

Soil development on the red beds of New Jersey

by

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INTRODUCTION

THE WILLIAM L. HUTCHESON MEMORIAL FOREST

The William L. Hutcheson Memorial Forest in Somerset County, New Jersey offers a special interest to the student of soil genesis. The forest, which has been preserved since the appearance of white man, serves as an excellent area for the study of soils under natural conditions.

Rutgers University scientists have for many years been using the forest as an outdoor laboratory. Information concerning the life of the forest had been collected in the fields of botany, zoology, and related disciplines. Because of the interdependence between the bio-factors and soil, it was felt that plant and animal distribution would be highly related to soil conditions. For this reason, detailed pedologic studies of the area were undertaken, and hopefully, this report will serve a useful purpose in understanding the ecosystem.

The soils of the red shale, because of their unique soil morphology, have always attracted the attention of pedologists. Field examination shows that the soils, despite their great chronologic age, do not show the full impact of soil-forming processes as do most other soils in the general area. Questions regarding the cause of this resistance to soil formation have led to extensive discussion; however, little factual information has been set forth. Consequently, it was felt that a study of these soils would provide information that would help in the understanding of the genesis and development of the red shale soils.

PURPOSE OF THE STUDY

The purpose of this study is threefold: 1) to characterize the major kinds of soils present in the Hutcheson Memorial Forest, 2) to provide a map of the distribution of these soils, and 3) to explain insofar as possible the pedological processes occurring or having occurred in the area. The possibility of disclosing the reasons for, and the distribution of, the unique soil morphology rests on establishing the nature of the geochemical phenomena that have taken place during the Triassic period and the subsequent processes that have acted on the bed rock to produce the present soils.

DESCRIPTION OF THE AREA

Locality. Hutcheson Memorial Forest is a woodland of 65 acres situated about one mile east of East Millstone (40° 30' N., 74° 34' W.) in Somerset County (figure 1). Peripheral to the woods are fields totaling some 71 acres which are part of the tract (Monk, 1958; Buell, 1957). The boundaries of the fields and forest are: north, Amwell Road (N. J. 514); northwest, New Brunswick spur of the Pennsylvania Railroad; south and east, farmland.

Climate. The dependent climate (Conrad, 1936) of New Jersey is governed by centers of action that are located over the interior of North America and the Atlantic Ocean in the winter, and over eastern Canada and the subtropical Atlantic Ocean in the summer; consequently, the northwest winds prevail during the cold season and the southwest winds in the summer. Since New Jersey is between the uniform seasonal distribution pattern of the New England states and the subtropical Atlantic type, characterized by late summer and fall precipitation, the rainfall in the state is rather well distributed (Biel, 1958).

Of particular interest in a soil study is the climate a few inches above the surface of the soil; this has been named microclimate. From August 1951 through August 1952, air temperatures were measured at 2 m., 20 cm., and 5 cm. heights above the ground of the forest while soil temperatures were measured at a depth of 4 cm. below the soil surface. Air temperature values between -3° and 96°F. were recorded in the woods. Soil temperatures in the woods (29° to 73°F.) showed less annual variation than in the open field (30° to 90°) (Sparkes and Buell, 1955.)

Vegetation. Bard (1952) considered Hutcheson Memorial Forest to approximate climax vegetation because of the large percentage of oaks and hickories and she stated that vegetation and soil conditions are presumed to be at the end point of a succession. Monk (1958), in a more recent study, casts some doubts about the theoretical existing climax. The number of small oaks is probably not sufficient to maintain the prevalence of the species (Monk, 1958). This condition is becoming more critical since more shade-growing species have ap-

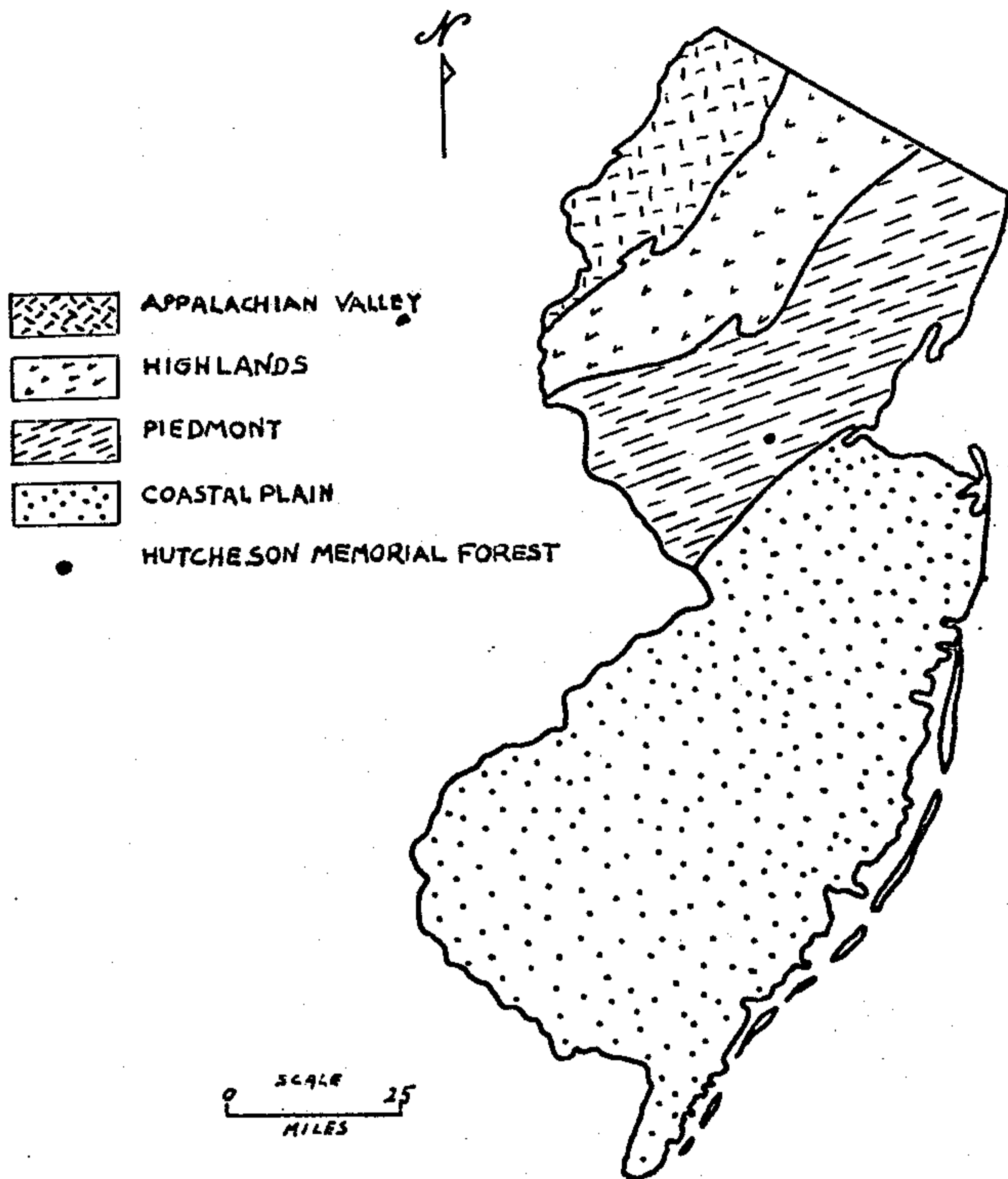


Figure 1. Geographic provinces of New Jersey and location of Hutcheson Memorial Forest. (Source: *Our New Jersey Land*, G.A. Quakenbush, 1955)

peared. In the recent history of the forest, two shade-tolerant species, the sugar maple (*Acer saccharum*) and the Norway maple (*Acer platanoides*), have made their appearance. Monk reports that both species seem to compete favorably with the predominant oaks.

The upland of the forest shows a distinct stratification of three layers. The main canopy is made up of white oak (*Quercus alba*), black oak (*Quercus velutina*), red oak (*Quercus rubra*) and red hickory (*Carya ovalis*). This canopy reaches a height of 95 feet. Always present in this group are white ash

(*Fraxinus americana*), red maple (*Acer rubrum*), sugar maple, Norway maple and beech (*Fagus grandifolia*). The understory is made up of dogwood with a maximum height of 35 feet. The shrub layer consists primarily of maple-leaved viburnum (*Viburnum acerifolium*), blackhaw (*Viburnum prunifolium*), choke-cherry (*Prunus virginiana*) and black cherry (*Prunus serotina*), all of which occur locally and prevail in the open areas. The herbs are characterized by the prevalence of May apple (*Podophyllum peltatum*). In areas with slow drainage, white oak, red oak, red maple, white ash and pin oak (*Quercus palustris*) are the most prevalent species (Monk, 1958).

Physiography. New Jersey is divided into two geographical provinces, the Appalachian and the Coastal Plain (Kümmel, 1940). The Appalachian province consists of four divisions named in order from west to east: Appalachian Plateau, Appalachian Valley, Highlands, and Piedmont Plain. All but the first are present in New Jersey. Hutcheson Memorial Forest is included in Piedmont Plain. The Piedmont Plain is situated in the easternmost part of the Appalachian province and forms a band across the state that attains a width of 32 miles along the Delaware River and 15 miles at the boundary with New York State. In general, the topography of the province is a gently rolling plain with altitudes between 11 and 250 feet.

Bedrock Geology. Triassic sedimentary rocks of eastern North America form a belt of discontinuous basins which follows the Atlantic coastline from Nova Scotia to South Carolina (Krynine, 1950). The Upper Triassic Brunswick formation, member of the Newark group, underlays the soils of the area (Kümmel, 1940). On the basis of lithologic features, the Newark group in New Jersey has been divided into three formations, here listed in the order from the oldest to the youngest: Stockton, Lockatong, and Brunswick. The thickness of the three formations was estimated by Kümmel (1896) as being between 11,800 and 14,600 feet. More recently McLaughlin and Willard (1949) concluded that "the Newark is not a group of three distinct formations successively deposited, but rather a series of interfingering, in part contemporaneously formed, continental facies."¹ If this is the correct interpretation, then the total thickness would be 11,100 feet at maximum rather than the maximum indicated by Kümmel.

The Brunswick formation is represented by soft red shale consisting of siltstone with a high percentage of clay and interbedded sandstone (Kümmel, 1940). Although the red color is the most

common, gray, gray brown, green, yellow and black colored members are found.

Hand specimens collected from exposures in the vicinity of Hutcheson Memorial Forest show glauberite crystal cavities which are partially filled with calcite, authigenically produced from the alteration of glauberite. Calcite also occurs as veinlets. Other authigenic minerals are barite, hematite, feldspar, and dolomite. It has been reported that the most predominant clay mineral of shale is a dioctahedral illite which presents two-thirds of the positions in the octahedral layer occupied by trivalent ions (Sturm, 1956). According to Kümmel (1896), the Triassic beds were formed by the deposition of material into shallow estuaries from adjacent land located on the northwest and southeast. Furthermore Kümmel states that the source rocks had to be highly altered because of the finegrain sediments and the absence of pebbles in the red beds. Ripplemarks, sun-cracks, and raindrop impressions and tracks of animals found in many beds were left on the beaches between tides. The Triassic basin was formed as a result of the uplift of folded Paleozoic formations on the west and the subsidence of strata on the east (McLaughlin and Willard, 1949). In this basin, under continental conditions the "long accumulated red regolith, formed from the sedimentary rocks of the peneplane to the west, was swept eastward into the new depression... simultaneously from the east feldspathic sands, pebbles and possible red mud streamed basinward from slightly uplifted Appalachia."² When the basin became wider, swampy conditions prevailed in its axial part where only the finer sediments were collected together with black carbonaceous debris. Under the reduction potential, the Fe ion passed from +3 to +2, thus losing its red color. The consolidation of this material produced the Lockatong formation which covered the Stockton formation. In turn, both formations were mantled by the Brunswick red shale which in the lower part was intermixed with beds of the Lockatong formation.

Sturm (1956) presents a similar hypothesis for the paleogeography of the Newark basin in New Jersey and provides data that are in agreement with the lacustrine origin of the Lockatong. Dorsey (1919) investigating the stratigraphy of the Newark group in Maryland reported that the Newark sediments had formed under terrestrial conditions. Richards (1944), studying fossil mollusks of the Triassic sedimentary rocks in Pennsylvania, concluded that the Newark series had originated in fresh water. Krynine (1950) reviewed the literature on the subject for the Connecticut basin and reported that the fossils found include fresh water shells, fishes, and

¹ McLaughlin, D. B. and Willard, B., 1949. Triassic facies in the Delaware Valley. Penn. Acad. Sci. Proc., V. 23, p. 35

² Ibid., p. 39-40

terrestrial vertebrates. Remnants of the fauna together with those of the flora suggest a warmer climate than at present, approximating a subtropical type of climate (Krynine, 1950).

Work done by Krynine (1950) on the Triassic sedimentary rocks of Connecticut brings out important facts about the origin of the red beds and the probable climate during the Newark epoch. This investigation is of particular interest to the problem of soils associated with the Triassic substratum.

In fact, the possibility of establishing a prevailing climate during the time the sediments were deposited would allow one to postulate, according to climatogenic theory, the type of soil present in the area during the time of deposition. If the soil of the general area was the potential source of the sediments of the Brunswick formation, then the characteristics of the rock would be related to the pre-existing soils. This assumption is valid only if diagenic processes have not modified the sediments during and after deposition. Krynine (1950) postulates that the climatic conditions which had produced the red-pigmented soils were possibly not different from those existing in the basin of sedimentation. A different point of view on the matter is presented by Sturm (1956) with respect to the Newark basin in New Jersey. He suggests climatic differences between the Highlands (the source of sediment) and the basin of deposition. These differences would be indicated by the different degree of crystallinity of illite, which is more sensitive to environmental changes than the red hematite pigment. According to Krynine (1950), the probable climate of the Newark epoch which occurred during the Upper Triassic time was rather uniform with fluctuations that did not alter the general prevailing type of climate. Krynine (1950) postulated a

"Savanna climate of that type — with a uniformly high temperature around 80°F. a totally rainless dry season lasting at least 3 months (and possibly longer), and a heavy rainfall exceeding 30 inches (and possibly much more in the valley proper) and in excess of 60 to 70 inches (or much more) in the erosion region of the scarp — seems to be entirely compatible with all the facts of the Triassic sedimentary record."³

Under these conditions, a red soil formed on granitic rocks comparable to the soil formed today in a warm humid climate may be considered under the inclusive name of lateritic soil. According to Sturm (1956), the above-mentioned climate assumed for the Connecticut basin needs to be modified in the case of the New Jersey Triassic basin, in that he does not accept Krynine's hypothesis; instead he indicates that there must have been a lower rainfall and a soil other than lateritic.

Surface Geology. The area included in Hutcheson Memorial Forest is situated south of the Wisconsin terminal moraine and therefore, was not covered by the last ice sheet. The topography of the area has not been altered by glacial invasions, but has been modified by rain and river erosion. Unconsolidated materials transported by rivers running southward from the moraine during the ice advances were deposited in the Millstone Valley and probably covered at least part of the land of the Hutcheson Forest. Erratics, either exposed at the surface or buried a few feet, are found in the woods and adjacent fields. Salisbury (1902) postulated that the gravels were deposited by the Raritan River which during the Pleistocene time flowed southwestward. Salisbury further supports the hypothesis that the Raritan River during the glacial period followed a southward pattern along the Millstone-Stony Brook-Shipetaukin-Assanpink route coming to a confluence with the Delaware River. The origin of the gravels found in the woods as related to the deposits left by the Raritan River cannot be further substantiated. The only thing that can be said is that at East Millstone gravel deposits are found as reported by Salisbury:

"At East Millstone and a little further north, there is a considerable body of gravel at fifty feet or above. The gravel is largely of red shale material, but with this material of local origin there is a small amount of glacial gravel. A more considerable deposit occurs at (1 mile away), while near Weston there is a considerable body of gravel which contains a larger percentage of glacial material than most of that in the valley to the south."⁴

Salisbury restricted the detection of these gravel bodies in the valley within the seventy foot contour; evidently, this was the only part that had significance as far as proving his hypothesis. The presence of a gravel mantle above the seventy foot contour has not been proven or disproven. It is of interest, however, to note that much of the Hutcheson Memorial Forest is included below the seventy foot contour which would encourage one to think that, topographically speaking, the forest was in the part of the valley which could have been the bed of the Pleistocene Raritan River or the adjacent flood plains.

Soils. Mention of the soils derived from the Brunswick formation was made in 1896 by Kümmel (1896). In this report, these soils have been described as shallow, clayey, red, and containing fragments of the shale. The shallowness of these soils was ascribed to the ease with which the fine products of rock alteration were removed even on gentle slopes. According to Kümmel (1896), erosion and

³ Krynine, P. D., 1950. Petrology stratigraphy and origin of the sedimentary rocks of Connecticut. Conn. State Geol. Surv., Bull. 73, p. 182

⁴ Salisbury, R. D., 1902. The glacial geology of New Jersey. N. J. Geol. Surv. Trenton, V. 5, p. 725.

not the slow disintegration of shale is the cause for thin soils. In 1923 appeared the first soil survey report of the Bernardsville area, New Jersey, which included the soils formed on red shale (Lee and Seltzer, 1923). These soils were classified under the name of Penn and they were further subdivided into textural classes. Penn silt loam was described as a dark reddish brown soil, which attains a depth at the surface of 8 to 10 inches. The subsoil retains the Indian-red color of the parent rock and the depth is estimated to vary from 20 to 36 inches.

Marbut (1928), in his discussion of the soils formed on the red beds, pointed out the resistance of shale to weathering. "In general, we usually think that shales, being soft rocks, decompose very readily and quickly form a thick layer of soil material, giving the climate and the vegetation an opportunity to develop a mature soil in a short time. The geologist thinks of shales as soft rocks. It is the universal experience of the student of soils, however, that soils developed on shales are shallow."⁵ Another character of immaturity of these soils, is according to Marbut, the persistence of red color inherited from shale. Normal soils of the area have a light brown or yellowish brown tinge, whereas soils formed on red shale are red in color. Such soils, therefore, cannot be considered mature until the red color disappears and the A horizon has developed a brown or grayish brown and the B, a deeper brown. Marbut compares the Penn soils to A-C soils of Stremme because they have only an A and a C and not a true B horizon. Penn soils were then included in the eastern part of the Brown Forest soils region (Marbut, 1928). Marbut (1935) later placed the Penn soils in the Gray Brown Podzolic, a new soil group earlier suggested by Baldwin (1927). By the mid-nineteen thirties the Russian pedological approach had made inroads on American concepts of soils and Joffe (1937) in keeping with the Russian concepts published pedologic data on the Penn soils. The total chemical analysis of the soils on profile basis offered by Joffe shows the degree of translocation of certain elements in the solum. In the case of Penn silt loam, there is virtually no translocation of constituents in the profile, and SiO_2 is uniformly distributed; therefore, this soil is considered an endodynamorphic or lithogenic soil, which implies that the parent material determines the major profile features. Up to 1937 the taxonomy of the Penn series was rather inclusive. Included with the Penn were soils formed on residuum, soils on glacial outwash or other surface deposits. The term Penn (Quakenbush, 1955) is now reserved for the true residuum conditions whereas names for other series were created for situations other than residuum.

⁵ Marbut, C. F., 1928. Soils, their genesis, classification and development. A course of lectures given in the Graduate School of the United States Department of Agriculture, p. 8-9 (unpublished).

The soils in Hutcheson Memorial Forest underlain by the Brunswick formation should not be called, therefore, Penn soils inasmuch as they are not entirely derived from red shale. Gravel, foreign to the area, is found at the surface or at some depth in the profile. The presence of this foreign material had contributed to the increase of the thickness of the solum and also permitted a deeper brown color to develop. Although information is not available about the precise thickness of such mantle of unconsolidated rocks, one may infer from the rather uniform texture of the soils which reflects a close association with the texture of the Brunswick red shale, that this mantle did not attain a great thickness prior to the formation of the soil.

Soils derived from the sandy substratum located in the western part of the forest were included with the Dunellen sandy loam (Lee and Seltzer, 1923). This series consists of soils derived from glacial outwash and local red shale material. A possibility that some of the sand is of wind-blown origin is suggested for areas along the Millstone River at East Millstone, Weston, and other localities (Lee and Seltzer, 1923).

Joffe (1949) was inclined to regard Penn soils as a climatogenically subdued soil type. With this expression, he meant that in the development of this soil other factors had played a more important role than climate and the biosphere. In the case of Penn, the parent material is the most important factor, and, therefore, this soil is also called lithogenic (Joffe, 1949).

