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SOUTHEASTERN GEOLOGY

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GEOLOGICAL RESEARCH IN THE SOUTHEAST - 1970 EDITION

By

James W. Clarke
McLean, Virginia

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A_2 HORIZONS OF COASTAL PLAIN SOILS PEDOGENIC OR GEOLOGIC ORIGIN[1/]

By

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ABSTRACT

Many lines of evidence indicate that the A2 horizon of soils in the North Carolina Coastal Plain is of pedogenic origin. Those demonstrating a pedogenic origin are the uniformity of sand sizes across the A2 — B horizon contact, the micro-intertongued A2 — B horizon contact, near absence of oriented clay in the A2 and its abundance in the subjacent B2, eluvial-illuvial nature of the clay bands in the A2 horizon, the thickening and thinning and changes in color and clay content of the A2 in response to changes in soil drainage, and the presence of A2 horizons in soils on eolian sands. Any one of these would not be enough to negate the hypothesis that the A2 horizons are a separate geologic deposit, but when considered as a group they are very strong evidence for a pedogenic origin for the A2 horizons or surface sands.

INTRODUCTION

Virtually all of the North Carolina Coastal Plain is mantled by

1/ Paper number 2989 of the Journal Series. Joint contribution from the Soil Conservation Service, USDA, and the Department of Soil Science, North Carolina Agricultural Experiment Station, Raleigh, North Carolina.

a sandy cover of varying thickness. This sand[2] is most prominent in the Sandhills and in the upper Coastal Plain and southern parts of the middle and lower Coastal Plain. Surface textures in the northern parts of the middle and lower Coastal Plain are finer, ranging from sandy loams to silt loams and the A horizons are not as thick. In the Sandhills, surface horizons are three to five feet thick and light gray to white sand to loamy sand. They cover both the interstream divides and the side slopes, and occur on several different geologic materials. The A2 horizons (surficial sands)[3] are indistinct and only 1/2 to one foot thick in many areas of the middle and lower Coastal Plain, although locally (in the southern parts) they may approach the thickness found in the Sandhills. In wetter areas of the Coastal Plain the A2 horizons ordinarily are less than a foot thick and are indistinct because the upper part has been darkened by organic matter. Exceptions are the white sands in the very sandy wet areas of the southeast Coastal Plain.

Some geologists have considered the A2 horizons as a separate deposit (Pirkle et al., 1964; Howard, 1955; Clark, 1912; Conley, 1962) whereas other geologists have thought of them as a weathering phenomenon associated with soil formation (Altschuler and Young, 1960; Hope, 1956). Soil scientists also have had mixed feelings. Before the introduction of a new soil classification system (7th approximation, Soil Survey Staff, 1960) forced people to look closely at their concepts of soil, it was accepted practice to consider most sandy surface deposits as A2 horizons only if they were less than 30 inches thick. This meant that the A2 and B2 horizons were believed to have formed by weathering from one supposedly uniform deposit. But if the sand was much more than 30 inches thick it was in many instances interpreted as a C horizon or a sandy overlay, and the underlying more clayey material was interpreted as either a buried soil or material from another geologic formation. These concepts frequently resulted in a small level field having a soil with an A-B sequence (no overlay) intermixed with a soil with an A-C sequence (a sand overlay). No difference in relief, no change in materials, nor any other features were evident. The only difference between these soils was that one had less than 30 inches of sandy surface and the other had more than 30 inches. This illogical

2/ In this paper sand refers to material dominated by sand-sized fractions.

3/ The A2 horizons of the pedologist are commonly called surficial sands by geologists. Because this sandy layer is considered part of a soil profile, we shall call this layer an A2 horizon throughout this paper. This will help prevent confusion with the term "surficial" which can mean anything from a thin surface sediment to a catch-all term used by Heron (1958) to include anything overlying the basal Cretaceous. We imply no genesis in the use of the term A2 horizon. The genesis of this layer is developed in the paper.

situation resulted from the concepts and theories then held. Based on recent studies (Rivers et al., 1963; Gamble, 1966; Daniels et al., 1967; Daniels and Gamble, 1967) and changing concepts we can now see that we must choose one of two hypotheses. Either both soils in the small area have genetic A2 horizons, or both have a cover sand (different deposit) of differing thickness.

These contrasting hypotheses on the origin of the sandy A2 horizon have many implications as to the origin of surficial deposits and soils in the Coastal Plain. If we assume that the sandy cover is a distinct stratigraphic unit then we want to know how it was deposited, when deposition occurred and the location of the source area. If the sandy cover is a weathering phenomenon then we want to know what processes are involved and how varying conditions of soil drainage and parent materials modify these processes.

A2 HORIZON CHARACTERISTICS

The A2 horizons of soils in the Sandhills and upper Coastal Plain usually have less than 10 percent clay (Figure 1) but a wide range in silt content is found in samples from the middle Coastal Plain. There is a considerable cluster of samples with 85 to 90 percent sand, 10 percent silt, and less than 5 percent clay. Samples from the upper Coastal Plain are clustered near the 90 percent sand 6 percent silt area of the textural triangle whereas those from the middle Coastal Plain range widely. It is evident that the A2 horizons do not have the same texture everywhere.

The contact between the A2 horizons and the underlying material is smooth but has many macro and micro features that change with soil drainage. Well-drained soils with yellowish red B horizons have considerable contrast in color and texture between the sand and the B horizon. The A2 horizons are light gray to very pale brown (10 YR 7/2 to 7/4) loamy sand[4/]. The underlying B horizon is yellowish red (5 YR 4/6) sandy clay loam to clay. This change in color and texture can occur across a boundary layer that is less than 1 inch thick, but which may be 2 or 3 inches thick. In most areas the boundary is relatively smooth but there is much variety in detail across the contact. A contact typical of well-drained soils is shown in Figure 2. Tongues of the A and B horizons 1/2 inch wide and 1 1/2 inches long extend into each other. On a vertical face many bodies of the B horizon appear to be suspended above the main contact (see Figure 2), but these multi-shaped bodies are actually connected to the B horizon when seen in three dimensions. These bodies are not like the clay galls commonly seen above a sand-over-clay contact in sediments. The sand grains in the A2 horizons are relatively clean with little or no coating material, and

4/ Texture based on the USDA system (Soil Survey Staff, 1951).