"The reason for the lack of profile development in the Penn soils lies in the makeup of the parent material — Triassic shale. Apparently, the shale represents a consolidated material of eroded soil and weathered rock of a period when the climate was conducive to lateritic weathering, the R_2O_3 constituents, with very little of other primary minerals. When the shale was exposed, and the agents of weathering and the factors of soil formation began their work, there were very few primary minerals that could be translocated and distributed in the profile. In short, there have not been enough reaction products to affect a differentiation into distinct horizons."⁶

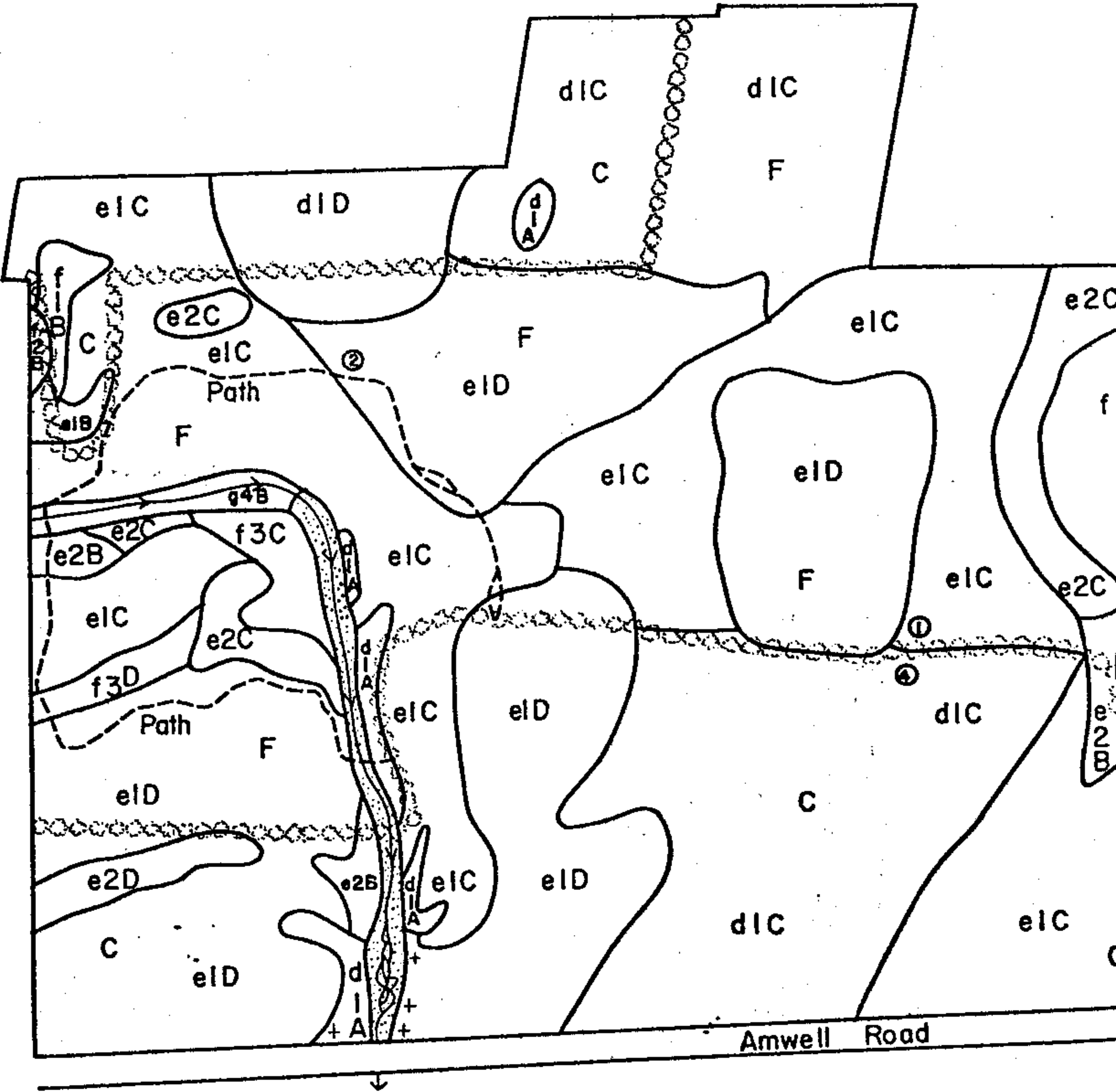
Joffe called Penn an endodynamorphic soil after Glinka's terminology.

Bard (1951) in an ecological study of Hutcheson Memorial Forest presented information on the soils of the fields and woods; but this research was not aimed to disclose the genesis of the soils. Losche⁷ provided data on clay mineralogy and chemical analysis of the soils. Connor *et al* (1957) determined trace element distribution in the shale soils of Hutcheson Memorial Forest. This investigation proves once more the dominant role of red shale in the

⁶ Joffe, J. S., 1949. Pedology. Pedology Publications, New Brunswick, New Jersey, p. 637.

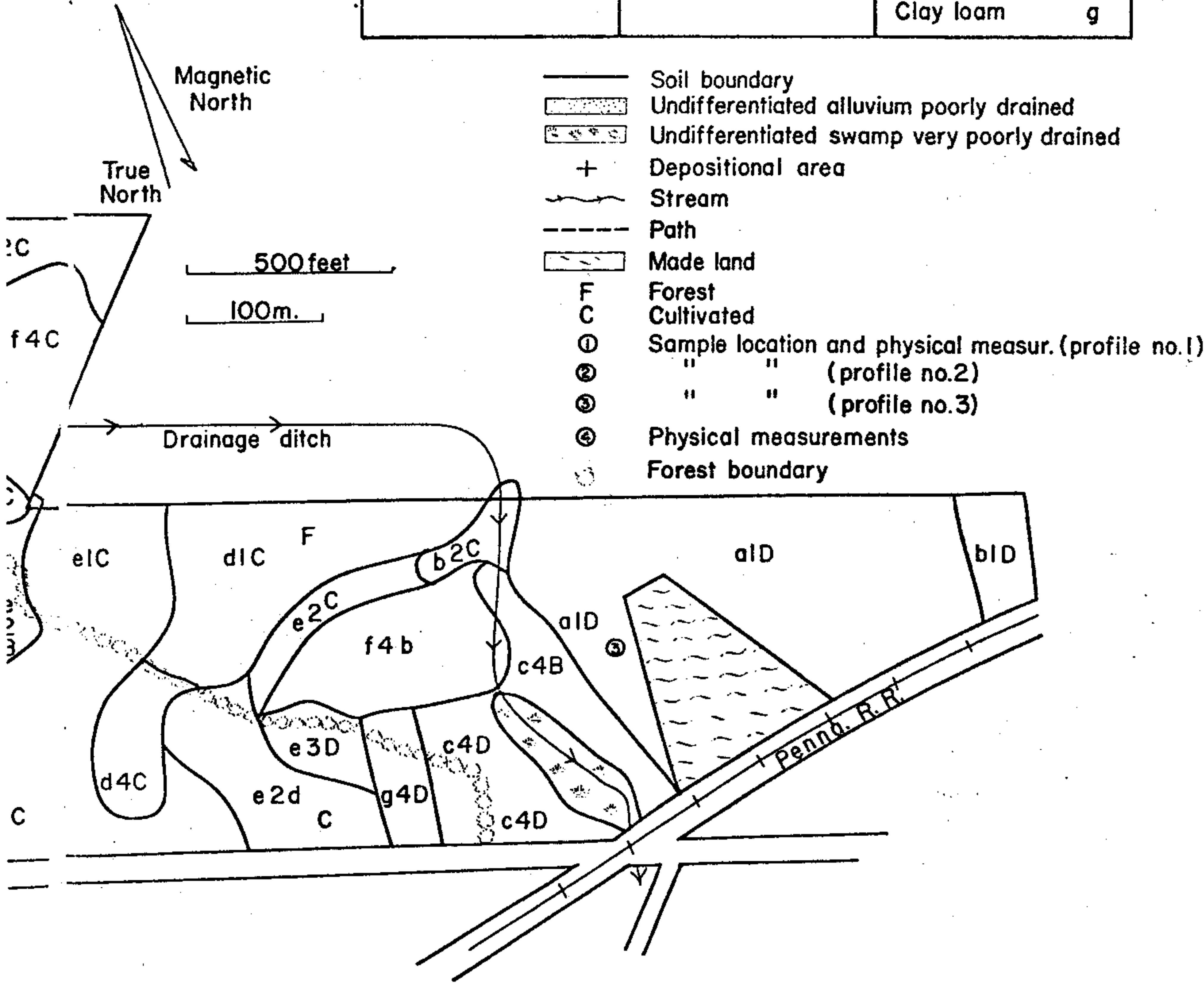
⁷ Losche, C., 1955. Senior Thesis, Rutgers University.

Fig. 2. Soil Map
of
William L. Hutcheson Memorial Forest



Symbols

Depth		Drainage		Texture	
0-12 inches	A	Well drained	1	Loamy sand	a
12-18 "	B	Mod. well drained	2	Sandy loam	b
18-24 "	C	Imperf. drained	3	Sandy clay	c
24 plus "	D	Poorly drained	4	Loam	d
				Silt loam	e
				Silty clay loam	f
				Clay loam	g



formation of this soil. Trace elements are uniformly distributed in the profile indicating a minor horizon differentiation.

Until the present study was initiated, no other work had been done to establish the true nature of the processes taking place in this soil. The student of soil genesis is, therefore, confronted with an enigmatic situation. While on one hand the term endodynamorphic has not been widely adopted to identify Penn soil, the inclusive name of Gray Brown Podzolic soils does not seem to be appropriate to typify the existing conditions.

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Figure 3. Soil profile pit #2. The spade rests on bedrock.

METHODS OF INVESTIGATION

SOIL DESCRIPTIONS

Three soil pits were excavated in areas which were representative of the locality. Once the pits were opened, the horizons were described and samples were taken for further analyses. The three sites are designated as profiles No. 1, No. 2 and No. 3 (figure 2). Most of the pedologic interpretations are based on information obtained from these three profiles. Profiles No. 1 and No. 2 were located on relatively deep well-drained soils of the upland and profile No. 3 from a well-drained fluvial deposit.

MORPHOLOGY OF PROFILE No. 1 *

Horizon	Depth in Inches	Description
A ₀	2-1	Leaves of deciduous trees.
A ₀	1-0	Dark gray to black, partially decomposed plant residues, very friable.
A ₁	0-2	Dark gray, discontinuous mineral matter, rich in organic matter.
A ₂	2-10	Brown, 7.5YR5/4, silt loam loose and friable, throughout; well aggregated in small round aggregates (weak crumbs). Individual peds are arranged in weak nuciform structure; worm holes and roots abundant.
A ₃	10-12	Transition zone; no abrupt change in the horizons.
B ₁	12-20	Reddish yellow, 5YR6/6, silt loam. Firm to compact consistency. Some zones appear to be aggregated; the rest firm.
B ₂	20-24	Yellowish-red, 5YR4/8, loam. Hard and compact, nuciform structure with peds less than 1 cm. on each edge; when crushed, yields a fine crumb aggregate.
C-D	24+	Red 2.5YR4/6, sandy loam, intermixed with local bedrock (Brunswick formation).

* Samples collected on July 15, 1957 under very dry conditions. (See figure 2 for location.)

MORPHOLOGY OF PROFILE No. 2 *

Horizon	Depth in Inches	Description
A ₀	2-1	Deciduous leaves, litter and twigs; mostly undecomposed.
A ₀	1-0	Organic debris either decomposed or partially so. High degree of aggregation.
A ₂	0-7	Light reddish brown, 5YR6/4, clay loam. The top of this horizon proves to be well aggregated but less than the preceding horizon. Consistency becomes firm with depth. Very few cleavage lines when the soil is broken. Roots, very abundant.
		Transition zone well defined.
B ₂	7-11	Reddish-yellow, 5YR6/6, clay loam. Firm consistency. Many small pores visible to the naked eye. Small aggregates, massive-like appearance. The soil breaks along no defined lines when it breaks into sub-angular-blocky peds.

- B₂C 14-20 Strong brown, 5YR5/6, clay loam. Fragments of unweathered red shale increase with depth. Erratics seem to be concentrated at this horizon. Hard consistency, gritty. Random cleavage, primarily on the adjoining area of the shale fragments. Harsh feeling when the aggregates are crushed.
- C 20-28 Dusky red, 10YR3/4, sandy clay. Compact, gritty layer of merely red shale at various stages of weathering. Clay skin appears at the surface of the fragments. Along the cleavage plane of the aggregates, spots of black coating are noticeable.
- D 30 Red 2.5YR4/6, shale (bedrock). Brunswick formation.

* Samples collected on August 7, 1957 under very dry conditions. (See figure 2 for location.)

MORPHOLOGY OF PROFILE No. 3 *

Horizon	Depth in Inches	Description
A ₀	1-0	Formed of deciduous litter partially decomposed, grading sharply into mineral soil.
A ₂	0-12	Dark brown 7.5YR4/4 loamy sand. Moist, uniform in color and texture. Single grain structure with some very weak crumbs.
B	12-29	Dark reddish brown 5YR3/4 sandy loam, very moist. Few and indistinguishable cleavage lines suggest a weak block-like structure. The horizon as a whole does not show cross bedding features.
C ₁	29-38	Yellowish red 5YR4/8 sandy clay loam massive structure, very wet but firm consistency.
C ₂	38-40	Transition zone more clay than above.
C ₃	40-48	Light reddish brown to reddish brown 5YR6/4 to 5/4 loam, very wet. Subangular type of structure with massive-like degree of aggregation. The consistency is firm. The horizon shows signs of impeded drainage in form of gray hues and partially mottled. Red shale material.
D	48-52	Dark yellowish brown 10YR4/4 shaly material that breaks along parallel plain, wet conditions. Considerable clay in the shale matrix. Materials of fluvial origin.

* Samples collected on January 17, 1958 under moist conditions. (See figure 2 for location.)

SOIL MAPPING

The soils in the forest and fields were classified according to texture, drainage, and depth characteristics. Texture was determined by the field method with certain samples further verified by laboratory techniques. Drainage classes were distinguished by the usual field methods; depth of soil was measured directly with the soil auger. Seven textural classes

were adopted:

Loamy sand
Sandy loam
Sandy clay
Loam
Silt loam
Silty clay loam
Clay loam

Four soil-drainage classes were used:

Well-drained
Moderately well-drained
Imperfectly-drained
Poorly-drained

Four depth classes were adopted:

0-12 inches to bedrock
12-18 inches to bedrock
18-24 inches to bedrock
> 24 inches to bedrock

By grouping texture, drainage, and depth, soil taxonomic units were formed, and indicated on the map.

SOIL SAMPLING

Field work was started during the summer of 1957 and continued intermittently until late 1958. Because of unusually dry conditions from July 30 to September 27, 1957, samples were taken during this time in the vicinity of profile No. 1 (forest) and adjacent field for soil moisture determinations. The samples were collected by the use of a Veilmeyer tube at 3-inch intervals in the forest to a depth of 36 inches and at 6-inch intervals in the field to a depth of 24 inches. Similar samples were obtained for pH measurements near profile No. 1. Samples for pH measurements were taken in triplicate on horizon basis including the A₀ and A₂. Sampling was started September 23, 1957 and continued to September 2, 1958.

A set of cores for permeability and bulk density (3-inches in diameter) were taken from the forest and field in November of 1958. The cores from the forest were collected near profile No. 1 in quadruplicate and on horizon basis (A₂, B₁ and B₂). Those from the adjacent cultivated field were taken according to depth, from 1 to 3 inches and from 6 to 9 inches.

MECHANICAL ANALYSIS

Air-dried soil samples of each profile were passed through a 2 mm. sieve; pebbles which remained on the screen were reserved for microscopic studies. Mechanical analysis was performed on the <2 mm. fraction. Samples were treated with 3 percent and again with 15 percent H₂O₂ and heated on a steam bath to remove organic matter. Organic matter-free soil was dispersed with 0.1N sodium hexametaphosphate and stirred with an electric mixer for 10 minutes. Clay and silt fractions were obtained by

the modified pipette method (Kilmer and Alexander, 1919). Sand was recovered by passing the soil suspension through a 300 mesh screen. The sand was then passed through a sieve-nest and separated into five fractions: very coarse, coarse, medium, fine, and very fine sands. The entire mechanical analysis was run in duplicate.

HEAVY AND LIGHT MINERAL ANALYSIS

The light and heavy minerals of the fine and very fine sands were separated by using bromoform with specific gravity of 2.85 as the separating liquid. Slides for microscopic study were prepared from both the heavy and light minerals. Aroclor (N=1.66) was used as a mounting medium for the heavy, and Lakeside #70 for the light minerals. Both separates were quantitatively analyzed petrographically by the field-count method. A variable number of fields (the number of fields is a function of the density of the grains in the slide) were chosen for each slide in such a way as to be equidistant to each other and to form a regular geometric pattern. A medium power objective (10X) was used for the count and the counted grains ranged from 250 to 300 for each slide. Such enumeration should maintain the error below 5 percent (Milner, 1950).

SEPARATION OF SOIL CLAYS

Soil clays (<2 micron) from profiles #1, #2, and #3 were recovered by sedimentation. The soil was dispersed with 0.1 N sodium hexametaphosphate; time and depth for the removal of the <2 μ particles were determined according to Stokes' Law. The suspensions were flocculated with dilute HCl; an aliquot of the clay was stored in a moist condition whereas another portion was dried at 100°C. A separate dispersion with NH₄OH was used to collect the soil clays from profile #1. All clays were left in a moist condition and in a dispersed state in order to be further subdivided. The separation of the clay into different fractions was accomplished with an International centrifuge (Tanner and Jackson, 1947). The following sizes were obtained: 2-1 μ , 1-0.5 μ , 0.5 μ . Prior to identification with X-rays, the clays were treated according to Jeffries' method to remove coatings (Jeffries, 1947). Such treatment, however, was not carried out for the 2-1 μ , 1-0.5 μ , 0.5 μ fractions.

X-RAY DIFFRACTION ANALYSIS

For X-ray analyses, the Debye-Scherrer powder method was used. Diffraction patterns were obtained with a Debye-Scherrer camera with a diameter of 114.6 mm. (Parrish and Cisney, 1948). X-rays were generated by a sealed-off tube containing a copper target. In order to produce the strong K characteristic component of general radiation, the tube was operated above the K excitation potential (35KV and 16 ma). The K α radiations were separated from

the weaker K β by a Nickel filter. X-ray equipment and cameras were manufactured by the North America Phillips Company, Inc. Clay samples were passed through a 325 mesh sieve and placed in a thin-walled lithium glass capillary tube (I.D., 0.2 mm.). Exposure time was fixed at 5 hours. The indexing of patterns was accomplished by obtaining directly the d spacing in angstrom units by superimposing the photographs on the scales for the cylindrical camera. Once the d spacing was obtained, the indexing was taken from the ASTM diffraction data cards and related literature. The intensity of the lines was estimated visually.

Thermal treatments were used in those cases in which X-ray patterns did not provide enough information for a positive identification of the minerals. The clays of A, B, and C horizons of all three profiles were again X-rayed after being heated respectively at 600°C. and 700°C. for two hours.

DIFFERENTIAL THERMAL ANALYSIS

Clay samples of each horizon of profiles #1, #2 and #3 were analyzed by the differential thermal method.⁸ A sample of untreated clay 400 mg. in weight was used for each determination. A high-temperature pressure-vacuum furnace as described by Lodding and Hammell (1959) was used for this purpose. Calcined alumina was employed as inert material in one of the sample holders. Each determination was run in dynamic air with 10 ounces of pressure. In order to avoid any oxidation of soil organic matter, samples from the surface were run in a nitrogen atmosphere. The heating rate employed was 14°C. per minute from room temperature to 1025°C. The thermograms were interpreted by comparing with standard reference clay minerals.

CHEMICAL ANALYSIS

The soils (<2 mm.) from the A₂, B₁, C-D horizons and clay from the A₂ of profile #1 were fused with Na₂CO₃ for total chemical analysis (Mellor and Thompson, 1938). Silica and sesquioxides were determined gravimetrically, iron volumetrically by the Jones reductor method, and titanium colorimetrically with a Bausch and Lomb Spectronic 20 colorimeter. A separate aliquot of soil was necessary for phosphorous which was determined colorimetrically. Alumina was obtained from the sesquioxides minus the oxides of iron, titanium and phosphorous. In the filtrate recovered from the separation of the sesquioxides, Ca and Mg were determined. Calcium and magnesium combined were determined by titration with versanate using Eriochrome Black T as indicator (Cheng and Bray, 1951). Calcium was precipitated with ammonium oxalate and later titrated with potassium permanganate (Prince, 1955).

⁸ The analyses were made at the Bureau of Mineral Research, Rutgers University, under the supervision of W. Lodding.

Table 1. Mechanical Analysis of Profile #1.

Horizon	Depth*	Very Coarse Sand	Coarse Sand	Medium Sand	Fine Sand	Very Fine Sand	Silt	Clay
		2-1 mm.	1-0.5 mm.	0.5-.25 mm.	.25-.10 mm.	.10-.05 mm.	.05-.002 mm.	<0.002 mm.
	ins.	pct.	pct.	pct.	pct.	pct.	pct.	pct.
A ₂	2-6	1.07	3.01	8.23	8.81	6.55	58.68	12.80
B ₁	12-20	0.86	3.08	7.93	6.91	8.80	55.75	16.66
B ₂	20-24	5.89	7.72	7.79	8.44	5.72	44.96	19.46
C-D	25-30	10.01	12.65	12.55	4.60	16.76	36.89	6.53

* Depth of sampling.

Magnesium was then estimated by differences. Another sample was sintered with CaCO₃ and NH₄Cl for K and Na determinations; both elements were estimated by flame photometer with a Perkin-Elmer flame photometer model 146.

The clays from A, B₁ and C-D horizons of profile #1 were fused for a partial chemical analysis. Silica, total sesquioxides, and iron were determined with the already mentioned techniques. Alumina was estimated by difference. Loss on ignition was determined by heating for 2 hours at 600°C.

Solutions recovered from cleaning the clay were wet-washed with aqua regia and treated with HNO₃ and HClO₄ in order to destroy the oxalates. Iron oxides, expressed as Fe₂O₃, were determined by the Jones reductor method.

SOIL ORGANIC MATTER

Soil organic matter of each horizon of profiles #1, #2 and #3 was determined by the chromic acid method. Total nitrogen in soil was estimated by the A.O.A.C. method and C:N ratios were calculated.

CATION EXCHANGE CAPACITY

Cation Exchange capacity of profile #1 was obtained by the ammonium acetate method (Schollenberger and Simon, 1945). The samples were leached with ammonium acetate solution with the pH adjusted to 7. After removal of the excess ammonia, the adsorbed ammonium was converted into gaseous ammonia and distilled into boric acid and titrated. The exchangeable cations present in the leachate were assayed with a Perkin-Elmer flame photometer model 52A using lithium as internal standard.

Table 2. Mechanical Analysis of Profile #2.

Horizon	Depth*	Very Coarse Sand	Coarse Sand	Medium Sand	Fine Sand	Very Fine Sand	Silt	Clay
		2-1 mm.	1-0.5 mm.	0.5-.25 mm.	.25-.10 mm.	.10-.05 mm.	.05-.002 mm.	<0.002 mm.
	ins.	pct.	pct.	pct.	pct.	pct.	pct.	pct.
A ₂	2-5	0.74	2.15	6.32	6.93	5.06	48.19	30.58
B ₂	10-14	0.63	2.27	6.62	7.45	5.31	49.91	27.79
B ₂ -C	16-20	3.50	4.01	5.28	6.63	5.17	37.64	37.75
C	22-28	10.31	14.20	10.07	8.54	5.13	13.50	38.20

* Depth of sampling.

pH

One part of soil and two of distilled water was used for preparing the suspensions for pH measurements. The suspensions were left overnight after repeated stirring and pH values were then determined with a pH meter (Beckman model H2) equipped with a glass electrode.

PERMEABILITY

Permeability and bulk density for forest soils and field soils were measured. Cores for the forest soils were taken from profile #1 on horizon basis (A₂, B₁ and B₂). Hydraulic conductivity under constant head was determined for 3 consecutive thirty minute runs.

MOISTURE

Moisture tests were made gravimetrically and reported on oven dry basis.

RESULTS

PARTICLE SIZE DISTRIBUTION

Table 1 and figure 4 show the results of the mechanical analysis for profile #1. Clay fraction attains its maximum in the B₂ horizon and its minimum in the C-D horizons. Silt tends to decrease with depth. The distribution of very fine sand behaves erratically at various depths, but it is found most abundant in the C-D horizon. A break in the percentages of sand fractions of A₂ and B₁ horizons is found at the level of the medium sand; the percentage of coarse and very coarse fractions increases quite drastically with depth. A more uniform distribution is found in the B₂ horizon and in the C-D

Table 3. Mechanical Analysis of Profile #3.

Horizon	Depth*	Very Coarse Sand 2-1 mm.	Coarse Sand 1-0.5 mm.	Medium Sand 0.5-.25 mm.	Fine Sand .25-.10 mm.	Very Fine Sand .10-.05 mm.	Silt .05-.002 mm.	Clay < 0.002 mm.
	ins.	pct.	pct.	pct.	pct.	pct.	pct.	pct.
A ₂	2-10	0.15	5.19	40.41	34.17	6.76	1.81	11.81
B	15-25	0.28	5.09	28.83	28.56	8.54	12.33	16.21
C ₁	34-40	0.21	2.83	16.77	22.33	11.65	23.29	23.21
C ₂	42-46	0.39	1.25	9.71	21.79	15.36	32.34	19.34

* Depth of sampling.

horizon. All five sand fractions, with the exception of the fine, reach highest values in the C-D horizon.

Profile #2 (table 2 and figure 4) manifests a rather different pattern with respect to profile #1 even though both share many common features. The clay of profile #2 shows an accumulation in the B₂-C and in the C horizons; there is not, therefore, a single horizon of accumulation. Silt which is rather uniform in the upper three horizons drops abruptly in C. Fine and very fine sand are uniformly distributed through the profile. Very coarse and coarse sands vary more than finer fractions, especially in the lower horizons. The clay content of this profile is almost twice that of profile #1. Higher values for the sand fractions are found in the C horizon.

Soil of profile #3, table 3 and figure 4, being sandy in nature, represents a remarkable variation in mechanical composition compared with the two previous profiles. Sand is more prevalent than silt and clay. Clay accumulates in the C₁ horizon, whereas the silt increases steadily with depth; similar behavior is shown by the very fine sand. Coarse, medium and fine sands decrease with increasing depth.

MINERALOGY OF VERY FINE SANDS

Heavy minerals of profile #1, table 4, figure 5, are not percentagewise more abundant than the heavies of other profiles with the exception of the C-D horizon. Black opaques, other opaques and zircon represent the bulk of the minerals in the A and B₁ horizons. A large number of heavy minerals

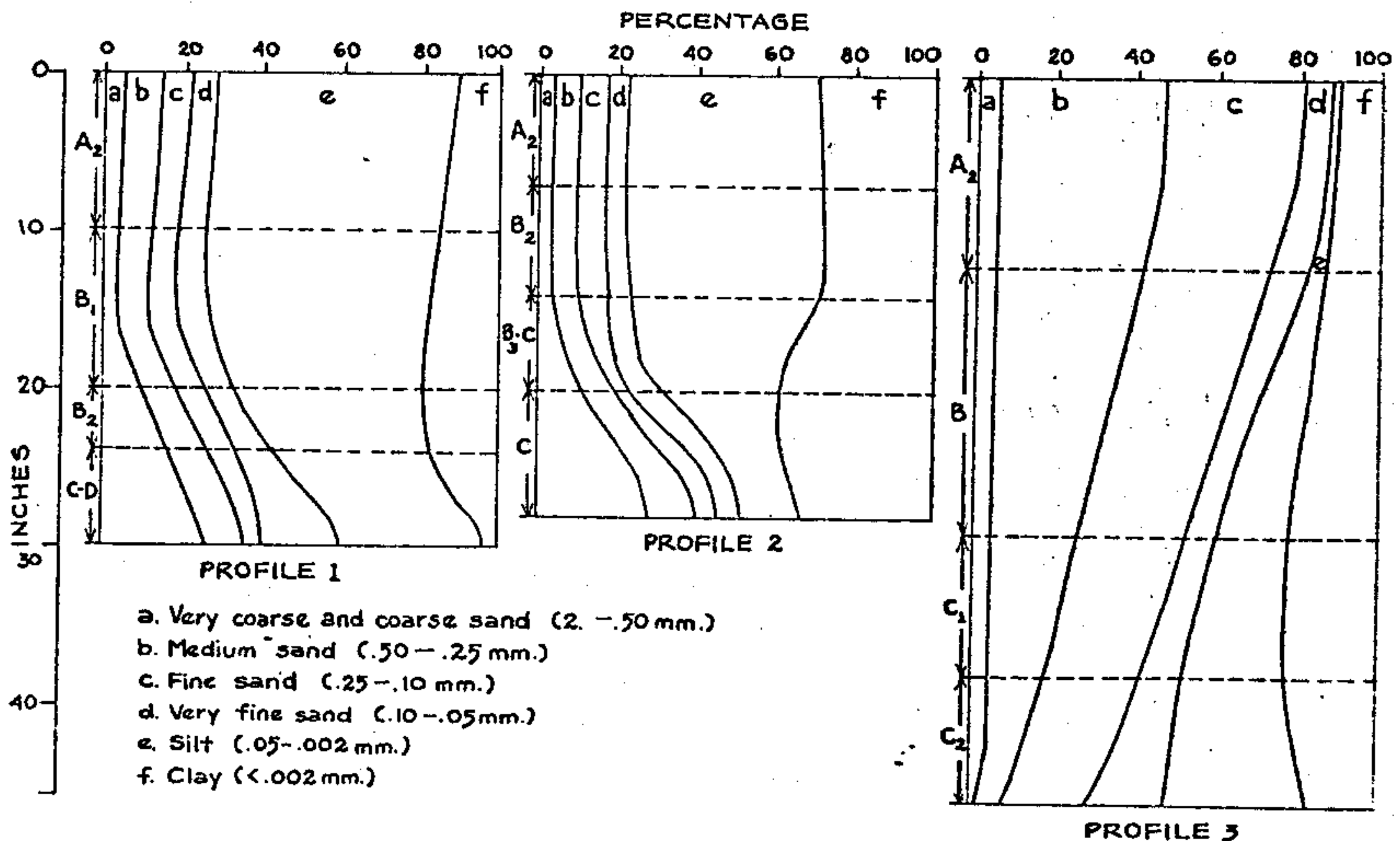


Figure 4. Particle size distribution in profiles 1, 2, and 3.

Table 4. Heavy Minerals of Very Fine Sand (0.10-0.05 mm.). Profile #1

Horizon	A ₂	B ₁	B ₂	C-D
Depth of Samplings	2-6 ins.	12-20 ins.	20-24 ins.	25-30 ins.
Heavy Minerals Percentage (by weight)	1.30	1.48	1.14	3.07
Minerals (by count)	pct.	pct.	pct.	pct.
Opaque (black)*	32	36	31	48
Opaque (others)**	41	43	41	35
Zircon	18	10	8	4
Tourmaline	4	1	4	2
Hornblende	1	1	3	5
Muscovite	—	t	9	1
Epidote	1	2	t	t
Staurolite	1	2	t	t
Chloritoid	1	1	1	t
Hypersthene	1	1	1	t
Garnet	1	t	t	1
Kaynite	—	t	—	t
Andalusite	—	t	—	—
Sillimanite	t	1	t	1
Rutile	—	—	1	—
Actinolite	—	1	—	1

* — including magnetite and ilmenite

** — including leucoxene, shale particles and iron coated minerals

t — trace — less than 0.5%

is present in the B₂ and C-D horizons, with a relatively high concentration of muscovite in the B₂ and hornblende in the C. Epidote, staurolite, hypersthene and andalusite are abundant in the B₁ horizon. Garnet prevails in C-D, while zircon decreases with depth, together with tourmaline and

Table 5. Light Minerals of Very Fine Sand (0.10-0.05 mm.). Profile #1

Horizon	A ₂	B ₁	B ₂	C-D
Depth of Sampling	2-6 ins.	12-20 ins.	20-24 ins.	25-30 ins.
Minerals (by count)	pct.	pct.	pct.	pct.
Quartz	81	83	58	50
Aggregates*	13	6	39	49
Opaque (black)**	—	3	—	—
Feldspars	6	6	4	t
Plagioclase	—	—	—	—

t — trace — less than 0.5%

* — mostly shale fragments

** — magnetite and ilmenite

epidote. The percentage of the other opaques is rather constant throughout the profile with the exception of C-D horizon where they decrease. The remaining heavy minerals are erratically distributed with depth.

In the light mineral fraction of this profile, table 5, quartz and mineral aggregates are most predominant. Quartz decreases and mineral aggregates show an increase with depth. Feldspars are fewer in the lower than in the upper horizons.

In profile #2, table 6, figure 5, the percentage of heavies in each horizon is rather uniform except in the C horizon where it decreases. In comparison with profile #1, mineral species are noticeably reduced in number. Black opaques and zircon still constitute the most common minerals, and they are fairly well distributed in the profile.

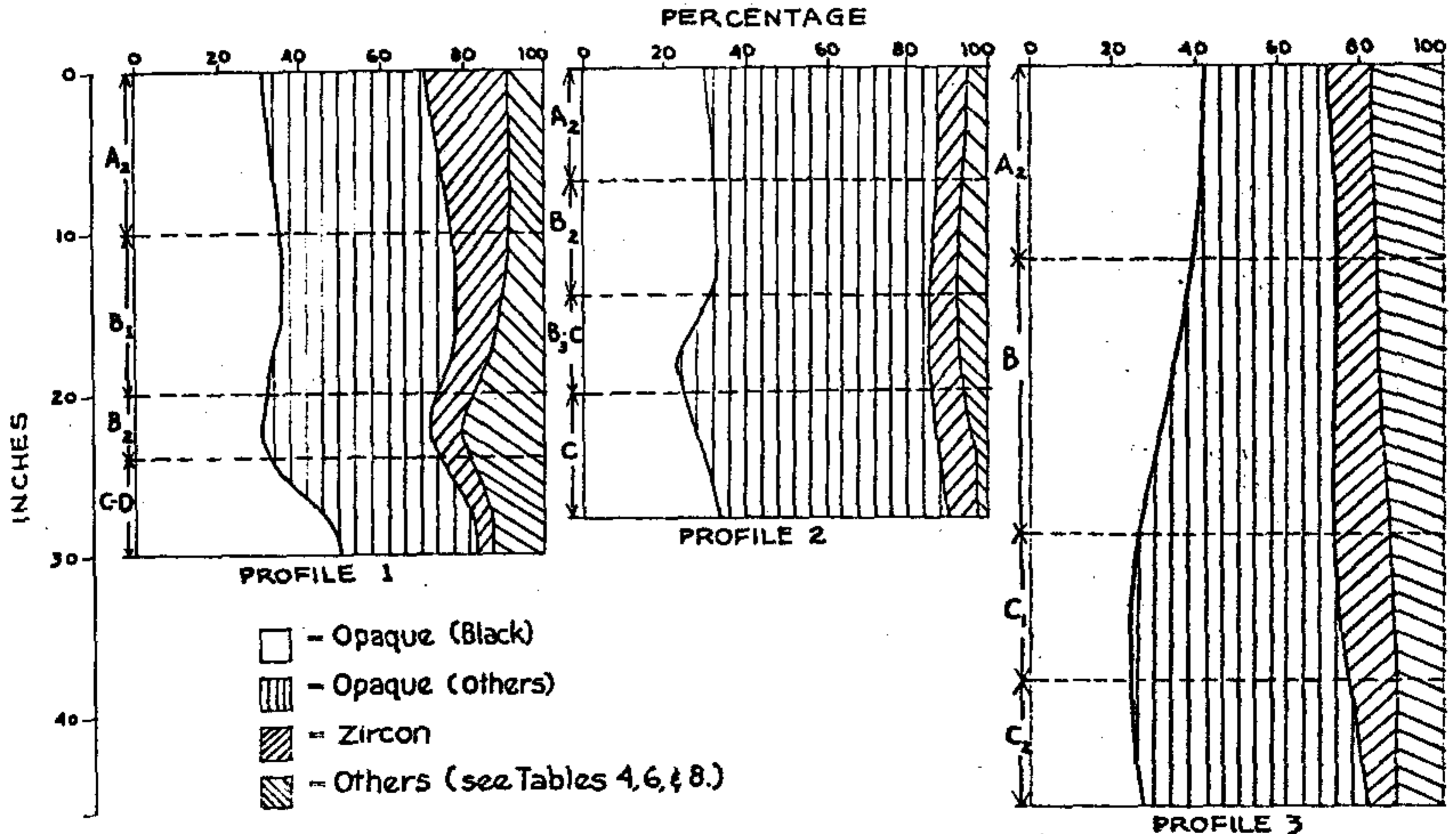


Figure 5. Heavy mineral suites of very fine sand from profiles 1, 2, and 3.

Table 6. Heavy Minerals of Very Fine Sand (0.10-0.05 mm.). Profile #2

Horizon	A ₂	B ₂	B ₂ -C	C
Depth of Sampling	2-5 ins.	10-14 ins.	16-20 ins.	22-28 ins.
Heavy mineral percentage (by weight)	2.0	1.50	2.33	0.73
Minerals (by count)	pct.	pct.	pct.	pct.
Opaque (black)*	31	33	23	32
Opaque (others)**	56	53	63	57
Zircon	7	6	7	8
Tourmaline	2	2	2	1
Hornblende	1	2	—	—
Hypersthene	1	†	—	—
Garnet	†	†	2	—
Epidote	1	2	1	†
Staurolite	1	1	†	†
Muscovite	—	†	—	—
Andalusite	1	—	—	—
Sillimanite	—	†	—	†

* — including magnetite and ilmenite

** — including leucoxene, shale particles and iron coated minerals

† — trace — less than 0.5%

Light minerals of profile #2, table 7 consist of quartz and mineral aggregates; with quartz decreasing and mineral aggregates increasing with depth. Very few feldspars are present in the C horizon.

Heavy mineral suites of profile #3 tend to be, percentagewise, uniform, table 8 and figure 5. Black opaques, other opaques and zircon are always the more frequent minerals, with some tourmaline, hornblende, epidote, chlorite, staurolite, sillimanite, garnet, chloritoid, mica, hypersthene, kyanite and rutile being present. Black opaques are more numerous at the surface than at depth; whereas other opaques decrease with depth. The number of zircon grains vary between two maxima; one in the A and the other in the B horizon, whereas tourmaline predominates in the lower horizons. Epidote and sillimanite are more abundant in the A and B than in the C₁ and C₂ horizons while staurolite is uniformly distributed together with chlorite, garnet and muscovite. Kyanite and hypersthene are absent in the C₂ horizon, but present in all other horizons.

Light minerals of profile #3 are represented almost entirely by quartz, table 9, mineral aggregates do not exceed 6 percent. Feldspars are somewhat constant numerically in the profile.

Table 7. Light Minerals of Very Fine Sand (0.10-0.05 mm.). Profile #2

Horizon	A ₂	B	B ₂ -C	C-D
Depth of Sampling	2-5 ins.	10-14 ins.	16-20 ins.	22-28 ins.
Minerals (by count)	pct.	pct.	pct.	pct.
Quartz	84	86	57	12
Aggregates*	9	3	37	85
Opaque (black)**	3	3	—	†
Feldspars	5	8	6	2
Plagioclase	†	†	†	—

* — mostly shale fragments

** — magnetite and ilmenite

† — trace — less than 0.5%

Table 8. Heavy Minerals of Very Fine Sand (0.10-0.05 mm.). Profile #3

Horizon	A ₂	B	C ₁	C ₂
Depth of Sampling	2-10 ins.	15-25 ins.	34-40 ins.	42-46 ins.
Heavy minerals percentage (by weight)	1.91	1.50	1.50	1.35
Minerals (by count)	pct.	pct.	pct.	pct.
Opaque (black)*	41	34	24	26
Opaque (others)**	31	40	53	55
Zircon	11	10	12	8
Tourmaline	2	3	5	4
Hornblende	3	4	1	2
Epidote	4	3	1	1
Staurolite	1	2	5	1
Sillimanite	2	1	†	†
Hypersthene	1	1	1	—
Chloritoid	2	1	—	—
Chlorite	1	1	2	2
Garnet	1	†	1	†
Muscovite	†	1	†	1
Rutile	—	†	—	—
Kyanite	†	†	†	—

* — including magnetite and ilmenite

** — including aggregates, leucoxene, shale particles and iron coated minerals

† — trace — less than 0.5%

Feldspars of all three profiles have been identified by the X-ray diffraction method; the 4.05 Å and 3.18 Å lines correspond to a K — feldspar.

CLAY MINERALOGY (X-RAY DIFFRACTION)

During the investigation of the clay components of the three profiles, the following clay minerals were identified by X-ray techniques.

Mica — This term is used to designate clay minerals that approximate an illite type. The name illite is reserved for those minerals whose basal reflections occurs at $10 + 0.2kX$; and upon heating to 500°C. for some hours the 10 kX lines become sharper with no major changes in the clay (Brown, 1951). Samples containing illite-mica were heated at 600°C. for 2 hours. This treatment results in most cases in a sharpening of the basal line without changing the rest of the structure. This illite-mica has been identified as a dioctahedral type because the (060) plane has an index of 1.50 Å (Brown, 1951). The weak 5.0 Å line indicates a ferriferous illite-mica in which the Fe ion prevails over the Al

Table 9. Light Minerals of Very Fine Sand (0.10-0.05 mm.). Profile #3

Horizon	A ₂	B	C ₁	C ₂
Depth of Sampling	2-10 ins.	15-25 ins.	34-40 ins.	42-45 ins.
Minerals (by count)	pct.	pct.	pct.	pct.
Quartz	92	93	91	90
Aggregates*	4	6	5	6
Opaque (black)**	—	—	—	—
Feldspars	4	1	4	4
Plagioclase	—	—	†	†

* — mostly shale fragments

** — magnetite and ilmenite

† — trace — less than 0.5%

ion in the central octahedral (Brown, 1951). A degraded illite-mica which is essentially the same as an illite-mica, except for certain ions (Fe) that have been removed by weathering, also belongs to this group. Removal of these ions is disclosed by the absence of the 1.50 Å and 2.37 Å lines which correspond respectively to the (060) and (133) planes (Sturm, 1956).

Vermiculite — The diagnostic spacing for vermiculite is the 14 Å line (002) which by thermal treatment at 700°C. for two hours is replaced by the 9.3 Å line which represents the basic talc-like layers (Walker, 1951). Vermiculite is thought to occur intermixed with hydrous mica.

Chlorite — Although chlorite gives a similar spacing as vermiculite, the two can be identified by heat treatments. A distinction between vermiculite and chlorite may be made also from the intensity of the lines. Vermiculite shows the 14 Å line as the stronger reflection and the 7 Å line as the weaker, whereas chlorite has both 14 Å and 7 Å reflections with the same intensity (Brindley and Robinson, 1951). This criterion could not be used in differentiating the two because the 7 Å line was weak in both vermiculite and chlorite. The samples were therefore, heated first at 600°C. and then at 800°C. for two hours. The 600°C. treatment did not make the 14 Å line disappear; the second treatment reduced the intensity to almost nil, but a trace of line still remains. On this basis the mineral has been called chlorite (Walker, 1951).