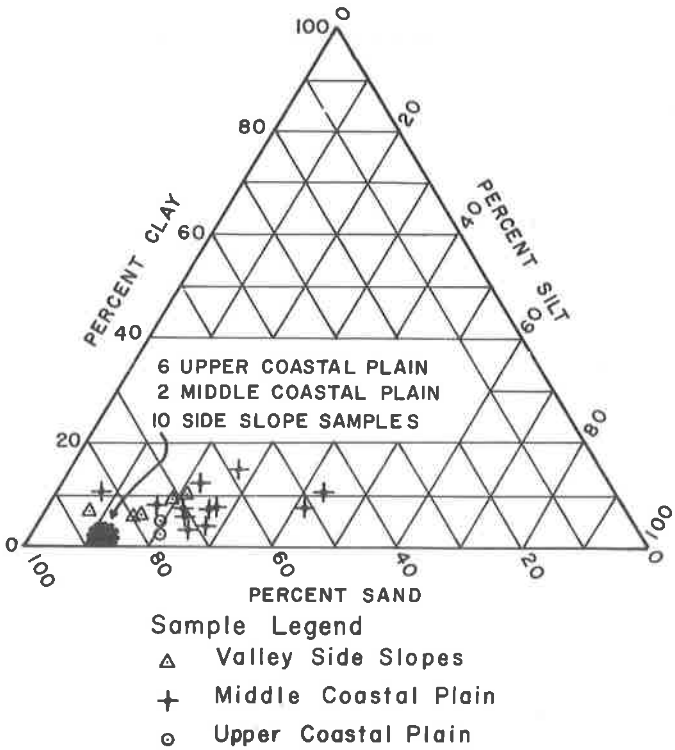

Figure 1. Sand, silt, and clay content of A2 horizon samples.

under a hand lens this condition grades across a boundary 2 to 3 mm wide to the B horizon where the sand grains are "floating" in a clay matrix. This same sort of "micro" boundary is found between the sand and the multi-shaped bodies of the B horizon surrounded by the sand.

In moderately well-drained soils the contact between the A2 and the B horizon is indistinct and has little color contrast. The intertongued effect found in the well-drained soils is absent or indistinct. The contact is a transitional layer an inch or two thick that has an abrupt (i. e., less than one inch) textural change from a sandy loam to sandy clay loam. This textural change must be felt with the fingers; it usually is not visible. Sand grains in the A2 horizons are coated with yellowish brown material, as are those in the underlying B horizon.

The A2 horizons in poorly drained soils are similar to those in moderately well-drained soils. Color contrast between the A2 and B horizon is low because both are gray or light brownish gray. The sand grains of the A2 horizons have a dirty appearance and about 1/2 the grains are coated with a coarse matrix material. The increase in clay from the A2 to the B horizon is readily apparent in the field and the contact is abrupt to clear.

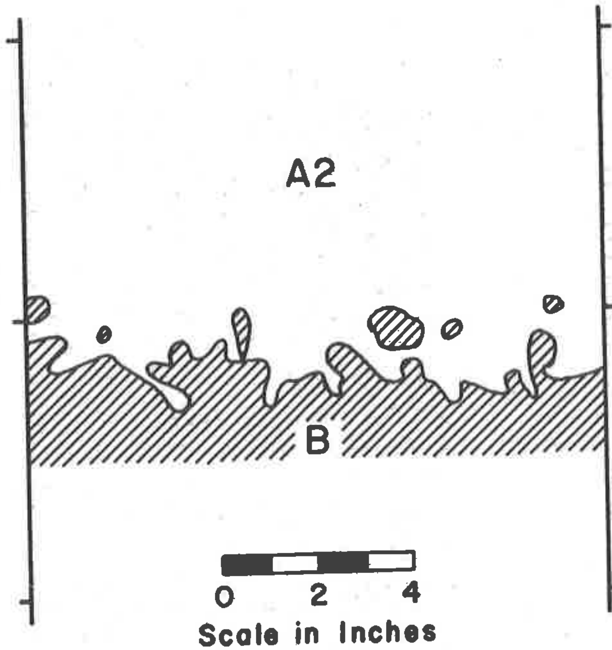

Figure 2. Typical intertongued A2-B horizon contact.

RESULTS OF FIELD AND LABORATORY STUDIES

A combined field and laboratory study was conducted in two phases to determine the size relation between the sand grains in the A2 horizons and those in the underlying B horizons. In the first phase eight sites from gently convex interfluves were sampled in detail from the A2 horizon across the contact into the underlying B horizon. The sample sites were on upper and middle Coastal Plain surfaces in Johnston and Wake Counties, North Carolina. Sites were at least 2 miles apart. A typical summation curve based on clay-free samples is shown in Figure 3. This cumulative curve is typical for 7 of the 8 sites. The eighth site (sample 31-6 Table 1) had an abrupt change in sand size at the contact between the A2 and the underlying B horizon. The B horizon was in the Macks Formation (Daniels et al., 1966) and the A2 Horizon in material eroded from the Pinehurst Formation and redeposited down slope on the Macks. Coarse sand (2-1 mm) content of the samples from these eight sites ranged from 9 to 26 percent and was randomly distributed on the landscape. The range in mean size of the sand and silt fraction of these samples is shown in Table 1. From this study it was possible to conclude that there was no regional pattern to the distribution of sand sizes, and the sand sizes in the A2 horizons and the underlying B horizon were in most cases similar.

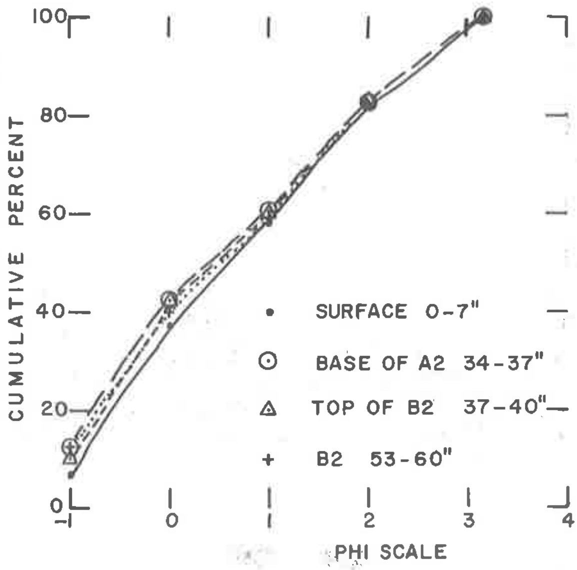

Figure 3. Typical cumulative curves for sands from A2 horizons and the underlying B horizon. Data from Soil Survey Laboratory, Beltsville, Md.