Kaolinite — The name kaolinite is intended to refer to that clay mineral that gives a distinct basal reflection in the 7.14 kX (001) regions (Grim, 1953). Well-crystallized kaolinite can be differentiated from a poor one on the basis of the broadening and weakening of certain reflections. In a poorly-crystallized kaolinite, the 7.14 Å line extends up to 7.20 Å and the (III) and (III) planes are not individually resolved (Grim, 1953). Since the second order of reflection of vermiculite and chlorite corresponds to the first order of reflection of kaolinite, positive identification of kaolinite cannot be achieved in the presence of a 14 Å mineral. A thermal treatment may be used to identify the two. A temperature of 600°C. for two hours or less is, in fact, sufficient to decompose kaolinite whereas chlorite is not appreciably affected by the treatment (Brindley and Robinson, 1951).

A mica approximating an illite is the most predominant clay mineral in profile #1, figure 6. This mineral shows a 9.50—10.50 Å line with variable intensity according to the abundance and the degree of alteration. The basal reflection tends to increase in intensity with depth reaching a maximum in the

C-D horizon and in the underlying bedrock; the degree of crystallinity also increases with depth and is shown by the augmented intensities of the 1.50 Å and 2.37 Å lines. The general trend of the illite-mica distribution is, therefore, an increase with depth. Vermiculite, on the other hand, prevails at the surface and decreases in the lower horizons. An X-ray diffraction pattern (<0.002 mm.) of the C-D horizon does not reveal traces of vermiculite, but different results are found when patterns are obtained from smaller fractions of the <0.002 mm. clay. These pictures show that vermiculite tends to concentrate in the fine clay particles (1-0.5 μ) and in the C-D horizon, it is found in less than 0.5 μ. Illite-mica does not seem to conform to this rule. Evidently, this mineral is segregated in the coarser clay fraction.

Kaolinite is present in the upper horizons and decreases in the lower; however, it does not attain great prominence even at the surface. Clay-size quartz remains constant in the profile except in the C-D where it decreases somewhat.

Clay minerals of profile #2 do not differ qualitatively from those of profile #1. A mica approximating an illite still remains the predominant clay; it appears to be more abundant in the B₂, B₂-C and C horizons and it is drastically reduced in the A₂ horizon. The degree of alteration of illite can be inferred from the absence or weakening of the 1.50 Å and 2.37 Å lines; in this profile the intensity of both reflections is noticeably reduced at the surface, indicating hence a higher status of weathering in the upper horizons. The upper horizon shows an accumulation of vermiculite which, contrary to profile #1, tends to persist throughout, even though it is less prominent in the lower part. Kaolinite is almost constant in the profile with a tendency to be widespread at the surface, however its quantity always remains small. Quartz persists in the profile with a reduction in the B and C horizons.

Clay minerals in profile #3 represent a variation with respect to the other two. A 14 Å mineral identified as chlorite is present in the A₂ horizon where it becomes the most abundant clay component. It is somewhat reduced in B, but it is still prominent in C, where, together with illite-mica, it makes the bulk of the clay. A reduction of chlorite is observed in the C₂ where the 10 Å mineral prevails. The latter shows an increase from the surface to the bed rock. The degree of alteration of this mineral is probably more pronounced than in the other profiles although this can be stated only on a qualitative basis. Kaolinite does not show a preferred distribution in any horizon and it is never very abundant.

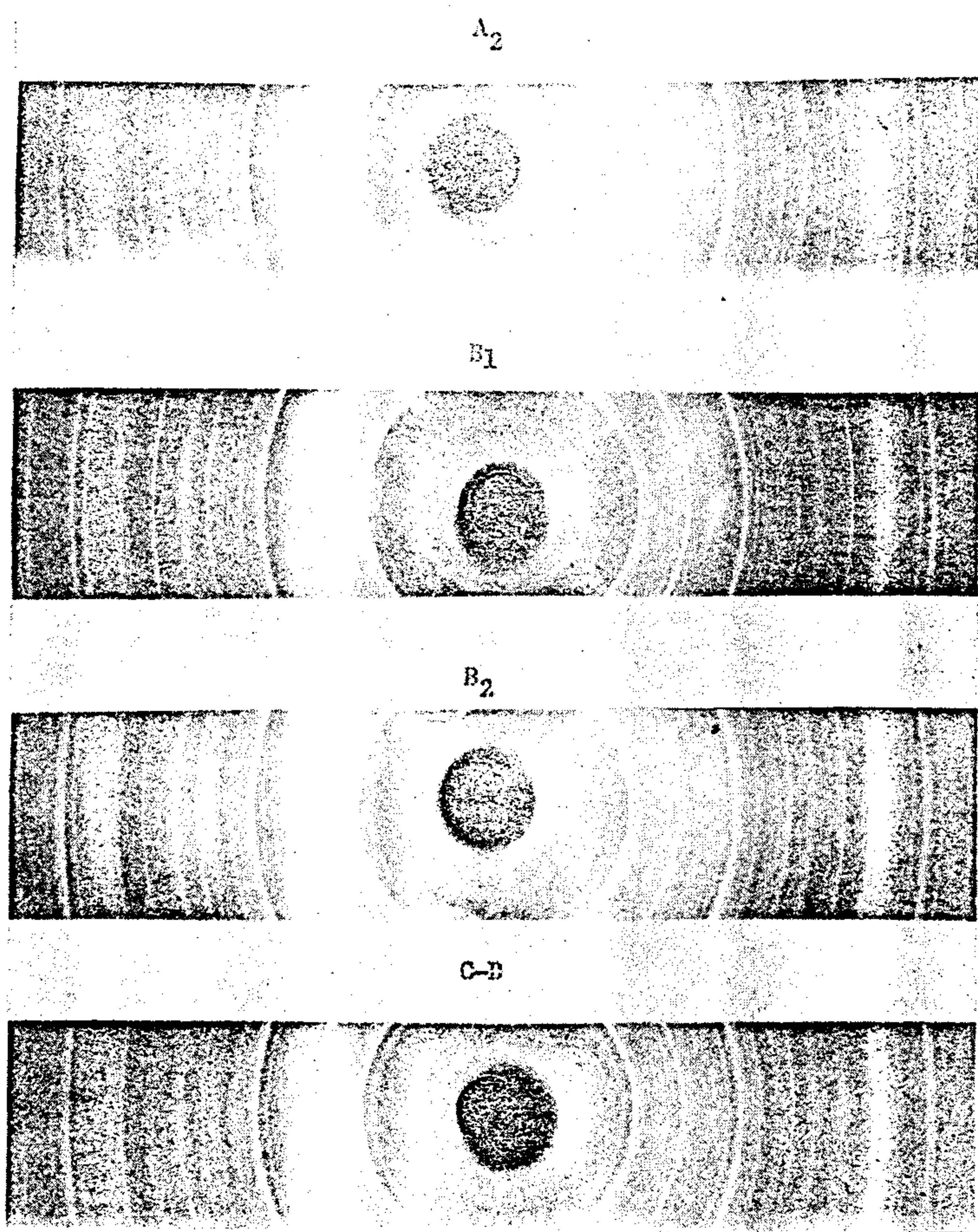


Figure 6. X-ray diffraction patterns from profile #1. Contact prints from powder camera. Copper radiation, $\lambda=1.54$; camera radius 57.3 mm.

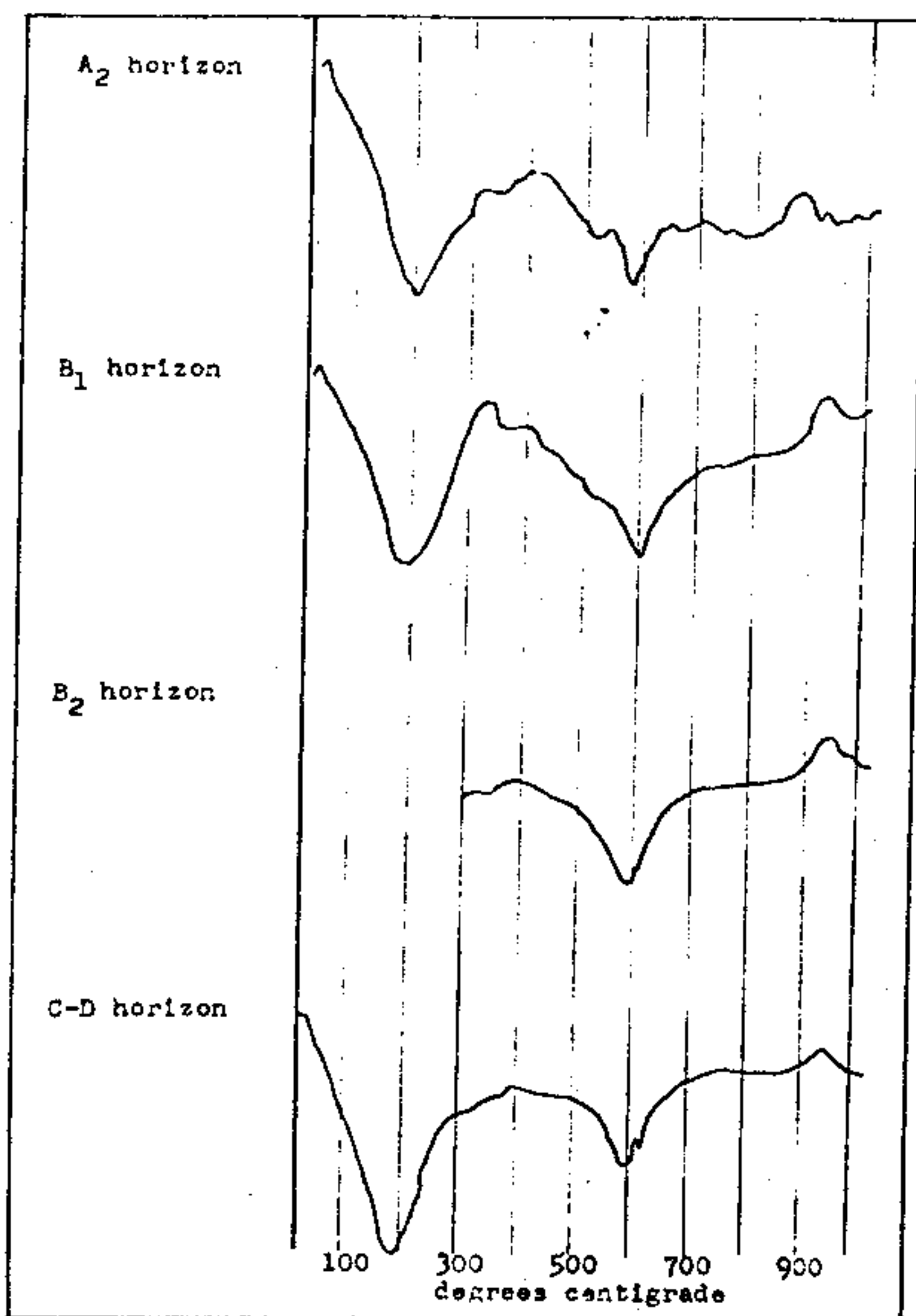


Figure 7. Differential thermograms of <0.002 mm. clay from profile #1.

DIFFERENTIAL THERMAL ANALYSIS OF CLAYS

Profile #1 — A₂ Horizon — Endothermic reaction at low temperature (100°–200°C.) (figure 7) is common to minerals such as illite, montmorillonite, and other 2:1 minerals (Kerr, *et al.*, 1949). A small endothermic peak at 350°C. is probably contributed by gibbsite (Bowen, 1954); another peak at 510°C. has not been identified. An endothermic reaction at 600°C. is due to illite (Bowen, 1954; Kerr, *et al.*, 1949; Grim, 1953). A small endothermic peak at 800°C. is believed to belong to vermiculite which shows also an exothermic reaction at about 800°C. (Grim, 1953). Illite also shows an exothermic reaction at 930°C. A small endothermic peak at 980°C. may be ascribed to kaolinite (Kerr, *et al.*, 1949; Bruijn and Van der Marel, 1954).

B₁ Horizon — This horizon still manifests the presence of vermiculite with a small endothermic reaction at 850°C., although illite may be responsible for this peak. Illite produces an endothermic reaction in the 600°C. region and an exothermic peak

at 950°C. Gibbsite is present with an endothermic reaction at 350°C. circa (Bowen, 1954). Kaolinite is present with a small endothermic reaction at less than 400°C. The 950°C. exothermic reaction for kaolinite is probably masked by the illite peak occurring at the same temperature.

B₂ Horizon — A typical thermogram of illite appears at this horizon. The 600°C. endothermic peak and the 950°C. exothermic peak are well pronounced; the endothermic peak at 850°C. is less clear and the gibbsite peak is also very poorly marked at 350°C.

C-D Horizon — This horizon shows virtually the same features as the previous horizon, with the exception of a more marked peak for the exothermic reaction at 950°C. Gibbsite is probably not present.

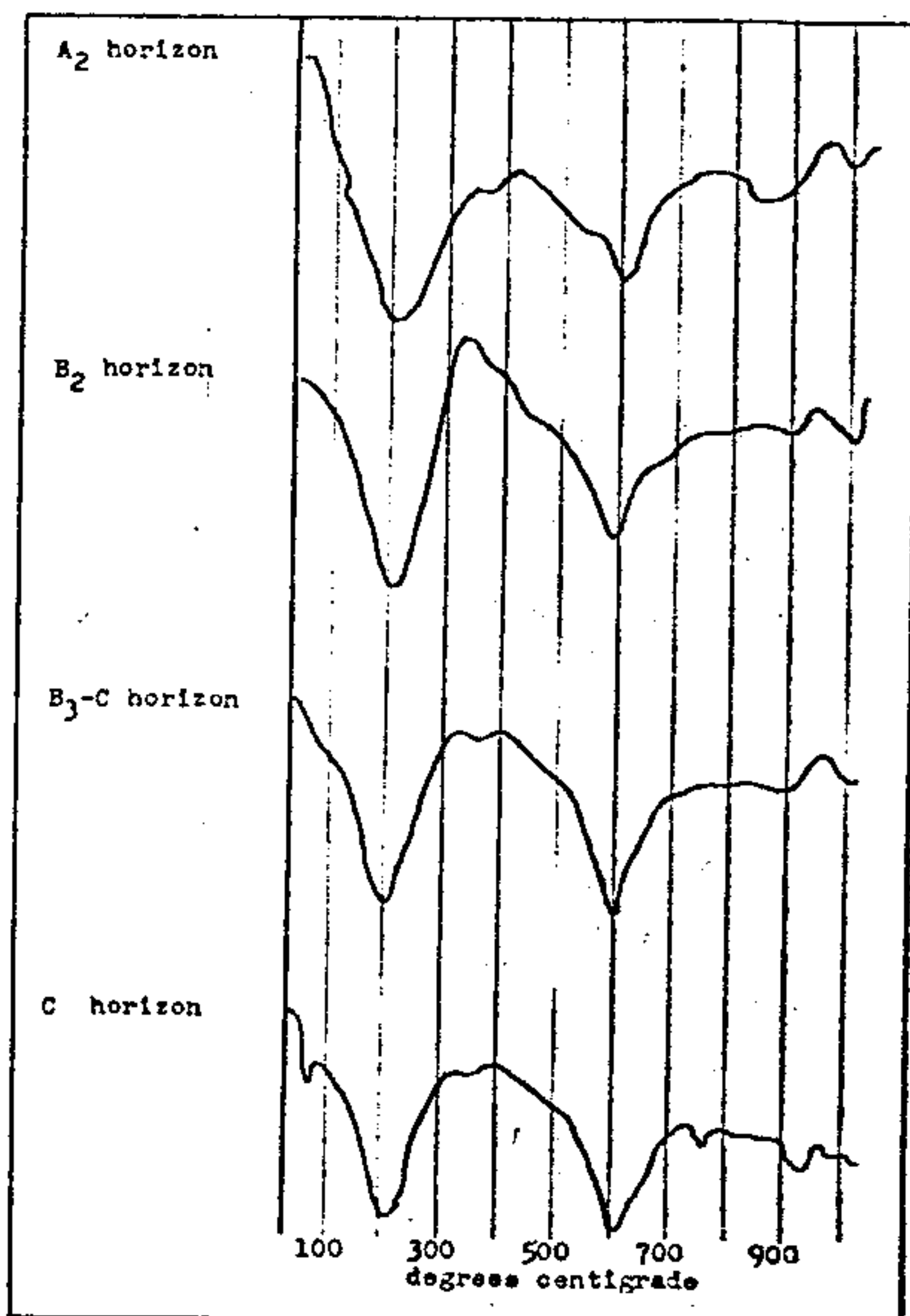


Figure 8. Differential thermograms of <0.002 mm. clay from profile #2.

Profile #2, figure 8. — Vermiculite is present in the A₂ horizon in minor quantities; a peak at 820°C. may be ascribed to the exothermic reaction of this mineral. Illite shows endothermic peaks in the 450, 600 and 850°C. regions and the exothermic one at 950°C. A small inflection at about 350°C. is from gibbsite. Kaolinite does not show any other charac-

teristics with the exception of a possible exothermic peak at 950°C. in concomitance with illite.

B₂ Horizon — Persistence of vermiculite can be detected from the endothermic reaction at 750°C. and from the exothermic one at 850°C; both are very weakly expressed. Illite is present with three endothermic reactions at 450, 580–600°C. and at 900°C. The exothermic reaction is present at 950°C. Gibbsite shows the endothermic peak at 350°C. Kaolinite is probably present.

B₃-C Horizon — Vermiculite does not show any thermo-features in this horizon. Illite appears to be the predominant mineral. Endothermic and exothermic reactions occur in the same positions as in the previous horizon. Gibbsite gives an endothermic reaction at 350°C. only slightly perceptible. No reactions attributed to kaolinite were recorded.

C Horizon — Illite still prevails in this horizon with the 400°C. and 900°C. endothermic reactions slightly more marked than in the B₃-C horizon. A sharp exothermic peak ascribed to illite occurs at 950°C. The endothermic reaction for gibbsite is weakly manifested.

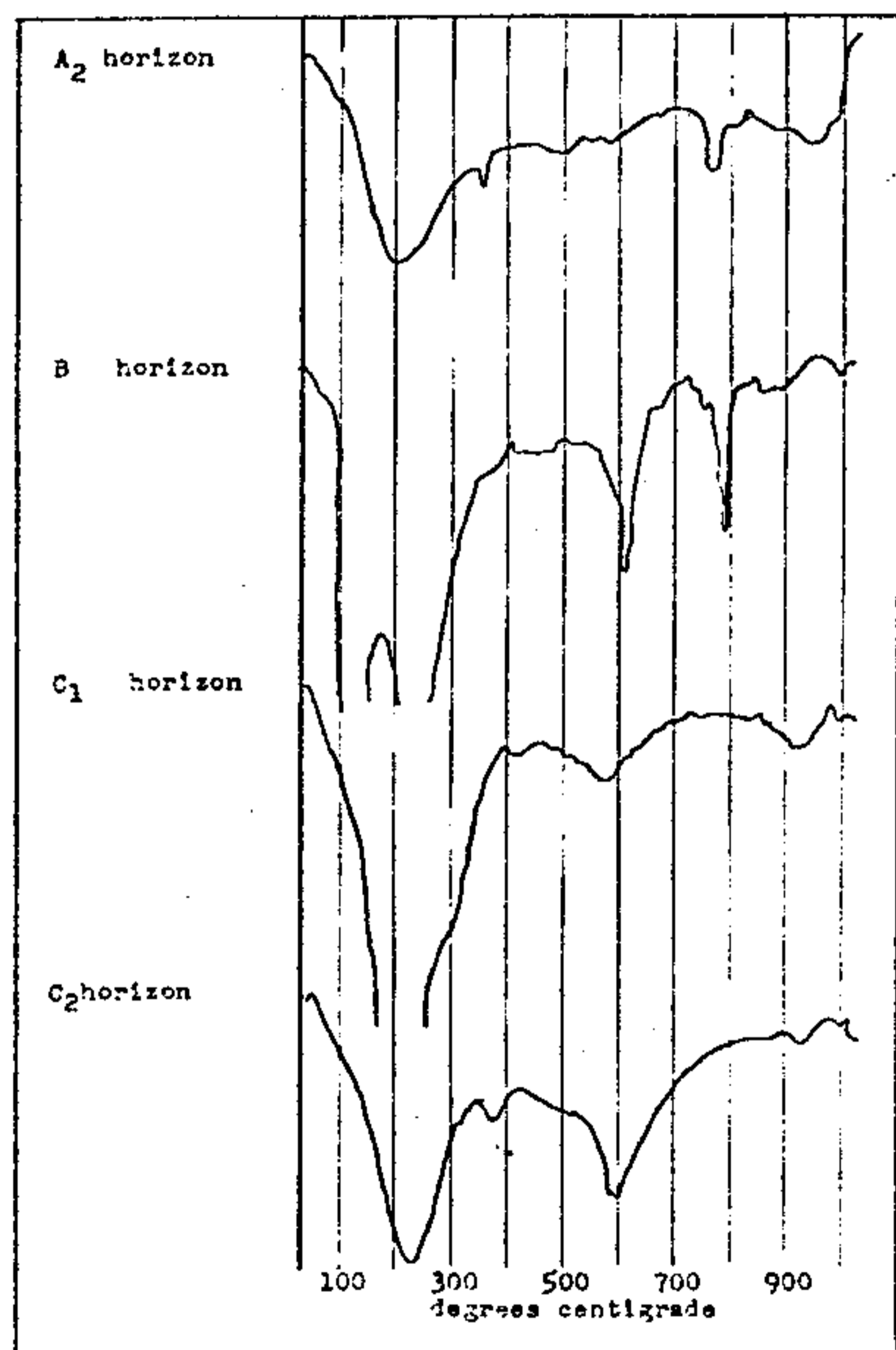


Figure 9. Differential thermograms of <0.002 mm. clay from profile #3.