Table 1. Mean Sizes of Silt and Sands in mm.

	Horizon				
Sample #	Ap	A2	B21	B22	B23
31-1	.27	.27	.31	.21	.21
31-2	.31	.29-.26	.24	--	--
31-3	.22	.23-.20	.18	.26	.30
31-4	.25	.23-.30	.26	.27	--
31-5	.15	.15-.16	.15	.15	.15
*31-6	.14	.13-.12	.09	.07	.06
40-1	.20	.25-.26	.25	.27	.28
40-2	.24	.22-.29	.25	.30	--

*Discontinuity between A2 and B horizons.

The second phase of the study was a statistical design to look closely at the variation in sand sizes between the A2 and the underlying B horizon on a local landscape unit (Gamble, 1966). The local landscape units consisted of a flat to gently convex interfluve and adjacent side slopes. Sample sites were in Johnston and Wake Counties on upper and middle Coastal Plain surfaces. The side slopes are Holocene surfaces (Daniels, et al., 1966).

Traverses consisting of five paired sites extending across each landscape position were used. Within each position there were ten individual sites in two parallel lines. This sampling scheme was used in seven areas. The paired arrangement was made to obtain an estimate of horizontal variation or error. The five pairs were about 40 to 60 feet apart and spaced more or less equidistant across each position. Sampling was done from small pits or by hand auger. The lower six inches of the A2 and the upper six inches of the B horizon were sampled. The presence or absence of discontinuities was evaluated in the field using criteria such as detectable changes in sand size or the presence or absence of stone lines or gravel layers.

Laboratory evaluations of discontinuities were made by comparing the cumulated weights of the sand fractions. Silt and clay size material were not used because a soil with a genetically related eluvial A2 and B2 horizon formed in homogenous material would have an abnormal accumulation of fine material. The A2 horizon, having lost fine material, would have an abnormal amount of coarse material. The sand fraction, being mainly quartz, is much less subject to particle size changes produced by soil development.

The phi means of the sand fraction from the surficial sands were compared to the phi means from the underlying B horizons of the interfluve summits and the side slopes. The differences between the means of samples on the interfluve summits from four areas are all less than the least significant difference (lsd) and they are less than the possible laboratory error for phi mean (Table 2). Thus there is no difference in sand sizes between the A2 and the B horizons on the interfluves. Similar results were obtained in the other three areas not reported in Table 2.

Sand sizes generally are finer in the A2 horizons than in the B horizons on the side slopes (Table 2), although this is not true at all sites. Finer sands would be expected on many side slopes because the upper and middle Coastal Plain stratigraphic units are coarse at the base and become finer toward the top. Any movement of sand toward a lower elevation on the slope would bring fine sand over coarser sand (Figure 4). The discontinuity in sand sizes on these side slopes where it does occur, does not always coincide with the A2-B horizon contact but may occur anywhere (Table 3).

AREAL VARIATION OF A2 HORIZONS

Detailed field work has established that the thickness of the A2 horizon and its clay content on a gently undulating surface varies systematically with soil drainage (Daniels et al., 1967). In similar materials the A2 is always thickest in well-drained soils and about equal in thickness in moderately well and poorly drained soils (Table 4). Within

Table 2. Phi Mean Sizes of Sands from Surficial Sands (A) and B Horizons: Areas x Horizons x pairs within Position Means.

Least significant difference = 0.2323 at 5% level
Experimental error = 0.0880

Interfluve Summits

	Horizon	Pairs 1	2	3	4	5
Area 1	A	1.0202	1.8987	1.9303	1.9695	2.0194
(Upper Coastal Plain)	B	1.1049	1.9433	1.9567	1.9196	1.9544
Difference		0.0847	0.0546	0.0264	0.0499	0.0650
Area 2	A	2.0904	2.0436	2.0143	1.9761	2.0242
(Upper Coastal Plain)	B	2.0697	2.0656	2.0218	1.9205	2.0148
Difference		0.0207	0.0220	0.0075	0.0556	0.0094
Area 3	A	2.1091	2.2361	2.3862	2.3721	2.5179
(Middle Coastal Plain)	B	2.0554	2.1350	2.2688	2.3124	2.4818
Difference		0.0537	0.1011[a]	0.1174[a]	0.0597	0.0361
Area 4	A	1.9184	1.9411	1.9047	2.1245	2.1486
(Middle Coastal Plain)	B	1.8548	1.8573	1.8259	2.0929	2.1028
Difference		0.0636	0.0838	0.0788	0.0316	0.0458

Side Slopes (Holocene Surfaces) Associated with Interfluve Summits

	Horizon	Pairs 1	2	3	4	5
Area 1	A	1.4654	1.5073	1.5198	1.5805	1.6787
	B	1.0083	1.3798	0.9906	1.2935	1.2456
Difference		0.4571[b]	0.1275	0.5292[b]	0.2870[b]	0.4331[b]
Area 2	A	1.5523	1.2559	1.6650	1.6282	2.0546
	B	1.3875	1.1393	0.9558	1.2200	1.7604
Difference		0.1648	0.1166	0.7092[b]	0.4082[b]	0.2942[b]
Area 3	A	2.1050	2.1347	2.1985	2.2192	2.2250
	B	1.7842	2.0580	2.0846	2.1184	1.9871
Difference		0.3208[b]	0.0767	0.1139	0.1008	0.2379[b]
Area 4	A	1.9141	2.0707	2.0265	2.1273	1.9589
	B	1.9502	2.0575	2.0141	2.1134	1.7903
Difference		0.0361	0.0132	0.0124	0.0139	0.1686

[a] differences greater than experimental error
[b] differences greater than least significant difference

the well-drained soils those with yellowish-red B horizons or yellowish-brown B horizons near the dissected edge of the surface have the thickest A2 horizons. Within well-drained soils the clay content of the A2 decreases from the wet edge (adjacent to poorly drained soils) to the

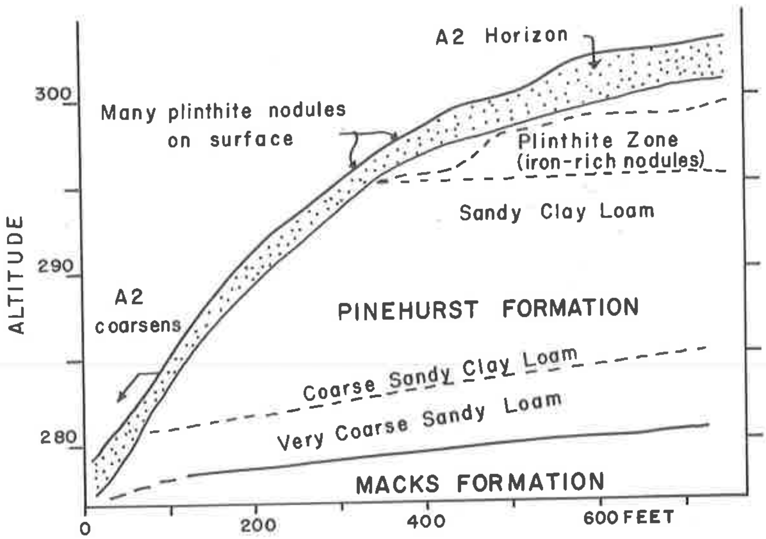

Figure 4. Relation of A2 horizons to sediments on side slopes.

dry edge, or dissected edge, of the surface (Table 5). This decrease in clay in the A2 horizon across well-drained soils is accompanied by an increase in clay in the upper part of the B horizon (Daniels and Gamble, 1967) (Figure 5). It is across surfaces such as these that we find no change in mean size from the A2 to the underlying B horizon.

A2 horizons are found over very weakly developed B horizons of soils formed in eolian sands. These eolian sands generally have less than five percent clay (Daniels et al., 1969), yet the sandy A2 horizon so characteristic of well-drained medium to coarse textured soils throughout the Coastal Plain is present.

THIN SECTION STUDIES

Thin section studies of the A and B horizons were used to extend observations to a greater level of magnification than was possible with a hand lens. Sections from the A2 horizons generally have sand grains with minor amounts of unoriented silt. Rare bodies of oriented clay with ragged edges are found in the A2 horizons. Oriented clay, usually interpreted as evidence of clay movement and redeposition (Thorpe et al., 1959) is common immediately under the intertongued A2-B horizon contact on four profiles studied in detail. Bodies of clay-size material occur discontinuously in the A2 as oriented and unoriented bodies. These bodies of clay grade horizontally and vertically into bodies with more coarse silt. In the B horizon fine particles are the major constituent. Bands of yellowish brown material about 3 mm thick

Table 3. Phi Mean Sizes of Surface Sands and Underlying B Horizons on a Valley Side and Relation to Laboratory and Field Recorded Discontinuities.

Pair	Site #	Phi Mean A B	Diff.	Field comments on discontinuity
1	11	2.1044		
		1.5414	0.5630*	Definite discontinuity at boundary
	12	2.1056		
		2.0269	0.0787**	Discontinuity below contact
2	13	2.0291		
		1.0785	0.0506**	Discontinuity below contact
	14	2.2404		
		2.1376	0.1028	Possible discontinuity at contact
3	15	2.1342		
		2.0376	0.0966	
	16	2.2628		
		2.1317	0.1411	Possible discontinuity at contact
4	17	2.1918		
		2.0629	0.1289	
	18	2.2467		
		2.1739	0.0728**	Discontinuity below contact
5	19	2.1464		
		1.9377	0.2087	Discontinuity above contact and surface sand sample
	20	2.3037		
		2.0366	0.2671*	Discontinuity likely at contact

* Significant difference in sand size between A & B horizon.
**Difference between samples less than laboratory error.
0.2323 = least significant difference.
0.0880 = experimental error.

are common in the thicker A2 horizons. Thin section studies indicate these bands are composed of sand grains and many bodies of oriented clay. The bands appear to be literally stuffed with clay. The upper band boundary is smooth and abrupt and the lower boundary is irregular. Very little fine material is found above the band but there are rare clay bodies with irregular edges just above the band. The amount of fine material increases below the band. This suggests that the bands might result from the "cleaning out" of small amounts of fine material in the A2, or perhaps from downward movement of fines added as local or regional "dusts" after the surficial sands have developed.

Table 4. Mean Values of A_2 Horizon Thickness by Drainage Class and Geomorphic Surface.

Geomorphic Surface & Location	Horizon Thickness (in.) Well drained: Yellowish red B	Well drained: Yellowish brown B	Moderately well drained	Poorly drained	LSD[a] 5%	LSD[a] 1%
Brandywine		16.40[b]				
		12.53	8.95	8.20	2.27	3.02
Coharie						
Newton Grove	13.42	9.78	7.92	6.80	1.39	1.83
Mount Olive		11.80	9.30	11.43	1.99	2.64
Sunderland		11.27	8.03	8.60	1.99	2.64

[a]L. S. D. = least significant difference
[b]Dry edge of surface

Table 5. Clay in A2 Horizons in Well-drained Soils

Area	% Clay for Different Landscape Positions: Wet Edge 1	Intermediate 2	Intermediate 3	Dry edge 4
Mount Olive	9.8	8.2	6.2	3.2
Benson	6.7	6.5	3.3	4.2

INTERPRETATION

The two hypotheses advanced at the start of this study were that the A2 horizons were either a separate geologic deposit or a weathering phenomenon related to soil formation. The strongest argument for a geologic origin is that the A2 horizon sand drapes across all parts of the landscape (Figure 4), but soil forming processes could result in the same feature. Assuming that the A2 is a separate deposit, deposition in or by water seems improbable because the A2 is a continuous mantle on valley slopes, interfluves and low terraces. The A2 could have been deposited by wind, but if an eolian mechanism is invoked, how does this fit with the other characteristics of the A2 horizon? The wide variety in texture of the A2 (Figure 1) could be the result of different source areas, and this could also explain the wide range in coarse sand found in samples studied initially. This almost restricts the sands to a very local source. The uniformity of sands across the