Profile #3, figure 9 — A₂ Horizon — The thermogram of the A₂ horizon shows the presence of chlorite. Diagnostic reactions of this mineral are the endothermic reaction at 750°C. and the exothermic reaction at 820°C. Illite has only a small endothermic peak at 550 and 950°C. A sharp peak at 350°C. is ascribed to gibbsite. Other clay minerals are possibly present but are not quantitatively sufficient to produce distinct reactions.

B Horizon — According to the features of this thermogram, this horizon should contain a large quantity of chlorite but this is not confirmed by the X-ray method. Illite is present with possibly three endothermic and one exothermic peak. Kaolinite is probably present but masked by illite. A "shoulder" shown at 350°C. is the result of gibbsite.

C₁ Horizon — No definite peaks can be attributed to chlorite. Illite is represented by the already mentioned reactions. The exothermic peak at 950°C. is sharper compared to the one of the previous horizon. Kaolinite is probably absent.

C₂ Horizon — Illite is evident in this horizon. The 450°–500°C. endothermic reaction is slightly perceptible, the 600° and 920° are more evident. A double peak occurs at 950° and 1000°C. as exothermic reaction. Gibbsite is here indicated by the endothermic loop at 360°C. Kaolinite is probably absent.

ELECTRON MICROSCOPY⁹

The clay fraction (<0.002 mm.) of profile #1 was examined and photographed under the electron microscope. The instrument used was a RCA model EMU-2A. The samples were magnified 10,000, 15,000, 18,750 and 25,000 times. The electron micrographs were compared with micrographs of standard clays. (Davis *et al.*, 1950; Grim, 1953; Bates, 1958.)

The electron micrographs (figure 10) confirm the presence of some of the clay minerals already established through X-ray diffraction analysis. Illite present as crystal aggregates and flakes is the dominant mineral; halloysite represented by tube-like particles may also be present. There is no evidence for vermiculite. Kaolinite is probably shown by subeuhedral flakes.

Some of the illite flakes tend to show an hexagonal outline; this characteristic seems to prevail in the lower horizons indicating and confirming that the degree of crystallization of this mineral increases with depth.

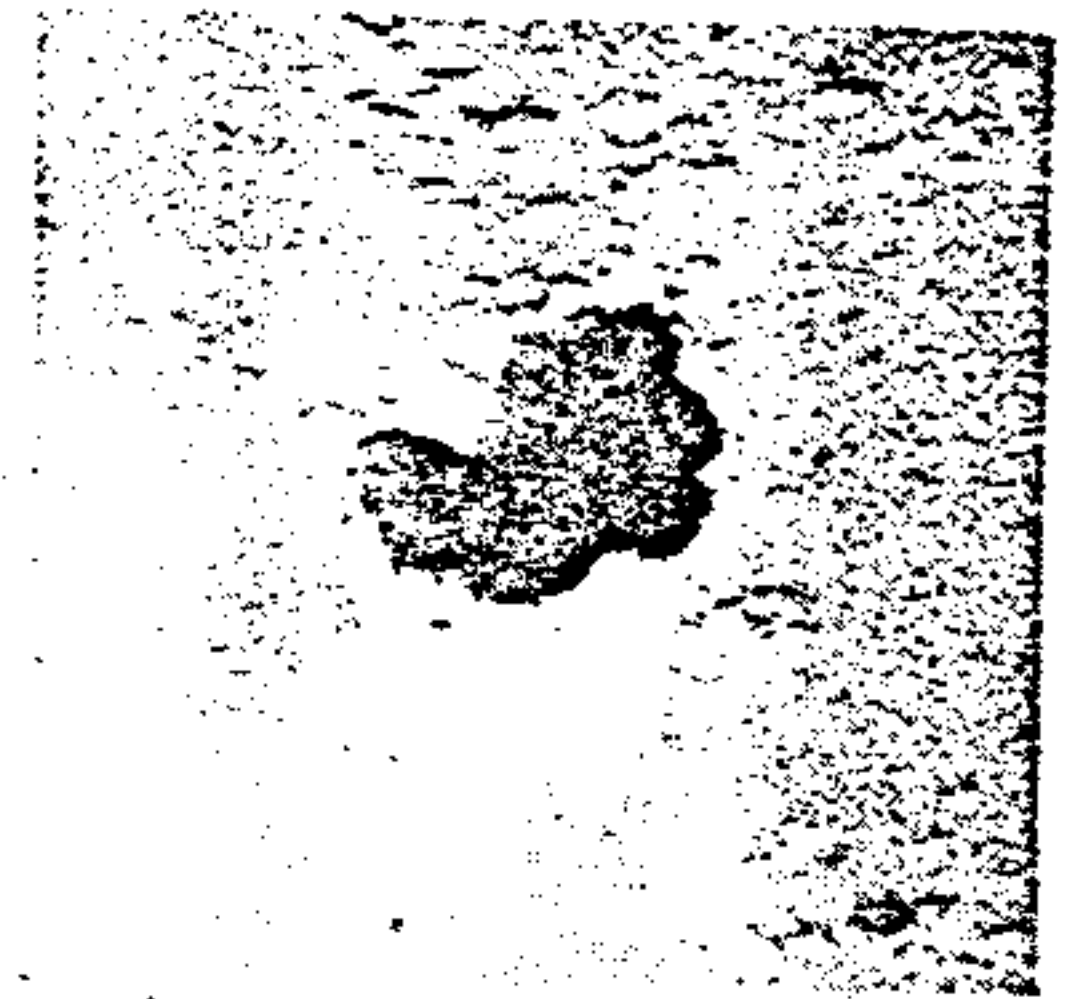
CHEMICAL ANALYSIS

Results of the total analyses of the soil of profile #1 and Brunswick shale are shown in table 10; those of the partial analyses of the clay of the same profile in table 11. Silica is uniformly distributed in the A₂ and B horizons, the same can be said for

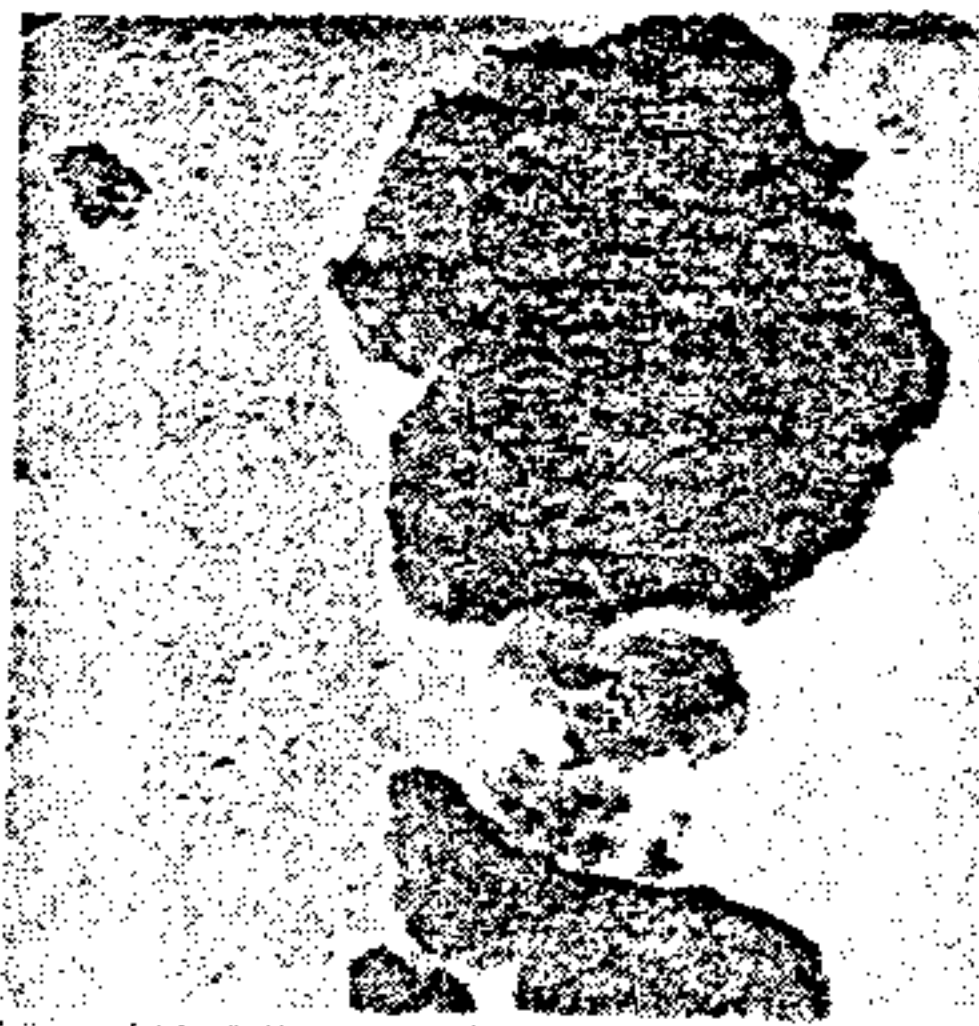
9 Micrographs by Prof. Pauline Holbert.



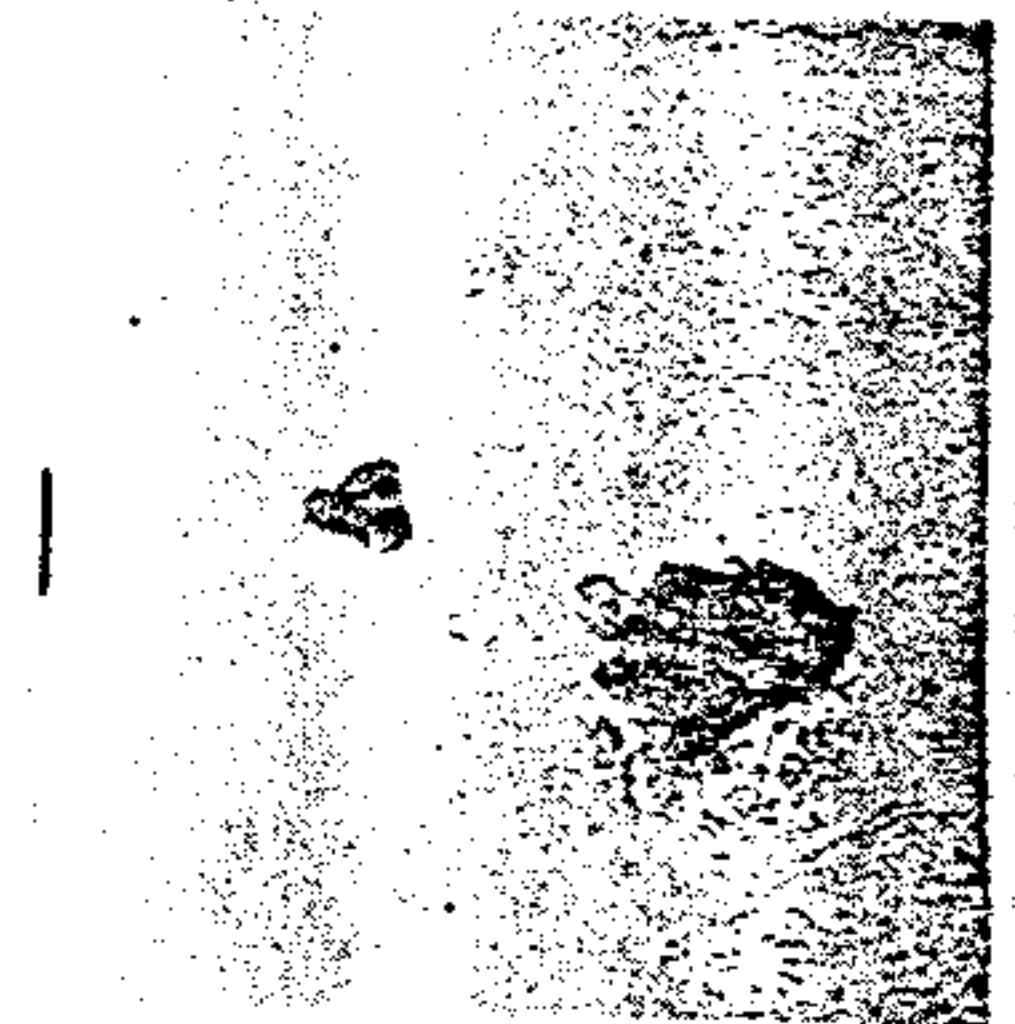
A₂ 10,000x



A₂ 15,000x



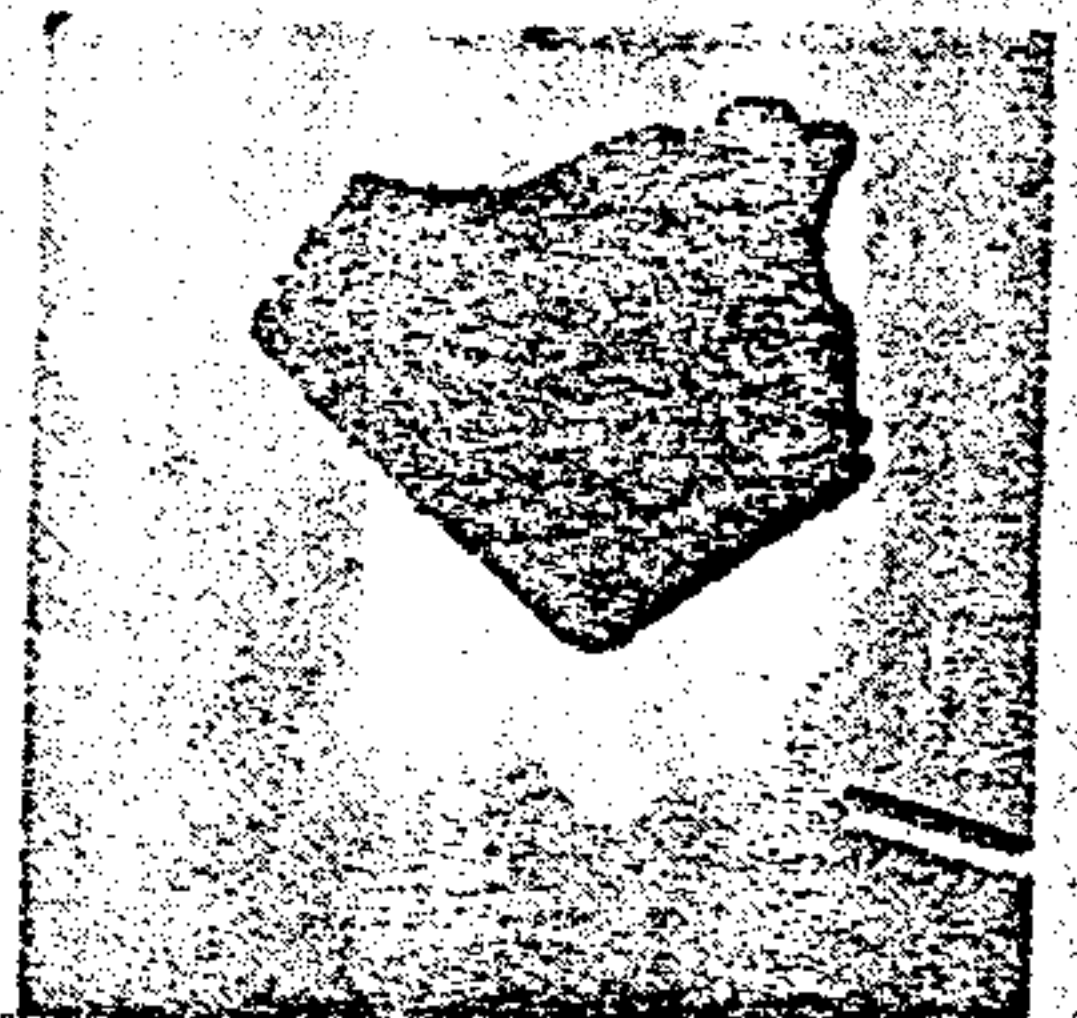
B₁ 15,000x



B₁ 15,000x



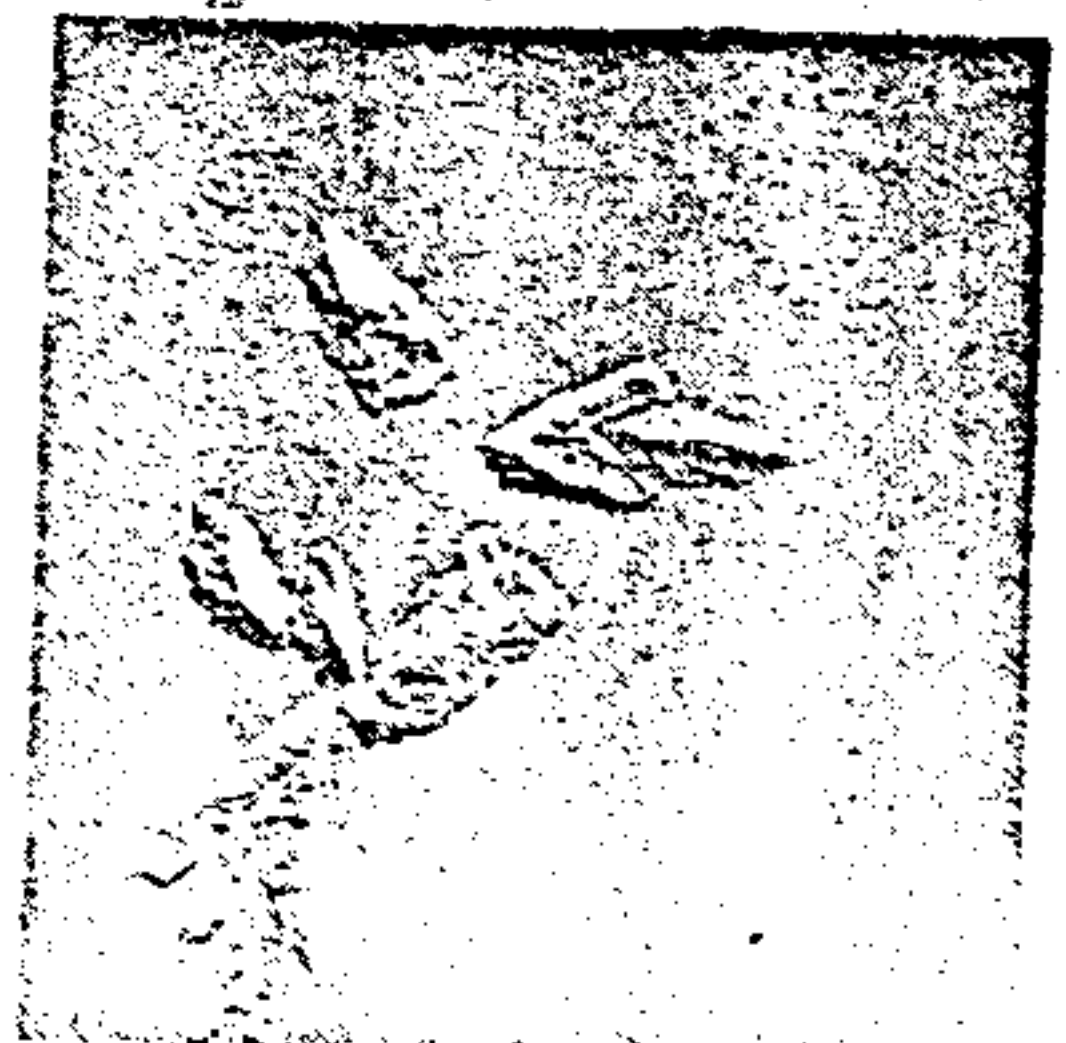
B₂ 15,000x



B₂ 15,000x



C-E 8,300x



C-E 15,000x

Figure 10. Electron micrographs of clay fraction (<0.002 mm.) of the four horizons of profile #1. Microscopy by Prof. Pauline Holbert.

Table 10. Total Chemical Analyses of the Soil (<2 mm.), and Bedrock. Profile #1.

Horizon	Depth*	SiO ₂	Fe ₂ O ₃	Al ₂ O ₃	TiO ₂	P ₂ O ₅	CaO	MgO	Na ₂ O	K ₂ O	Loss of Ignition 600°C.	SiO ₂ **	
												R ₂ O ₃	Al ₂ O ₃
	ins.	pct.	pct.	pct.	pct.	pct.	pct.	pct.	pct.	pct.	pct.		
A ₂	2-6	76.55	4.35	8.23	0.91	0.10	0.35	0.60	1.32	1.71	6.24	11.1	14.5
B ₁	12-20	78.28	5.37	8.51	0.90	0.64	0.53	1.14	2.17	2.04	3.58	11.2	15.6
C-D	25-30	66.33	7.97	13.77	0.97	0.94	0.52	0.73	1.52	3.56	4.25	6.0	8.2
Brunswick													
shale	30+	51.52	10.44	19.81	0.97	0.61	0.90	1.06	1.04	6.57	6.82	3.3	4.4
A ₂ clay	2-6	41.37	9.15	25.10	0.99	—	0.99	0.70	0.98	2.50	17.36		

* Depth of sampling

** Molar ratio

Table 11. Partial Chemical Analyses of the Clay (<0.002 mm.). Profile #1.