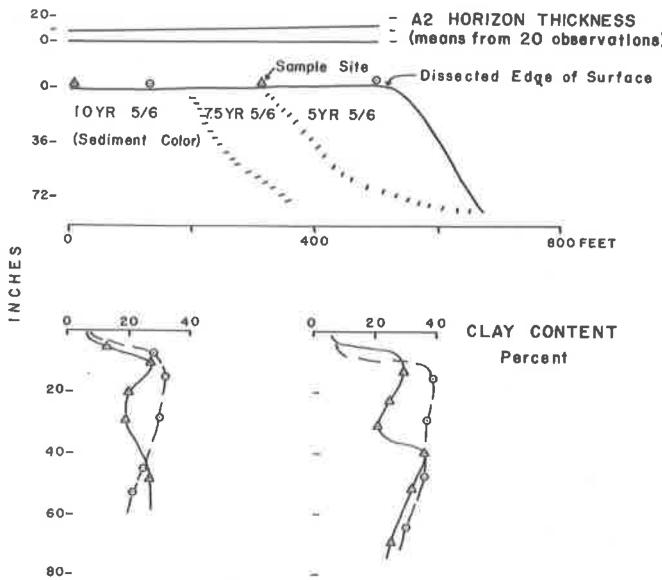

Figure 5. Relation of clay content of B horizon and the thickness of A2 horizon across the dissected edge of a surface. The A2 horizon thickens toward the dissected edge of a surface.

contact between the A2 and B horizons on gently sloping to level interfluve summits (Table 2) is the strongest evidence against the eolian hypothesis. Even on side slopes where erosion would have a chance to move finer sands over coarser sands (Figure 4, Tables 2, 3), only about 1/2 the sites have a significant change in sand sizes across the contact. On these side slopes, the discontinuity as seen in the field could be at, above, or below the A2-B horizon contact. The fact that sand sizes are uniform across this contact on interfluve summits at one or two sites could easily be a chance relation. But since ten sites in seven areas, a total of 70, on interfluves show no difference in sand across the contact, a stratigraphic discontinuity at the base of the A2 horizon is extremely unlikely. This argues for uniform material. Additional data arguing against the eolian hypothesis are the close relation between thickness of the surface sand and soil drainage (Table 4) and the change in clay content of the sand from the wet to the dry edge of a geomorphic surface (Table 5). A2 horizons also are found in local eolian sand bodies on the Coastal Plain (Daniels et al., 1969) that have less than five percent clay.

The intertongued contact between the A2 and the B horizon probably is not a depositional feature, but is more likely a post-depositional feature produced by soil formation. The eluvial-illuvial relation between the surface sand and the B horizon is demonstrated by the rariety of oriented clay in the A2 and the abundance of oriented clay in the subjacent B horizon. The oriented clay in the few thin bands in the surface sand is additional evidence.

From the preceeding lines of evidence it seems unlikely that most, or even a small part, of the A2 horizons in the Coastal Plain are a geologic deposit stratigraphically distinct from the underlying B horizon. It can be argued that on side slopes (Holocene surfaces) in dissected areas there is a good chance of movement of sand by running water. Our evidence suggests this origin for some side slopes, but the field and laboratory data also show that there may be a discontinuity between transported and residual material anywhere in the soil profile - in the A2, at the A-B contact, or in the underlying B horizon. In this case, formation of the A2 horizon is independent from, but probably influenced by, the presence or absence of a stratigraphic discontinuity.

If the A2 horizons are a weathering phenomenon, how are they formed and what factor or factors influence the rate and degree of formation? These A2 horizons are horizons of clay loss, and in some cases the underlying B horizon is the horizon of clay gain, an illuvial horizon. The best evidence for clay loss from the A2 horizon is its distinctly low clay content (Figure 1), and the eluvial-illuvial nature of the thin bands in the A2. Evidence suggesting that the underlying B horizons are illuvial horizons is found in the oriented clay below the A-B contact, and the clay bulge in the upper part of the profile on the dry parts of the landscape where the A2 horizons are the thickest and have the lowest clay content (Figure 5, Tables 4, 5). The increase in intensity of A2 formation from the wet to the dry soils probably is tied in with water-table regimes (Daniels et al., 1967). The mechanisms in moving clay from the A2 to the B2 horizon are not fully understood. Laboratory work with soil leaching columns and field observations plus micromorphological work with thin section (Thorpe et al., 1957, 1959) suggests that drying is necessary before clays can be moved. Once clays are dried, they are subject to dispersion upon rewetting and may be physically moved from the A-B contact to a point deeper in the profile.

APPLICATION TO STRATIGRAPHIC AND GEOMORPHIC WORK

One of the most difficult areas in the Atlantic Coastal Plain for a stratigrapher or geomorphologist to work is in the Sandhills area of North and South Carolina. Here are thick sandy sediments, old surfaces (in places at least), and soils with thick A2 horizons. Field experience suggests that on any given landscape the thickest A2 horizons are where the sediments are the sandiest. The thinnest A2 horizons are over clayey sediments. The landscape in the Sandhills is dissected so that water tables generally are low and A2 horizon formation should be at its maximum intensity (Daniels et al., 1967). In mapping surficial formations in the Sandhills area, it is unwise to map an eolian formation wherever the "surface sands" are two or more feet thick as Bartlett (1967) did in areas of Moore County. A2 horizons two feet thick are common in this area and many exceed five feet. In many parts of the Sandhills the solum may be 15 or more feet thick. A two- to five-foot A2 horizon is not out of proportion with the rest of the profile. We do not imply that all thick sands are A2 horizons, but we do point out that sands several feet thick can be and frequently are A2 horizons. If the unit is massive and overlies more clayey material (anywhere from 5 to 60+ percent clay) it is wise to consider a pedogenic weathering origin for the sand until proof to the contrary is found. Most evidence indicates a continuous soil mantle in the Coastal Plain and an A2 horizon is only the upper part of a Coastal Plain soil.

Soil weathering in the sandy rim on some Carolina bays must also be considered. Many of these bays may have rims of eolian sand, but between the Cape Fear and Neuse Rivers many bays have only 1 to 1 1/2 feet of sandy surface over a more clayey B horizon. Because the rim is as high or higher than any part of the local landscape and is one of the best drained parts of the landscape, A2 horizon formation is at its maximum intensity at these sites. Each site must therefore be evaluated because it is doubtful that one hypothesis is applicable to the sandy rims of all Carolina bays.