Horizon	Depth*	SiO ₂	Fe ₂ O ₃	Al ₂ O ₃	Loss of Ignition 600°C.	SiO ₂ **		Fe ₂ O ₃ ***
						R ₂ O ₃	Al ₂ O ₃	
	ins.	pct.	pct.	pct.	pct.			pct.
A ₂	2-6	39.11	11.43	24.73	18.77	2.1	2.7	5.00
B ₁	12-20	45.75	14.43	26.25	10.64	2.2	2.9	7.20
C-D	25-30	44.12	13.66	26.12	8.21	1.9	2.5	7.50

* Depth of sampling

** Molar Ratio

*** Removed from the clay by Jeffries' method; calculated on total clay basis.

both Al and Fe. Silica sesquioxides molar ratio shows evidence of this condition. A decrease in SiO₂ is noticed in the C-D horizon which in turn shows an increase in Al and Fe hydroxides. The Brunswick formation as compared to the soil manifests a further decrease in SiO₂ and an increase of Fe₂O₃ and Al₂O₃. Partial analysis of the clay confirms the characteristics already partially disclosed by the soil analysis.

CATION EXCHANGE CAPACITY

Cation exchange capacity of profile #1 (table 12) fluctuates between three maxima: at the surface, in the B₁, and in the C-D horizon. The variations among these values are not pronounced. A rather high value is obtained from the C-D horizon which

may be considered unusual as far as the general trend of the cation exchange capacity goes. Exchangeable cations show Ca as the prevailing cation followed by the Na and K ions. These values expressed in millequivalents are rather low for the three cations.

pH VALUES

The measured pH on a one-year basis are presented in table 13 and figure 11. Annual variations are not very pronounced, although there is a tendency to follow a certain pattern throughout the year. The highest pH values were recorded during the winter and summer months; the lowest values occurred during the spring and fall.

Table 12. Chemical Data of Profile #1.

Horizon	Depth*	Cation* ^o exchange capacity m.e./100g.	Exchangeable cations: m.e./100g.			O.M. pct.	C pct.	N pct.	C:N pct.
			Ca	K	Na				
A ₂	2-6	8.5	0.50	0.10	0.20	4.18	2.09	0.15	13.4
B ₁	12-20	5.7	0.53	0.12	0.15	2.68	1.34	0.06	21.3
B ₂	20-24	9.2	0.67	0.20	0.23	1.84	0.92	0.51	17.9
C-D	26-30	9.0	0.64	0.18	0.20	0.21	0.11	N.D.	N.D.

N.D. Not detected

* Depth of sampling

SOIL ORGANIC MATTER TOTAL N AND C:N RATIO

Tables 12, 14, and 15. A steady decrease of organic matter with increasing depth is observed in profile #1. Total N follows a similar pattern. The C:N ratio is lowest at the surface; it increases in the B₁ horizon and decreases again in the B₂. An increase of almost 2 percent in the organic matter content is recorded in the A₂ horizon of profile #2 with respect to profile #1 for the same horizon. Organic matter and total N decrease with depth; the C:N ratio maintains a pattern similar to profile #1. Organic matter and N achieve lower levels in the case of profile #3. Both components register a reduction in value with increasing depth; the C:N ratio shows a similar tendency.

MOISTURE, PERMEABILITY AND BULK DENSITY

On July 30, 1957, a minimum percent of moisture was recorded in the soil of the forest. A small amount of precipitation during August affected mostly the A₁ horizon and the first 3 inches of mineral soil. During the course of sampling, only slight variations were registered up to 12-15 inches; below this depth, moisture varied almost negligibly throughout the soil.

Values of bulk density determined in the area corresponding to profile #1 and the adjacent field are presented in tables 16 and 17 and figure 12. Due to mechanical disturbance, the cultivated field has a higher bulk density than the undisturbed woodland. This condition is also reflected in the permeability of the two soils. Rate of infiltration expressed in inches per hour is lower in the field and higher in the forest. Infiltration decreases with increased depth in the forest soil, whereas the opposite is true in the field (tables 16 and 17 and figure 13).

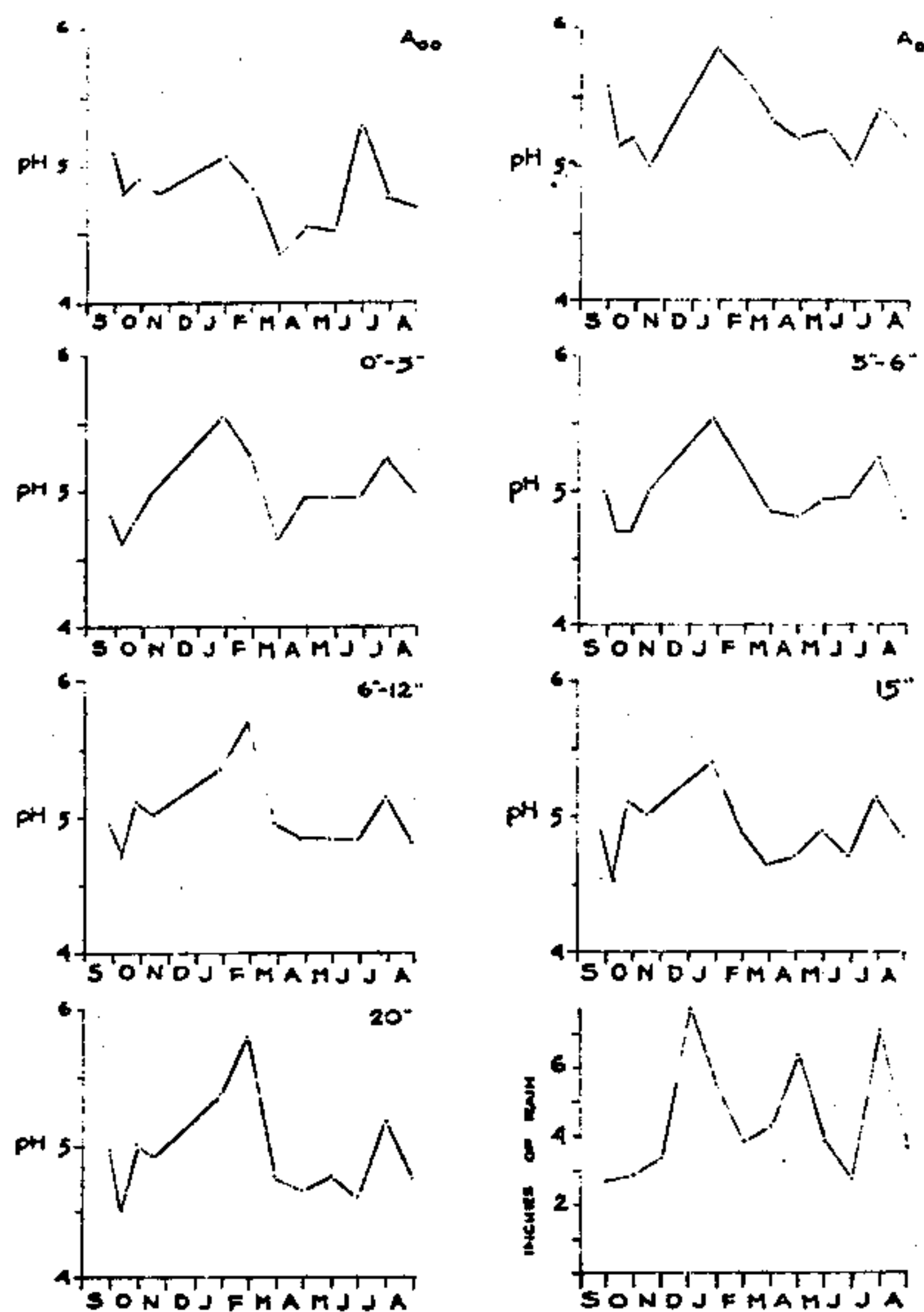


Figure 11. Monthly distribution of pH according to soil depth and monthly distribution of rainfall (Sept. 1957—Aug. 1958).

Table 13. Soil pH values at Various Depths and Time Intervals* 1957-1958.

Depth	Sept.	Oct.	Nov.	Jan.	Feb.	March	April	May	June	July	Aug.
A ₀₀	5.1	4.8	4.9	4.8	5.1	4.8	4.4	4.5	4.5	5.3	4.7
A ₁	5.6	5.2	5.2	5.0	5.9	5.6	5.3	5.2	5.3	5.0	5.2
0-3"	4.8	4.6	4.8	5.0	5.6	5.3	4.6	5.0	5.0	5.2	5.0
3-6"	5.0	4.7	4.7	5.0	5.6	5.2	4.9	4.8	5.0	5.3	4.8
6-12"	5.0	4.7	5.1	5.0	5.4	5.0	4.9	4.9	4.9	4.8	4.8
15"	4.9	4.5	5.1	5.0	5.4	4.9	4.7	4.7	4.9	4.7	4.9
20"	5.0	4.5	5.0	4.9	5.4	5.8	4.7	4.7	4.8	4.6	4.8

* Samples taken near profile #1

Table 14. Chemical Data of Profile #2.

Horizon	Depth*	pH	O.M.	C	N	C:N Ratio
	ins.		pct.	pct.	pct.	
A ₂	2-5	5.1	6.10	3.05	0.23	13.2
B ₁	10-14	4.9	2.40	1.20	0.08	15.8
B ₂ -C	16-20	4.6	0.66	0.33	0.07	5.0
C-D	22-	4.9	0.66	N.D.	N.D.	N.D.

N.D. — not detected

* Depth of sampling

Table 15. Chemical Data of Profile #3.

Horizon	Depth*	pH	O.M.	C	N	C:N Ratio
	ins.		pct.	pct.	pct.	
A ₂	2-10	6.0	1.42	0.71	0.04	17.7
B	15-25	5.6	0.75	0.37	0.03	12.3
C ₁	34-40	5.7	0.40	0.20	0.03	6.6
C ₂	42-46	5.5	0.36	0.18	N.D.	N.D.

N.D. — not detected

* Depth of sampling

Table 16. Bulk Density Values and Infiltration Rates in the Forest.*

Horizon	Depth ins.	Bulk Density	Infiltration Inches Per Hour
A	3-6	1.07	1.89
B	13-16	1.42	0.60
C	20-23	1.48	0.54

* Profile #1

Table 17. Bulk Density Values and Infiltration Rates in the Cultivated Field.

Depth ins.	Bulk Density	Infiltration Inches Per Hour
1-3	1.30	0.38
6-9	1.54	0.54

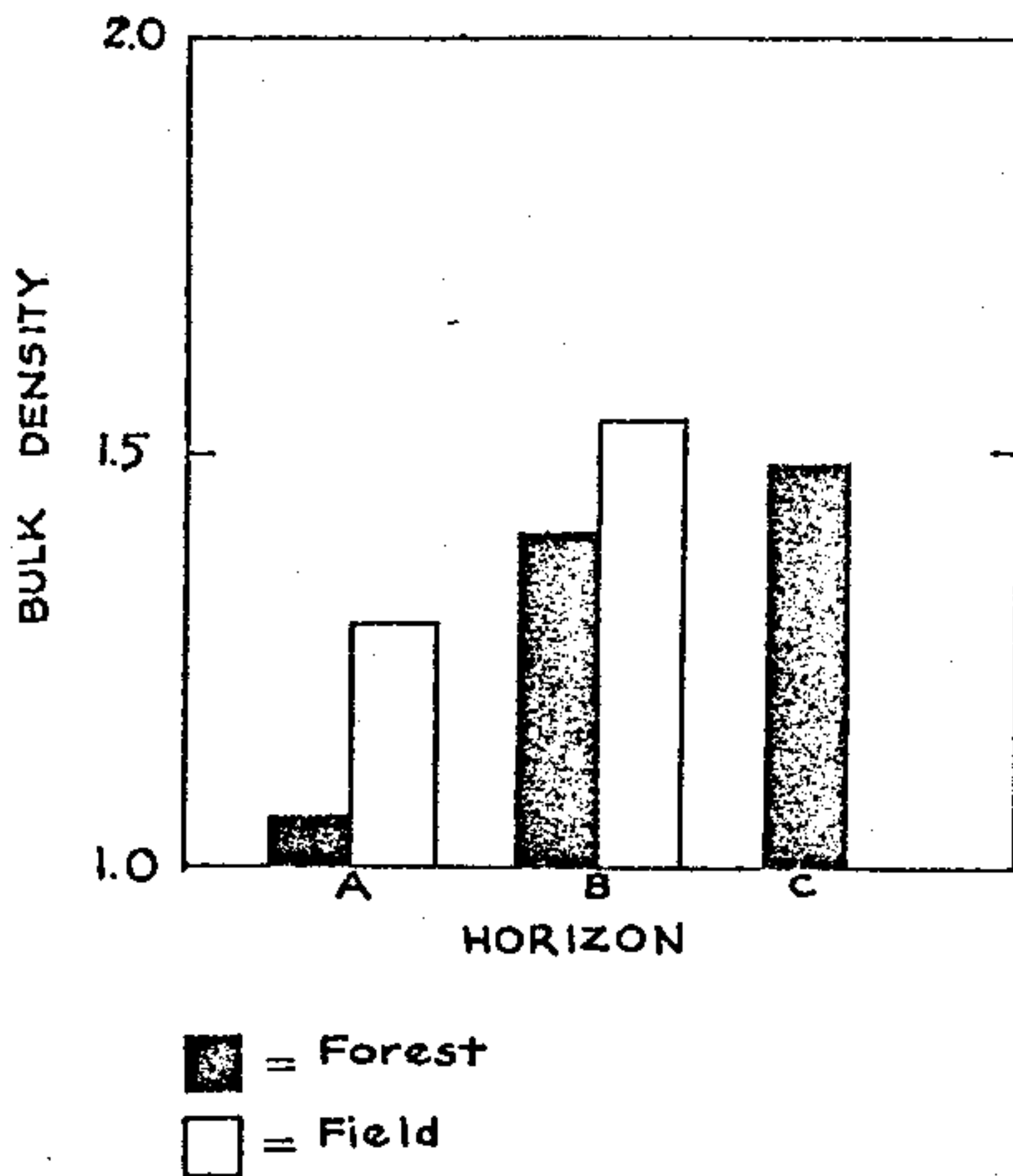


Figure 12. Bulk densities of forest and field soils.

DISCUSSION

GENERAL

Glinka coined the expression andodynamorphic soils to distinguish those soils in which the internal characteristics "chemical composition and physical character of the parent rock interrupt the development of the soil type that would take place normally under the action of the external forces."¹⁰ The term has not, however, acquired a wide usage among American pedologists. One of the reasons is probably due to its long orthography whereas the term

¹⁰ Glinka, K. D., 1927. Great soil groups of the world and their development. Translated by C. F. Marbut, Ann Arbor, Edwards Bros., p. 33.

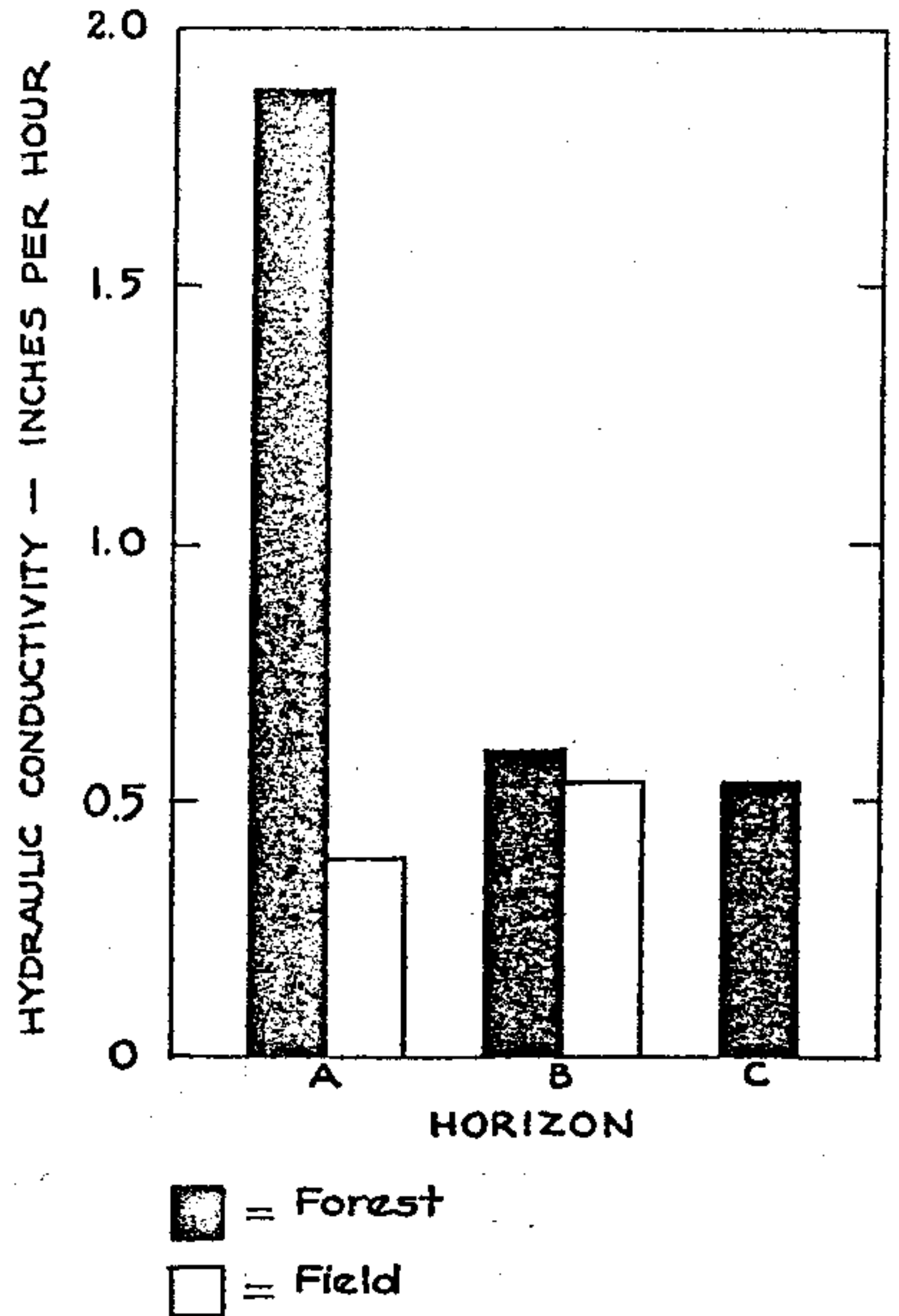


Figure 13. Hydraulic conductivities of forest and field soils.

"lithogenic" expresses a similar meaning. The greatest objection to this expression is that in Glinka's classification, the endodynamorphic soils and the ektodynamorphic ones comprise the first division in the classification of soils (Glinka, 1927). Since ektodynamorphic soils are by far more numerous than endodynamorphic, the student of soils is faced with a picture of a balance whose beams are unproportionally loaded. Of course, when Empedocle of Agrigento (V cent. B.C.) laid down the four principles of the world, later adopted by Aristotle, he had water, air, earth, and fire. Undoubtedly, the question of quantitative equivalence was not considered in his choice. Also, when the biologists divided the animal kingdom into Protozoa and Metazoa, the number of types of each group was not considered important for the finality of the classification.

Ramann considered climate the main factor in the development of soil; he admitted, however, the existence of factors which change the properties of a soil to a considerable extent. Soils whose

properties have been derived from these factors are called "local" soils. One of these factors is the parent material.

"The nature of the parent rock exercises less influence the more extreme the climate. In Eastern Europe Loess, Clays, Granite and Limestone are all uniformly covered by the Black Earths. The more temperate the climate the greater is the effect produced by the parent rock on the soil properties, both chemically and physically."¹¹

In this connection, Ramann mentioned that Fallou, who was active in Central Germany, made his classification on the basis of the parent rock because in that part of the country the climate produces only a moderate weathering and the soils thus do not acquire outstanding features (Ramann, 1928). Polynov (1927), father of paleopedology, considers four categories of soil and soil material that have been affected by paleophenomena. The fourth category, products of ancient weathering, deals with substrata whose characteristics are connected with geological climate. Thus, the red-lands found in U.S.S.R. are regarded as remnants of Tertiary laterites. The soils in Hutcheson Memorial Forest, especially those derived from the Brunswick shale, offer an example of soils whose genesis has been partially dictated by the underlying bedrock. Their characteristics are associated with a Triassic climate rather than with the present under which the soils have developed. The persistence of the typical red color of the shale in the solum, the lack of differentiation into horizons, the limited translocation of certain compounds are all indices of a retarded manifestation of the pedogenetic factors. Such soils in the words of Glinka, Ramann and Polynov would be called, respectively, endodynamorphic soils, local soils, and paleosoils. It is not intended to give the idea that these three terms are synonymous, and therefore interchangeable; but ironically enough each one could be used in the case of Hutcheson Memorial Forest soils with justification. These soils may be considered endodynamorphic inasmuch as the Brunswick shale had played a paramount role in their formation. They may be regarded as local soils because in Nova Scotia, under a colder climate, soils formed on Triassic shales display more podzolic characteristics,¹² and finally they are in some respect paleosoils because some of the features present in these soils are related to conditions existing during Newark epoch. It is, therefore, felt that great importance must be attached to the nature of the parent material, its age, its origin, and mode of formation. All these characteristics become even more significant since we are confronted with a sedimentary rock which represents the final product of many different processes. Any speculation on the

behavior of these rocks must be, therefore, related to the history of the rock itself. This will involve the determination of the source of the sediments, the mechanism of dispersal of the removed material, the mode of transportation, the medium in which the sedimentation occurred and, ultimately, the diagenesis and lithification after deposition (Pettijohn, 1949).