If we recognize that the upper part of surficial formations in the Coastal Plain has some of the thickest and oldest soils in the United States we may avoid the sampling difficulty that plagued Adams and Thom (1968) in their work in South Carolina. These authors sampled "just below the humic zone in the A2 soil horizon for size analysis at each sample location." Adams and Thom recognized the difficulties in such an approach but felt that the C horizon was too deep to sample and that the B horizon had been subjected to illuviation. Their choice of the A2 horizon, however, gave them the horizon with the most variable and intense eluviation. A better choice would have been part of the B2 horizon 20 to 30 cm below the contact to the A2. This part of the B would have some illuvial clay on the best drained sites, but in poorly drained sites there would be little change.

SUMMARY

Some geologists have considered the surficial sand in the southeastern Coastal Plain as a separate stratigraphic unit from the underlying clayey material (Pirkle, et al., 1964; Bartlett, 1967). But the bulk of the evidence from North Carolina indicates that the sands are a pedogenic weathering phenomenon associated with soil formation. This holds even on side slopes where there are stratigraphic discontinuities and on the bodies of eolian sand. These surface sands or A2 horizons

are primarily the result of clay movement from the A2 to the B2 horizon, or the destruction of clay and movement of its components to lower horizons. The resulting material is a gray to nearly white sand with 2 to 10 percent clay.

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PETROGRAPHY AND GEOPHYSICS OF THE ROCK HILL GABBRO PLUTON, YORK COUNTY, SOUTH CAROLINA

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ABSTRACT

A 16 square mile gabbroic pluton near Rock Hill, South Carolina, was studied by petrographic and geophysical means to determine the outcrop pattern, petrographic variation and subsurface shape. The pluton is a member of a series of gabbro bodies intruded into the Charlotte belt, probably during Late Paleozoic time.

Petrologic data indicates that the amount of plagioclase decreases from 95 percent at the center to 57 percent toward the contacts. Chemical alteration of primary minerals to hydrated secondary minerals is believed to have been caused by a hydrothermal fluid derived from the parental magma.

Gravity and magnetic surveys were undertaken to obtain information on contact relationships and thickness respectively. Magnetic survey data indicated that the pluton is relatively thin. The density contrast between gabbro and adjoining rocks was not always sufficient to warrant supposition concerning the location of the contact.

INTRODUCTION

Directly south of Rock Hill, South Carolina, a mafic pluton intrudes the Charlotte Belt rocks (Figure 1). The intrusion is comprised mainly of gabbro, anorthositic gabbro and anorthosite, and it is roughly oval in outcrop pattern. The pluton has an area of approximately 16 square miles and is now called the Rock Hill gabbro.

Over a century ago geological investigation was undertaken in York County by Tuomey (1844, 1848), who included a geological map in his second report. Geological research prior to 1965 was concerned primarily with mining and economic aspects of the prospects scattered throughout York County. Overstreet and Bell (1965a, 1965b) included discussions of the geology of York County. Butler (1966) has published a reconnaissance map of the county.

The Charlotte Belt as mapped by Butler (1966) in York County consists of medium to high rank metamorphic rocks, such as schist,

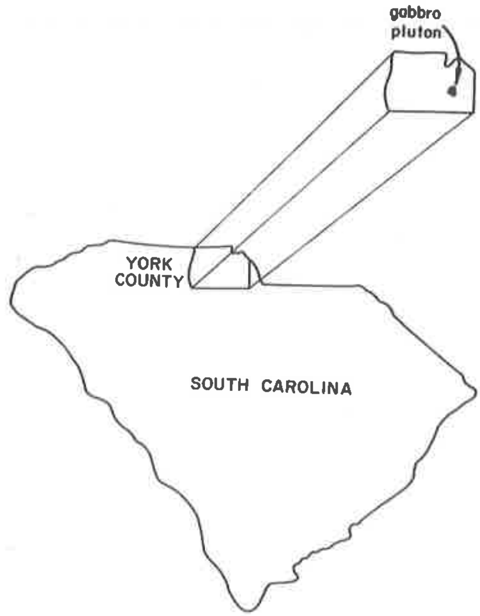

Figure 1. Index map of the York County area.

gneiss, amphibolite and metagabbro, which are intruded by igneous rocks that range in composition from gabbro to granite. Diorite, tonalite and foliated adamellite are the predominant igneous rocks.

The Charlotte Belt maintains an approximate northeasterly structural trend through York County. Phyllites, gneisses and schists that form the Charlotte Belt metamorphic rocks are thought to have formed from a sedimentary sequence of shale, graywacke, siltstone and tuff which is tentatively assigned an Early Paleozoic age (Overstreet and Bell, 1965a).

The Charlotte Belt rocks are crosscut by dikes of granitic composition and by quartz veins. Diabase dikes are commonly found transecting the northeast regional trend of the Charlotte Belt.

Field Relations

The Rock Hill gabbro stock crops out as an irregularly shaped oval, flattened on the northeast side (Figure 2). Consideration of the topography and soil associations proved an asset to field mapping. The gabbro body being less resistant to weathering and erosion than the quartz-rich rocks which circumscribe it, typically forms a topographic low. Contact areas may be located by a positive change in slope at the edge of the gabbro.

The Iredell-Mecklenburg-Davidson soil association forms from

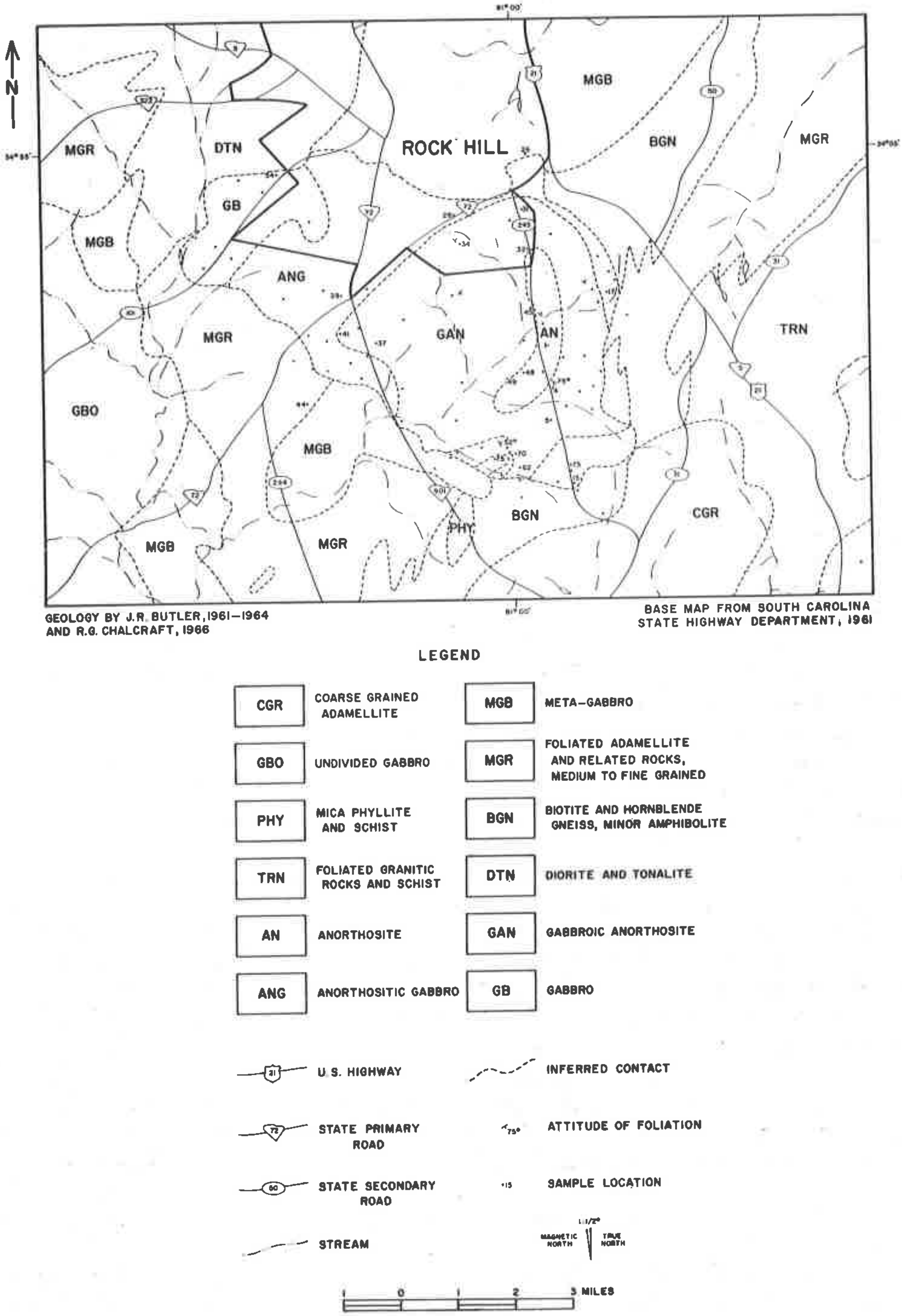

Figure 2. Geologic map of the Rock Hill area.

the weathering of gabbro. The Iredell soil group, which greatly predominates, typically forms from the weathering of basic rocks. The Iredell soils are moderately well drained, grayish brown to light brown loam. Soil color and absence of quartz grains aid in differentiating the gabbro from the units of the Charlotte Belt. The Rock Hill gabbro, unlike many of the Piedmont rocks, has more fresh rock at the surface in outcrops or residual boulders. Large gabbro boulders occur over nearly the whole outcrop area and were sampled for petrographic examination.

The Rock Hill gabbro is a medium to coarse grained, intrusive rock whose major mineral constituents are subhedral to anhedral. High plagioclase content, a characteristic of this pluton, is reflected in the color index of the rock. Parts of the gabbro are massive; others show the development of structures formed during the magmatic stage. This layering is readily visible in the eastern portion of the pluton. Large residual boulders and exposures have layering shown by alternation of lighter plagioclase-rich bands with darker bands rich in olivine and augite. The layers rarely exceed five inches in thickness, and generally average one inch or less. Strike of the layering is found to be subparallel to the gabbro contacts and dip decreases from near vertical in the eastern portion of the pluton to approximately 45 degrees east near the westernmost part (Figure 2). Mafic mineral layers are preferentially weathered. Linear alignment of olivine grains is detected on weathered surfaces by small pits in the rock that are surrounded by a yellow-brown stain. Chemical alteration of the plagioclase laths lightens the rock color and after alteration the darker mineral grains are easily visible within the whitish gray altered plagioclase mass. The chemical alteration accentuates the layers of lighter colored rock within the purple to greenish gabbro.

The Rock Hill gabbro pluton has an associated zone of pyroxene hornfels and a garnet and epidote-bearing skarn zone. The extent of the aureole cannot be accurately determined because the contact of the gabbro with adjacent Charlotte Belt rocks cannot be observed directly. Butler (1966) noted that the contact aureole is in some instances several hundred feet wide. Rocks of the aureole crop out along the southeastern margin of the pluton and the aureole attains its greatest width at the southern portion of the pluton.

Quartz veins and aplite and pegmatite dikes, two to five feet wide and of unknown length, are found within the gabbro pluton in the area immediately south of the Rock Hill city limits and along S. C. 72. The dikes and veins noted are too small to be included on Figure 2.

Acknowledgments

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PETROGRAPHY

Mineralogy

Microscopic examination reveals that plagioclase, olivine and hypersthene are the major primary mineral constituents of the Rock Hill gabbro. Hornblende, biotite, magnetite, pyrite and spinel are also present. Chemical alteration of varying intensity produced actinolite, chlorite, serpentine, epidote and muscovite in some areas.

The plagioclase normally occurs as subhedral laths ranging from 1-5 mm in length with an average of approximately 3 mm. Carlsbad and albite twins are observed throughout the gabbro suite. The anorthite content of the plagioclase was determined by measuring the maximum extinction angle of albite twins cut normal to (010) following the method of Kerr (1959). The anorthite content ranges from 35 to 75 percent. The greatest concentration of anorthite values is between 47 and 59 percent with 53 percent as the average value and the plagioclase is therefore mainly a sodic labradorite.

Alteration of the plagioclase is a very common feature. Saussuritization yielding epidote and sericite are the most common alterations. The plagioclase also alters to a clay mineral. Alteration to epidote is sporadic and of varying intensity throughout the extent of the gabbro pluton. Labradorite is the most resistant major mineral constituent while ferromagnesian minerals are always extensively altered in samples that display only slightly or moderately altered plagioclase.