Krynine (1949) is the fautor of humid temperate climate (Savanna) which prevailed during the Newark epoch; Raymond (1927) and Dorsey (1919) also share this opinion. Krynine (1949), in his classification of the Red beds, considers the Triassic sediments primary red beds which implies that the red pigment was the result of weathering which occurred in the source area of the future sediment at the expense of iron-bearing minerals (silicates and carbonates) and later it was incorporated directly into the resulting sediments. The prerequisite necessary and sufficient to produce a red soil is, that the type of weathering is conducive for a red end-product. Such a requirement is fulfilled when a silicate is exposed to an average annual temperature above 60°F. and over 40 inches of annual precipitation. The bulk of our present day red soils (95 percent according to Krynine) are formed where annual temperature is 60°F. or above, and precipitation is 25 or more inches. "Most of the 5 percent of so-called non-humid red soils were not formed at the present time, but were inherited from the pluvial Pleistocene period. The so-called red sands of the desert are mostly fiction insofar as the Recent goes, are practically always produced by the reworking of the pluvial Pleistocene red soils or of ancient red sediments."¹³ The preservation of the red color in the sediment is due only to one factor, that is the persistence of oxidation over reduction reactions, a condition that can occur in any climatic condition. If this is the case, then the Brunswick formation is nothing more than a reworked product of the Triassic "soils". Terra Rossa soils form under climatic conditions which approximate 30 inches of mean annual rainfall and a mean temperature of 60°F. These soils are very common in the Mediterranean region and possess a striking red color. In addition to the coloration in the Triassic sediments, there are other characteristics in common with Terra Rossa soils. The predominant clay of Terra Rossa is illite (Cecconi, 1955), the $\text{SiO}_2/\text{R}_2\text{O}_x$ molar ratio is 1.8. In the Brunswick shale the ratio approximates 3.3. The Fe:Al ratio in the Mediterranean soils is 1 to 2, in the shale it is 1 to 2 (Comel, 1933). Although the existing similarities tempt one to regard the soil on the flanks of the Triassic hills similar to Terra Rossa, a major difference exists between Terra Rossa and the Newark sediments. While Terra Rossa is very sensitive to

¹¹ Ramann, E., 1928. The evolution and classification of soils. Translated by C. L. Whittles, Cambridge, Heffer, p. 25.

¹² Tedrow, J. C. F. Personal communication.

¹³ Krynine, P. D., 1949. The origin of red beds. Trans. N. Y. Acad. Sci., V. 2, p. 61-62.

climate changes, the red sediments are not. In Italy, in the more humid sections, a similar type of red substratum as is present in the Newark basin is changed to a braunerde, and in the high mountains a podzol is found (Comel, 1933). Triassic sediments in New Jersey are less susceptible than Terra Rossa to environmental variations. In fact, in New Jersey, a general area of podzolized soils, the soils derived from the red shale, do not conform to this pedological characteristic.

Despite what has been said about the lack of a decisive type of soil formation in the soils of Hutcheson Memorial Forest, certain qualitative processes have taken place. Colorwise, the A, B, and C horizons show minor differences. Clay accumulation is present in the B horizon, iron compounds which exist as coatings appear to be more abundant in the lower than in the surface horizons. Clays appear to be more highly altered at the surface than they are at the depth. This indicates that red shale soils are not entirely a monotonous layer of soil material without genetic features.

MECHANICAL ANALYSIS

Mechanical analyses of profiles #1 and #2 (tables 1, 2 and figure 4) definitely show the presence of an illuvial horizon. In spite of this similarity, the trend of the silt and clay is not the same in the two profiles. In profile #1, an accumulation of clay in the B₂ is followed by a decrease in C-D. Profile #2 shows, instead, an increase of clay in both the B and C. Silt behaves somewhat differently in that more is present in profile #1 than profile #2. Profile #2, on the other hand, manifests a sharp reduction of silt in the lowest horizon. If it is assumed that silt represents the potential source of the clay, then some implications about the degree of weathering in the two profiles can be drawn. In profile #1 the process of weathering has been disturbed by a mixing action, whereas in profile #2 there has been sufficient time for the residuum to develop a more normal weathering condition. Therefore, in profile #1 the weathering of the silt to clay is still in progress, whereas in the second profile such transformation has already been effected. In addition to this, some windblown material may have affected the area of profile #1. (Tedrow and MacClintock, 1953.)

The main feature of profile #3 (table 3 and figure 4) with regard to particle size distribution is the preponderance of sand fractions, compared with the other separates. One observes that among sand fractions, medium and fine sands prevail. Horizons C₁ and C₂ which show a decrease in sand content and an accumulation of silt and clay, are probably not genetically related to the solum. Silt and clay found here are not the product of a pedogenic translocation but rather the result of water deposition. Sandy

material superimposed on these two horizons was also deposited by fluvial process. Horizon B shows an illuvial quality manifested by the concentration of silt and clay.

HEAVY MINERALS

Heavy mineral suites of the two profiles derived from the Brunswick shale show a number of mineral species that can not be ascribed to the local Triassic material (tables 4, 6 and figure 5). The existence of a stratum of unconsolidated rocks deposited during the Pleistocene on top of paleosoils is therefore highly possible. Petrographic analysis in addition to disclosing the existence of such surface deposit, shows also mixing action that has taken place between those two strata. Minerals, such as hornblende, hypersthene and garnet while not present in shale are found in the soils indicating that some of the surface material had been transported from other areas and subsequently mixed. Mixing is ascribed to the mechanical disturbances produced by falling trees (Lutz, 1940), root channels (Gaiser, 1952), burrowing animals wind activity and frost processes. A prominent increase of some minerals at certain depths as with the case of profile #1 cannot always be explained. A concentration of muscovite in the B₂ horizon of profile #1 can be partially ascribed to a segregation of this mineral from the shale fragments because of the close proximity of this horizon to the bedrock. Sturm (1956) has pointed out that the disk-shaped fragments of weathered shale tend to break along their short diameters. The two sides of the broken piece reveal a cover of muscovite flakes which can be released in the soil. These laminations produced by the muscovite had their origin during deposition. Micaceous flakes, together with clay particles, would settle together due to the hydrolic equivalence of the two and would form the upper part of each micro-sedimentation unit. The disappearance of muscovite at the surface is probably due to weathering; this is further substantiated by X-ray work and by the appearance of vermiculite as a prominent clay. Hornblende also offers an accumulation in the lower horizons, where it occurs in corroded grains indicating an acid weathering (Van der Marel, 1949).

Heavy mineral suite of profile #2 shows, in comparison with profile #1, a noticeable decrease in mineral species indicating, therefore, a more residuum condition. It is also important to observe that the mineral components decrease with depth; this behaviour is interpreted as the result of a less mixing effect. The decrease of mineral species in profile #2 may give information about the provenience of the sand mantle. Since the mineral species decrease from profile #3 (the westernmost) to profile #2, it may be inferred that the original sand deposit was decreasing in thickness in the eastern direction. This would imply that the sand deposit

now found in the western part of the forest once extended over most of the woods. The consistency of certain minerals in the three profiles may validate this hypothesis. Preponderance of certain heavies such as black opaques, zircon and tourmaline demonstrate the maturity of the sand deposit, whose origin can be postulated from the association with detrital minerals of different types of rocks. Metamorphic acid and basic igneous rocks may be considered as the source of this deposit. Light minerals of profile #1 and #2 bring more evidence regarding the degree of mixing of the two profiles. In fact, the quartz grains persisting in the C-D horizon of profile #1 are substantially reduced in the corresponding horizon of the second profile. The aggregates are largely made up of shale fragments and increase with depth in both profiles. Heavy minerals equivalent in size to a very fine sand separated from ground Brunswick shale were collected from profile #1 and petrographically analyzed. These heavies consisted of stained fragments which, with crossed nicols, showed a colored fringe. According to Martens¹⁴ we are dealing with fragments of muscovite which have been impregnated by a hematitic pigment. When these fragments are seen with incident light, they appear as little aggregates rather than flakes. Stained muscovite is also found in minor quantity. Minerals showing magnetic properties obtained from the ground shale were observed microscopically. A comparison of these magnetic minerals with the corresponding ones separated from various horizons of profile #1 proved very rewarding. The magnetite in profile #1 is definitely detrital, whereas the magnetic minerals in the Brunswick shale are made up of aggregates impregnated with magnetic substances. This qualitative difference between the two kinds of magnetic material helps to prove again the heterogeneity of the soil material constituting the profile. On the basis of these findings, it can be concluded that profile #1 and profile #2 are partially two-story profiles.

CLAY MINERALS

X-Ray Analysis. The clay mineral suites of profiles #1 (figure 6) and #2 are very similar as far as components are concerned. Clays of profile #3 are somewhat different. The most prominent clay of the first and second profile is an illite-mica which by alteration produces a 14 Å mineral identified as vermiculite. Such alteration is more conspicuous at the surface than in lower horizons. Kaolinite is only a minor component. Illite-mica has been identified as a ferriferous dioctahedral type with a spacing of 1.50 Å for the 060 plane (Brown, 1951).

The illite-mica present in profile #1 is derived from the bedrock from which it has been released.

This clay has undergone a major transformation since its appearance in the soil and a new mineral, vermiculite, has been produced by the weathering of the illite-mica. Since vermiculite has formed at the expense of illite-mica in the horizons where the former prevails, the latter is drastically reduced. The disappearance of illite-mica is manifested by the weaker and broader basal diffraction lines. The degree of crystallinity of such minerals increases, therefore, with depth. Illite-mica represents, according to Jackson, *et al.*, (1948), an intermediate stage in the weathering sequence. Progress in the weathering of this mineral is represented by the appearance of hydrous mica and by a montmorillonoid mineral. Such degradation is observed in the A₂ and B₁ horizon of this profile. The broad 10 Å line observed at the surface tends to become sharp by heating, demonstrating that the hydration of mica is due to molecules of water between the sheets (Stephen, 1952). According to Jackson, *et al.*, (1948), the release of the fixed K in mica is the starting point of its alteration. Brown (1951) offers a theory about the weathering of illite. A decrease in potassium is the first step in this process, followed by adsorption of water which causes the mineral to expand. Starting with an illite, a montmorillonoid mineral is later produced. When vermiculite is associated with mica, it owes its formation to a similar process. The appearance of vermiculite as a product of weathering is, on the other hand, more commonly reported in trioctahedral mica than in dioctahedral (Brown, 1951). MacEwan (1949) reports that dioctahedral micas do not show a well-defined stage of weathering as do the trioctahedral, but they alter to a swelling material resembling montmorillonite rather than a vermiculite. Stephen (1952) also reports vermiculite as an alteration product of biotite (trioctahedral mica). Other investigations corroborate these findings (Barshad, 1948; Mortland, 1958, and Mortland *et al.*, 1956). Jackson, *et al.*, (1952) however, observed that in the weathering of mica, illite, "intermediates", vermiculite and montmorin are produced, which implies that a mica can give an expanding 2:1 layer silicate after it has been weathered to a vermiculite. Therefore, the swelling material tested by MacEwan (1949) from a dioctahedral mica is the last stage of weathering. In addition to this, Jackson *et al.*, (1952) relates the chemical weathering processes to the pedogenetic processes. From this comparison it results that under podzolic condition an intermediate dioctahedral illite weathers with preference to vermiculite rather than to montmorillonite. This confirms that a dioctahedral illite alters to a vermiculite under acid weathering. Alteration of illite is further substantiated by the reduced intensity of certain lines. The 1.50 Å and 2.37 Å spacings appear very weak in those horizons where vermiculite prevails. The 1.50 Å and 2.37 Å lines

¹⁴ Martens, J. H. C. Personal communication.

are more intense and sharp where illite prevails over vermiculite. Furthermore, the 10 Å line of the A₂ horizon clay is rather diffused and forms a band between 9.5 Å and 10.5 Å. The band tends to produce a sharp line at 10 Å upon heating the specimen for two hours at 600°C. This, therefore, indicates the state of hydration of the mica, which has been recognized as the prelude to weathering (Brown, 1951; MacEwan, 1949). The presence of vermiculite is somewhat masked when an X-ray picture is taken of the entire clay (< 2 μ). Further subdivision of the 2 μ clay into smaller fractions reveals, however, that vermiculite tends in some instances to be concentrated in the smaller clay fractions. Such behavior is not found at the surface where the 1-0.5 μ particles show an intensity of the 14 Å line only slightly greater than the 2-1 μ fraction. In the B₂ and C-D horizons, vermiculite prevails instead in the smaller fraction (< 0.5 μ). This may be interpreted as the result of the weathering sequence: illite-mica → vermiculite. Longer exposure of the surface horizons to the alteration factors had given enough time for a more complete transformation of illite. In the lower horizons, weathering forces being of a lower magnitude attack only smaller particles (Walker, 1949). Their concentration in the B₂ and C-D horizons may also be due to translocation rather than to weathering. Appearance of some clay skins suggest that some vermiculite may have been translocated from the upper horizons, but this facet was not further investigated. If this is the case, these particles actually originated in the A₂ or possibly in the B₁ horizon. If the two horizons were reintegrated by these fractions, the proposed weathering sequence: illite → vermiculite, would still hold because the < 0.5 μ separates would prevail on the other fractions. The smaller fraction (< 0.5 μ) shows broad and more diffuse diffraction lines than the coarse one, indicating that as the size decreases the degree of crystallization is also reduced. Reduction in size shows also that certain minerals, like illite and kaolinite, become less abundant in the smaller fraction. In fact, both minerals are more prominent in the 2-1 μ size. Quartz follows the same pattern, being reduced in the < 0.5 μ separate (MacEwan, 1949). According to Jackson *et al.*, (1948) kaolinite is produced by a process of desilication from other members of the series such as: illite, hydrous mica, and montmorillonite. Since kaolinite is more easily perceived at the surface, the hypothesis of Jackson *et al.*, is confirmed. It should be remembered, however, that kaolinite is not only a product of alteration, but also a precipitation product.

Clay minerals in profile #2 show a composition and a distribution similar to profile #1. Vermiculite, however, tends to persist at depth in the profile. The C horizon of profile #2 contains slightly more vermiculite than the corresponding horizon of profile

#1. Furthermore, since in the second profile no subdivision of the clay was attempted, it is possible that once the coarser particles were removed the residue would show even a more marked increase in vermiculite. Illite and kaolinite are virtually the same as in profile #1.

Clay minerals of profile #3 offer in some respect a different composition than those in #1 and #2. A 14 Å mineral identified as chlorite is present in the profile. Its distribution varies according to the horizons, with a general tendency to decrease with depth. Identification of this mineral has been based mainly on heat treatments (Walker, 1951). Similar techniques can also be used to differentiate chlorite from the 14 Å spacing of montmorillonite. According to Bradley, "in the ranges up to 200°C. or 300°C. montmorillonite layers collapse to 9.5 or 10 Å either in monominerals or mixed layer assemblages; chloritic layers deviate in form factor and later collapse in ranges from 300°C. or 400°C. up to as much as 600°C."¹⁵ Since in this mineral the first order reflection persists after heating to 600°C., the name chlorite appears to be justified. As far as the origin of this mineral is concerned, the occurrence of chlorite in the sand grains may be considered as the source of the clay. Some type of correlation, in fact, exists between the occurrence of chlorite grains and presence of chlorite clay. Thus, for example, the A₂ horizon has the highest percentage of chlorite grains (chlorite + chloritoid) and at the same time the highest content of chlorite clay. This correspondence is followed in the C₁ horizon but not in the B and C₂ horizon. Stephen (1952), considering the trend of weathering of biotite in soil, shows that chlorite is first formed at the expense of biotite and subsequently altered to vermiculite. If this sequence holds true, then the presence of chlorite indicates an intermediate stage of weathering. The mechanism of transformation of chlorite to vermiculite takes place with the removal of the Mg ion in the brucite layer under acid condition and a subsequent hydration (Stephen, 1952). If it is assumed also that muscovite during the weathering to vermiculite has chlorite as an intermediate stage, then the soils developed on the Brunswick shale represent chronologically an old stage. In this connection, Stephen states: "The vermiculite product may apparently be reached either through mixed-layer mica-vermiculite intermediate stage . . . or mixed-layer biotite-chlorite and chlorite-vermiculite stages . . ." ¹⁶ The assumption, therefore, appears valid. As far as the illite-mica and kaolinite are concerned, no major differences exist with respect to the corresponding horizons of the other two profiles. Illite in profile #3 tends to

¹⁵ Bradley, W. F., 1953. Analysis of mixed layer clay mineral structures. *Anal. Chem.*, V. 25, p. 729.

¹⁶ Stephen, I., 1952. A study of rock weathering with reference to the soil of the Malvern Hills. *Jour. Soil Sci.*, V. 3, p. 31.

increase with depth except in the C₁ horizon where chlorite again becomes prevalent. The degree of degradation of this illite seems to be less than in the other profiles. Kaolinite is never abundant in any horizon and it stays rather constant throughout; kaolinite is always a "fireclay" type.

Differential Thermal Analysis — The major types of clay minerals identified with this technique correspond in general to those identified with the X-ray method. Gibbsite which had not been indicated by Roentgen analysis has been detected by the thermal method. Certain minerals like vermiculite and chlorite that appear distinct in the diffraction pattern are less evident in thermal analysis.

Thermograms of profile #1 (figure 7) indicate the tendency of illite to increase with depth. The degree of crystallization increases also with depth. Such improvement is relative, however, because even the illite of the C-D horizon cannot be called a well-crystallized clay. Grim and Bradley (1940) have shown that a well-crystallized illite is not destroyed with the expulsion of the OH at 600°C.; in fact, illite retains its structure up to 850°C. This temperature corresponds to the third endothermic reaction which occurs in concomitance with the destruction of the illite structure (Grim and Bradley, 1940). The illite of this profile shows only a slight endothermic peak implying that the crystalline structure is not maintained at this temperature because of a poorly crystallized material. The exothermic peak (900—1000°C.) that follows the destruction of the illite is due to a phase-change; at this temperature illite is changed to spinel (Grim and Bradley, 1940). Vermiculite was identified at the surface; but only traces were present in the B₁ and lower horizons. The first endothermic peak of vermiculite which should occur at about 150 to 200°C. cannot always be identified since illite shows the same reaction in the same region. This peak corresponds to losses of interlayer water the quantity of which depends on the type of exchangeable ion. Water lost at this temperature is also part of the hydration shell around the Mg ions of the brucite sheet (Grim, 1953). A second endothermic peak occurs between 700 and 800°C. and is due to the loss of hydroxyl water (Grim, 1953). An exothermic reaction at about 800°C. is produced by the appearance of enstatite (Walker, 1951). Kaolinite is only present in some samples, although a positive identification cannot be provided since the two peaks can be masked by illite. Grim (1947) reports that mixtures of one-third kaolinite and two-thirds illite has little effect on the final thermal reaction for illite. Considering that the X-ray analysis showed kaolinite only as an accessory mineral, the failure of identifying this mineral with the thermal technique when associated with illite should not be considered very critical. Gibbsite never attains a marked prominence. The exothermic loop present in all thermograms

between 200°C. and 600°C. is ascribed to the presence of organic matter. Supply of air during the heating of the sample allows uniform oxidation process of the organic fraction.

Thermograms of profile #2 (figure 8) show virtually the same characteristics as profile #1. Vermiculite is prominent at the surface, but tends to decrease with depth; illite increases instead with depth as is shown by the more pronounced endothermic reaction at 600°C. The degree of crystallinity of this mineral in this profile is considered superior to profile #1 on the basis of a more marked endothermic peak in the 850—950°C. region. Gibbsite is more prevalent than in the previous profile. Kaolinite is rather difficult to identify because of its small quantity.

In profile #3 (figure 9) the presence of chlorite in the A₂ horizon is well substantiated by two reactions typical for this mineral, namely, an endothermic peak at 750°C. and an exothermic one at 800°C. (Grim, 1953; Bowen, 1954). The first peak is due to the dehydration of the brucite layer, which marks the destruction of the chlorite structure. To this follows an exothermic reaction which marks the appearance of a new phase, olivine (Grim, 1953). According to Brindley and Ali (1950), since olivine appears at about 800°C., this type of chlorite is a penninite. Illite increases with depth together with the degree of crystallization. Gibbsite is more prevalent in the C₂ horizon. Since the endothermic reaction of this mineral coincides with that of goethite (Grim and Rowland, 1942), a further treatment is necessary to differentiate the two. The sample was, therefore, leached with 0.1 N HCl (300 ml. at 60°C.) and heated again to obtain a new thermogram. The area under the peaks before and after the treatment was evaluated and compared with a standard curve. This analysis indicated that gibbsite is present at one percent level.

CHEMICAL ANALYSIS

Total soil fusion analysis was one of the first tools used by students of soils to disclose the nature of the soil-forming processes and to establish the degree of movement of certain compounds. The total fusion analysis provides also an inventory of the potential minerals present in the soil. As far as a pedological interpretation is concerned, however, chemical analysis of soil colloids is more important than the total soil.

An increase of SiO₂ is shown by the A and B horizons in comparison with the C-D (Table 10). This would imply that this compound tends to accumulate in the soil. The presence of quartzose sand deposited on the area obliterates this increase and renders the figures unrealistic.

Iron increases with depth, showing that the source of this element is bedrock material. The A₂ horizon

has the least quantity of iron. This, despite the absence of a distinct horizon of iron accumulation, is probably due to a leaching process. A mixing at the surface with non-ferriferous sand could also produce the same result. Aluminum behaves in a similar way. From the figures shown in table 10, it is clear that considerable aluminum is lost. There is no agreement with what is reported in the literature about the mobility of aluminum in soils. Mattson (1933), however, proved experimentally that aluminum tends to be mobilized before iron in the podzolization process.

Titanium is constant in the profile. Joffe and Pugh (1934) report that in Podzols there is a tendency to accumulate TiO_2 , especially in horizons where the R_2O_3 are concentrated.

The tendency of phosphorus in profile #1 is to increase with depth; such behavior is not generally observed in podzolic soils. A concentration of phosphorus in the A horizon seems to be very common in the Gray Brown Podzolic soils. In fact, it is so common that Baldwin (1927) used it as a criterion to distinguish soils of this group. This usual concentration of phosphorus in the surface of the soil is explained as a result of organic phosphorus in the organic matter of the surface. A second peak is reported also in the B horizon in those cases in which a second accumulation of organic matter is present. Joffe (1940) reports a very small amount of P_2O_5 in the leaching of a Gray Brown Podzolic soil.

With the exception of potassium and calcium, the other bases, magnesium and sodium, show an increase in the B horizon. While this seems to be in harmony with the conditions found in podzolic soils, the amount of sodium appears to be rather anomalous considering the nature of the bedrock. When Sturm (1956) analyzed chemically the shale of the Brunswick formation, he obtained similar results. He indicated that impurities of clay-sized fractions such as analcime ($NaAlSi_3O_8 \cdot H_2O$) could be responsible; but X-ray analyses of many samples rich in sodium did not reveal the presence of analcime. No plagioclase was found in profile #1. Another possibility offered by Sturm (1956) is that some of the illite is brammalite, a Na-illite. The difficulties involved in separating the two renders impossible a definite identification. It is probable, however, that some of the illite is brammalite, a partial confirmation is that there is a concomitant increase of sodium with increase of clay in the profile. Potassium is found to increase steadily with depth. In this connection, a close correspondence exists between the increase of potassium and the increase of illite. Magnesium tends to prevail on alkaline earths; this is probably due to the fact that magnesium is less mobile than calcium. Wiegner and Jenny (1927) studying the mobility of ions, found that migration of calcium was higher than

any other divalent cation. Loddosol (1932) also substantiated these findings. According to Mattson (1933), the difficulty to displace magnesium is due to the presence of stable silicate resulting from the interaction of magnesium with silica in the soil.

The degree of translocation of certain compounds in the soil profile can be better detected by computing molar ratios between the oxides. The most common ones are the SiO_2/R_2O_3 ratio, the SiO_2/Al_2O_3 ratio, the SiO_2/Fe_2O_3 and the K_2O+Na_2O+CaO/Al_2O_3 ratio. Similar ratios can also be obtained from the bedrock and a comparison can be made between these and the ratios of the soil. Such ratios have been particularly stressed by the Russians and Europeans. Polynov (1945) reviews this subject and concludes that the proposed methods to determine the degree of leaching of an element can be used only "when it is possible to compare the chemical composition of the weathering crust in toto with that of the initial mother rock. As to separate soil horizons or individual samples of weathering products, it must be said that the application of these calculation (sic) leads to errors."¹⁷

It has been shown conclusively that many of these ratios collimate very closely with morphological features. Silica sesquioxide ratio of the whole soil of profile #1 shows a marked invariability in the A and B horizons. The same ratio in the C-D horizon and in the Brunswick shale is further reduced. When the SiO_2/R_2O_3 ratio of the soil is compared with that of the bedrock, an accumulation of SiO_2 seems to be evident. However, the same ratio obtained from the colloidal fraction does not reveal such a feature (table 11); very few variations, in fact, are shown among the clays of the soil horizons. An accumulation of silica in the profile is not, therefore, the result of pedogenic processes, but rather due to deposition of quartzose sand. The constancy of the SiO_2/R_2O_3 ratio as well as of the SiO_2/Al_2O_3 ratio is symptomatic of poor profile differentiation implying that chemical weathering has not been very effective on these soils. This is also confirmed by the work of Connor *et al.*, (1957) on trace elements of a soil profile in Hutcheson Memorial Forest, table 18. They concluded that the lithological characteristics of red shale are maintained in the soils despite their great chronologic age. In addition, trace elements distribution in the profile does not show a major differentiation in the horizons. From the chemical standpoint, these soils are not, therefore, affected by podzolization process. As Jenny (1941) has pointed out, the word podzolization means different things to different people. For Marbut, it is mainly a leaching process, whereas for Thorp an accumulation of clay in the subsoil is sufficient to deserve this term. Therefore, soils derived from red shale may be

¹⁷ Polynov, B. B., 1945. Total analyses of soils and their interpretation. *Pedology Acad. Sci. of U.S.S.R.*, No. 1, p. 61 (English Summary).

Table 13. Inorganic elements in soils of Hutcheson Memorial Forest.^a

Horizon	Cu	Mn	Zn	Oven Dry Weight						
				Ni	Co	Sn	Pb	Cr	V	Ga
Norton Soil — (Total)										
A	30	520	86	38	5	12	30	20	55	3
B	29	330	122	77	4	8	101	15	69	13
C	41	390	218	45	5	9	116	33	82	14
Norton Soil — (Clay Fraction)										
A	69	500	400	121	10	80	44	167	193	32
B	65	330	253	103	13	50	75	157	198	32
C	88	170	231	84	12	52	72	164	215	31

^a Taken from Conner et al., 1957.

considered podzolized soils with respect to clay distribution and at the same time non-podzolic from the standpoint of Fe and Al translocation. Russian investigators (1959) have recently coined a new word, illimerization (clay migration) which should be used in those cases in which an undifferentiated clay translocation exists. The word podzolization should be used instead when, due to the destruction of the clay at the surface and a selective remotion of these products, the clay composition varies in the different horizons. It seems, therefore, that for red shale soils, the word illimerization is appropriate.

CATION EXCHANGE CAPACITY AND EXCHANGEABLE CATIONS

Cation exchange capacity of profile #1 (table 12) shows two pronounced maxima, one at the surface and one in the lower horizons (B₂ and C-D). The nature of these two maxima is ascribed to the organic fraction in the A₂ and to a clay accumulation in the B₂ horizon. A rather high cation exchange capacity of the C-D horizon may appear quite abnormal. However, the mineralogical composition of the sand fraction of the two horizons provides a reasonable answer. Sands of the A₂ horizon consist mostly of quartz; while those of the C-D horizon are composed of aggregates, namely, micaceous fragments of red shale. It is obvious that the exchange capacity of the quartz particles is nil compared with the clay impregnated particles of the red shale. While at the surface, exchange sites are available only in the finer fractions, in the C-D horizon the coarse particles play a similar role. This acquires particular importance for the vegetation of the woods inasmuch as there is no defined zone in the profile from which the nutrients are available, for the entire soil represents the source of mineral nutrition.

The exchangeable cations of profile #1 offer a usual pattern with respect to their distribution. In podzolic soils, the adsorbed cations appear generally in the following order: Ca > K > Na. In this soil the same order is preserved with the exception of Na which approximates K. The possibility that some of the illite is brammillite, that is, a Na-illite, could help in explaining the correspondence of the two cations. The degree of release of the two ions

would then be a function of the clay breakdown.

ORGANIC MATTER AND C:N RATIO

The layer of humus which covers the forest attains different thickness and structure according to the frequency of the plants, texture of the soil, drainage and related conditions. This layer in profiles #1 and #2 has been identified as a medium mull (Heiberg and Chandler, 1941). The corresponding layer in profile #3 has been named a fine mull. Quantity of soil organic matter varies in the three profiles. Profile #2 (table 14) is the most rich in organic matter, profile #1 (table 12) moderately rich, and profile #3 (table 15) the poorest. In all three profiles as a general rule the organic matter tends to decrease with depth. Total N decreases in quantity passing from the A to the B horizon of all three profiles; but it increases again below B. Carbon:nitrogen ratio in the first profile is rather narrow at the surface indicating a type of organic residue at a well-advanced stage of decomposition. This ratio becomes wide in B₁ and narrow again in the B₂ horizon. The increased ratio in B₁ is possibly the result of undecomposed organic matter provided by the roots of trees. Reduction of microbial activity with depth could be responsible for the persistence of this organic matter. A relative increase of N in the B₂ tends to lower the ratio in this horizon. Reports on the increasing quantity of N with depth in the soil appear rather common in the literature on this subject. McBeth (1917) showed that NH₄OH can be fixed by the clay in an irreversible way. More recently, Allison et al., (1953) indicated that NH₄OH fixation occurs preferentially in the subsoil. The same authors found that illite and vermiculite upon drying and heating can fix ammonia. The fixed ammonia becomes less available to nitrifying bacteria, especially when the fixing clay is a vermiculite. Stevenson (1957) found that N-NH₄OH increases with depth in the soil. The carbon:nitrogen ratio of profile #2 offers a similar pattern although the values for such ratio are further reduced. Anderson and Byers (1934) studying C:N in relation to soil classification found that although wide differences exist in the Gray Brown Podzolic soils, the general trend was a decrease of the ratio with depth. Profile #3 registers the least amount of soil organic matter. The sandy nature

of the soil is directly and indirectly responsible for this condition. An open texture favors a good aeration of the soil, with a consequent rapid oxidation of the organic debris. Indirectly, the relatively low level of nutrients in sandy soils does repress any luxuriant vegetation and, therefore, the production of organic matter is reduced. Carbon:nitrogen ratio of this profile shows a decrease with depth; while a wide ratio occurs at the surface, a lower one appears in the C₁ horizon. The same considerations cited for the two previous profiles can be applied also in this case. It can be added that the initial C:N ratio of the aerial part of shrubby plants is higher than that of the roots. Differences in composition of plant material may play a role in narrowing the ratio.

pH

The annual pH variations observed in the soil seem to follow a definite pattern, table 13 and figure 11. The highest pH values were recorded in winter and summer, whereas the lowest during spring and fall. Hammerling (1927) mentions an investigation near Moscow dealing with the reaction of a podzolized soil where it was found that pH values were related to soil moisture content. High pH values were obtained at low moisture levels and vice versa. Climatological data available from Blackwell Mills Station, New Jersey, show, in general, that the kind of relationship suggested by the Russian investigators between rainfall and reaction of the soil does not always exist. In fact, in those months with the highest pH, the highest precipitation always occurred. Even if the winter months are grouped together, they always show the highest precipitation of the year. A closer relation between soil moisture and pH may exist during the summer. Lutz and Chandler (1946) report that drying the soil lowers the pH. The variations are always less than 0.1 or 0.2 pH unit. Nehring, working on German soils, (Lutz and Chandler, 1946) found the highest pH values in the winter and the lowest in the summer. These variations did not exceed 0.8 pH unit. Joffe (1949) reported that pH values in forest humus layers were highest in autumn due to release of bases from the fallen leaves. Coile (1937) pointed out that the tree species that contain high Ca in the leaves produce high pH values. Chandler (1939) also found that the litter of some species can affect the pH and base saturation of the soil. He determined the Ca content of mature foliage of 27 forest-tree species and related this content to soil. The group of trees that includes oaks, birch, hickory, and elm seems to have a neutral effect on the soil. According to Chandler (1939), an oak hickory community will not affect the status of Ca in the soil. Since Hutcheson Memorial Forest confines itself to this category, it can be implied that the vegetation will not affect di-

rectly the soil pH values. From the values reported (table 13 and figure 11), it can be observed that the degree of decomposition of the forest litter influences the pH of this layer. The A₁, which is made up mostly of undecomposed leaves, has a lower pH value than the A₂ horizon in which the advanced decay had time to release the bases present in the tissues. The general trend of the pH in the profile can be related to microbial activity in the soil. This activity starts in spring, and after a lag during the dry summer months, starts up again in autumn. Intense biological activity of spring causes the production of organic acids as the result of the decomposition process, together with emission of CO₂ from respiration. Should the organic materials have a wide C:N ratio, ammonia from the soil may be used, thus freeing anions. The same processes repeat themselves in the autumn. The lowest pH readings in the forest are recorded during spring and autumn, which seems to indicate a concomitance between the microbial activity and the soil reaction.

SOIL GENESIS

Genesis of the soils in Hutcheson Memorial Forest has been to a great extent monopolized by the underlying bedrock. The persistence of the red color in the soil, the persistence of the Na ion in the complex, and the inherited clay are all indications of the influence of the Brunswick shale on soil development. Krynine (1949) suggested that the red beds were derived from highly weathered soils formed under a savanna type of climate. This, however, appears somewhat incongruent because shale reflects only partially a highly weathered status. In fact, its composition if on the one hand, reveals the presence of an hematitic pigment, synonymous strong weathering on the other hand, it shows the existence of an illite mineral which is anything but a last stage of alteration. The association, therefore, of these two minerals, far apart in the weathering sequence, in the same sedimentary rock, cannot be very well accepted. Despite all this, however, the coexistence of hematite and illite can be still justified if a process of diagenesis is considered between the phase of deposition and that of lithification. As has been pointed out by Grim (1951), Ross (1943) and Bradley (1953) clay minerals in degraded state (illite, chlorite and possibly kaolinite) when sedimented in alkaline waters "would quickly absorb any available K⁺ or Mg⁺⁺, with attendant increase in the perfection of the mica structure. It seems likely that such reconstitution of the mica is a most important diagenetic process."¹⁸ If a depositional environment with alkaline waters is postulated for the Triassic sediments, then the reconstruction of degraded clays

¹⁸ Grim, R. E., 1951. The depositional environment of red and green shales. *Jour. Sed. Petr.*, V. 21, p. 226-232.

to illite can be explained. Although the Newark rocks are considered terrestrial sediments, it is possible that in the shallow lakes and lagoons during the dry season an alkaline medium was created. The existence of fossils of fishes that thrive in brackish waters might substantiate this hypothesis. These alkaline conditions would help in turn to explain the high Na content of the Brunswick shale. The selective absorption of K^+ and Mg by the degraded minerals to rebuild or build the crystalline structure would leave an excess of Na, which could enter into the structure of illite competing thus with K. The hematite on the other hand, being the last stage of a weathering cycle, could not progress any further. A reversal in the weathering sequence of this mineral should involve its chemical reduction (Jackson *et al.*, 1948) and since this has not occurred, hematite has remained stable. The stability of this mineral is further shown in the persistence of Fe in the profiles. Trials to leach this element from the C-D horizon of profile #1 with HCl solutions at pH 2.5 and 2.9 have given virtually no results. The extractable Fe at pH 2.5 was 0.019 percent of the total and at pH 2.9, only 0.006 percent. Both pH values are very unrealistic for this soil. The insolubility of the Fe is accomplished by a similar insolubility of Al. In fact, the SiO_2/Al_2O_3 ratio remains constant in the profile. However, while the lack of mobility of the Fe may be attributed to its hematitic form (Sturm, 1956; Kry-nine, 1950) for the immobility shown by Al, there is no satisfactory explanation. Reifennberg and Ewbank (1933) studying the red soils of the island of Cyprus theorized about the stability of the sesquioxides found in these soils. According to these authors, the sesquioxides, peptized by colloidal silica, would be coagulated by the rising electrolytes during the dry season and made irreversibly by the summer heat. This process seems to occur preferentially on calcareous rock but it is also found to take place on igneous rocks.

Although there is no intention to postulate such a process for the soils of the Newark epoch, it is opportune to remember that also in the realm of soils, the criterion used by the geologists that "the present is the key of the past" could sometime be used. As far as the acidity of these soils is concerned, it has been shown that the soil pH values are in the 4.5--5.5 range most of the year. While under these acid to subacid conditions, the sesquioxides are not affected, the acid weathering of the clay becomes rather effective. The main type of clay which has been identified as a dioctahedral illite has weathered to a vermiculite mineral. It is assumed that the weathering has been active for a long time. To support this view, it may be added that in the second profile, vermiculite could have reached low horizon through illuviation. However, if this is so, the horizons of

clay accumulation should show a prevalence of vermiculite which is not the case. The distribution of this mineral indicates a concentration at the surface and a decrease with depth which is in agreement with a weathering process.

Profile #3, derived from sandy material, shows a different pattern of weathering. The chlorite mineral found here is either a product of mechanical abrasion of the coarse chloritic particles or a product of weathering of illite. Probably, it is the result of both. As a product of the weathering of illite, chlorite indicates an intermediate stage of alteration in the general sequence illite \rightarrow vermiculite. This stage is considered as an index of youth as compared with the red shale soils.

In the systematic study of soils, red shale soils have been considered in the Gray Brown Podzolic, a group introduced by Baldwin (Baldwin, 1927). This grouping is, in some respects, justified inasmuch as the features that the climate imposed on these soils are podzolic in nature. However, according to Fridland (1959), the mere process of clay translocation is not enough to justify the term podzolization. If this is the case, then the soils of Hutcheson Memorial Forest would not be included with Gray Brown Podzolic and their placement into a specific great soil group remains unresolved.

In this connection, it may be pointed out that the original Baldwin concept of Gray Brown Podzolic soils wanted to include a specific type of soil preferentially formed on calcareous rocks. This concept, however, became more indistinct and more inclusive; virtually all the soils of the eastern United States between Prairies, Podzols and Red-Yellow Podzolics were included with the Gray Brown Podzolic soils. Actually, we must consider that this group was conceived by Glinka (Baldwin, 1927) who never accepted the existence of an independent soil type for the mid-latitude zone. In fact, the Brown Earths described by Ramann were considered by Glinka as "the weathering layer B of the Bleached Earths..."¹⁹ This implies that between the Red-Yellow Podzolics of the south and the Podzols of the north there is no buffering zone; but one group encroaches upon the other producing a transitional zone. The soil of Hutcheson Memorial Forest due to its unique characteristics cannot be included, orthodoxically speaking, in any great soil group. On the other hand, it would be more correct to consider these soils as a type of Brown Earth under podzolic environment. This would be also in agreement with what Ramann states about this group, that the parent material plays a large role in its development.

¹⁹ Ramann, E., 1928. The evolution and classification of soils. Translated by C. L. Whites, Cambridge, Heffer, p. 87.

SUMMARY

The soils present in Hutcheson Memorial Forest have developed from the Triassic red shale and from a fluvial sand deposit. The red shale soils are not, however, considered residuum soils because allogenic minerals uncommon to the area were detected in the solum. Many of the features of the sandy soils are connected to depositional factors rather than to pedogenetic ones. Soils derived from shale, despite being exposed for a long time to soil-forming processes, have not acquired an independent individuality.

The persistence of the red color in the solum, the weak horizons differentiation, the lack of translocation of certain compounds are all characters of an immature soil. Despite this, it seems that certain features gained by the soil with respect to the bedrock cannot be underestimated. The weathering of illite to vermiculite, the presence of an illuvial B horizon, the quantity of iron coatings in

the same horizon indicate that the pedogenic processes, even if weak, had operated on the soil profile. Most pedologists held the opinion that the bedrock of the Penn soils represents an end product of a previous weathering cycle and it appears as an exhausted substratum unable to react to the soil forming factors. This point of view can be shared only with reserve inasmuch as the clays and the chemical composition do not indicate such status. The 'quality' of the soil components rather than their 'quantity' had played a role in the genesis of these soils. This concerns mainly the low solubility of the sesquioxides which probably have been irreversibly precipitated during past geological eras. The resistivity of the red shale to the factors of pedogenesis is nothing else than the resistivity of the sesquioxides to solubilization and translocation. While the inclusion of these soils into a great soil group is not yet a clear-cut issue, it is suggested that Ramann's braunerde is the most appropriate.

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APPENDIX

*Moisture Determination of Profile #1 (Forest)
at Different Dates (1957)*

Depth	July 30	Aug. 4	Aug. 17	Aug. 31	Sept. 14	Sept. 27
ins.	pct.	pct.	pct.	pct.	pct.	pct.
A ₁	21.9	53.7	30.3	26.0	8.3	75.0
0-3	9.7	28.2	12.5	11.9	16.6	11.5
3-6	8.1	8.2	11.5	11.6	15.6	12.0
6-9	7.1	7.8	9.3	10.2	14.2	10.2
9-12	6.0	7.8	8.6	9.1	11.2	10.0
12-15	6.1	8.1	8.0	8.1	8.6	10.8
15-18	8.5	9.1	10.1	8.6	11.6	13.5
18-21	8.9	12.1	10.8	8.4	12.5	14.1
21-24	9.6	12.3	11.3	8.7	11.6	17.8
24-27	9.2	10.8	10.2	12.3	10.3	10.6
27-30	8.1	12.9	9.3	9.8	9.2	5.6
30-33	8.5	13.7	8.3	7.4	9.2	8.0
33-36	9.0	14.5	8.4	7.0	8.1	10.0

*Moisture Determination in the Cultivated Field
at Different Dates (1957)*

Depth	July 30	Aug. 4	Aug. 17	Aug. 31	Sept. 14	Sept. 27
ins.	pct.	pct.	pct.	pct.	pct.	pct.
0-6"	6.5	4.1	57.1	7.2	13.0	11.5
6-12"	6.7	7.0	58.9	8.2	8.0	7.4
12-18"	8.6	8.2	7.1	8.2	8.6	9.7
18-24"	8.5	11.7	7.3	9.3	9.0	9.2