The most abundant mafic mineral is augite with subhedral or anhedral grains up to 4 mm in length. Twinning is common with a (100) twin plane. Chemical alteration has greatly affected the augite in many localities and uralitization is the most frequent alteration process. A fibrous aggregate of actinolite forms at the expense of the augite. In some cases the outer ring of alteration, composed of antigorite with a few flakes of chlorite, gives way to actinolite α = pale green, β = grayish yellow, γ = light green, and finally to unaltered augite at the grain center. A rarer alteration is the chemical transformation of augite to epidote, calcite and chlorite.

Hypersthene is present in amounts to 8.5 percent as determined by modal analysis. Subhedral or anhedral crystals of hypersthene may reach 4 mm in maximum dimension, however, the mean grain size is about 2.5 mm. Strongly pleochroic reddish coronas surrounding polygonal olivine grains are common in specimens with higher modal percentages of hypersthene. Chemical alteration of hypersthene may take place yielding either a uralite or serpentine. Hypersthene alteration in the Rock Hill gabbro is predominantly uralitization in which actinolite is formed at the expense of the hypersthene.

Subhedral and anhedral grains of olivine may reach 5 mm in diameter. Chemical alteration yields magnetite in distinct grains or dendritic networks and is usually found along irregular fractures which characterize olivine grains. Serpentine and iddingsite are also developed by chemical alteration and the extent of their development appears to be a function of the intensity of the chemical alteration caused by hydrothermal fluids. Polygonal olivine grains are often surrounded by reaction rims of hypersthene.

The gabbroic rocks with a high mafic content contain up to 11 percent hornblende by volume. The hornblende shows strong pleochroism α = greenish brown, β = moderate brown, γ = moderate yellowish brown). The hornblende has two modes of occurrence in this rock suite. Primary hornblende crystals, generally subhedral and anhedral, with a maximum size of 2 mm, are found in unaltered gabbros. Aggregates of hornblende and magnetite are frequent in gabbros with high mafic content. Reaction rims of coronas enclosing grains of augite, magnetite, olivine and hypersthene are strongly pleochroic hornblende.

Spinel occurs as anhedral crystals less than 1 mm in length and is always found in contact with magnetite. The spinel has an emerald green color in section and was tentatively identified as the Mg-Fe spinel pleonaste.

The gabbroic rocks that lie close to the Charlotte Belt contact may contain anhedral grains of quartz up to 3. 5 mm in maximum dimension. Modal percentage of quartz may reach 18 percent.

Pyrite inclusions are found within magnetite grains. It is more common in the altered rock samples where it may reach approximately 4 percent of the total rock volume. It appears that the fluid that induced alteration contained sulfur. Combination of this sulfur and the iron that was already present may be responsible for the greater amounts of pyrite in the altered samples.

Biotite may approach 2 percent of the total rock volume. The biotite grains are very small, rarely exceeding 1 mm in diameter. The paragenetic association of biotite with hornblende, magnetite and spinel is apparent. The dark reddish brown to dusky yellow pleochroism of the biotite grains allows easy identification regardless of grain size. Biotite alteration is difficult to determine accurately in the extensively altered samples. Scattered flakes of chlorite and muscovite in the altered rock may have been derived from previously existing biotite.

Magnetite is the primary opaque mineral in euhedral to subhedral grains. The magnetite grains are very small and rarely exceed 1 mm in maximum dimension. Biotite, hornblende, spinel and magnetite form a mafic aggregate in many samples. Olivine alteration yields magnetite as discrete grains and as dendritic or arborescent networks of minute magnetite crystals. A moderate yellow stain on the altered specimens was related to magnetite grains and represents the iron oxide hydrate, limonite.

Fibrous aggregates of actinolite are the primary products of augite and hypersthene alteration. That the fibrous amphibole is strongly pleochroic indicates that it is actinolite rather than tremolite. The strong pleochroism indicates a high iron content.

Muscovite flakes less than 1 mm in diameter are found in the gabbros that have been subjected to chemical alteration. Minute muscovite flakes can be seen within the plagioclase and alteration of the plagioclase is the major source of the muscovite. Olivine alteration in selected samples has yielded a very fine-grained sericite like product. The muscovite observed in the Rock Hill gabbro is undoubtedly of secondary origin.

The antigorite variety of serpentine is present as exceedingly small grains formed by chemical alteration of augite and hypersthene. Antigorite typically forms the outermost ring of alteration surrounding the pyroxene grains. Antigorite veinlets irregularly cutting through plagioclase are indications of active hydrothermal chemical alteration, probably related to magma emplacement. An aggregate fibrolamellar structure is characteristic of both antigorite occurrences. Olivine alteration may also yield serpentine (chrysotile) along the irregular fractures within the grain.

The clinochlore variety of chlorite is a secondary mineral in the altered gabbros and occurs as small flakes generally less than 1 mm in length. Twinning according to the mica law was detected on a few larger flakes. Clinochlore was derived by the chemical alteration of primary augite and hypersthene but the conversion of biotite may also have been a source.

Epidote is observed as euhedral to anhedral columnar grains up to 1.5 mm in length. Clinozoisite of similar form has been identified but is rare. Epidote has not been identified as a primary mineral in unaltered samples.

Anhedral grains of calcite, 1 mm or less in length, are products of pyroxene alteration. The calcite occurrence is predictable since transformation of pyroxene to actinolite requires the release of calcium. When present, the calcite is associated with epidote, chlorite and actinolite.

Anhedral grains of garnet less than 1 mm in diameter occur within the gabbro at the contact. Precise identification of variety was not made.

Textural Variation

Samples from the Rock Hill pluton demonstrate a wide range of textural variation. Dominant is a hypidiomorphic-granular texture. The subhedral or anhedral grains of plagioclase, augite and olivine form an interlocking network that accounts for the toughness and hardness of the rock. Subophitic texture characterized by plagioclase laths partly enclosed in augite, occurs in most thin sections, and ophitic texture, complete enclosure of plagioclase by augite, is commonly observed. Most informative were the varieties of poikilitic texture.

Plagioclase chaddacrysts appear in hornblende, hypersthene or olivine. Poikilitic textures involving plagioclase inclusions were not restricted in areal distribution but appear to occur irregularly within the gabbro body. Samples with higher olivine content display well-developed intersertal texture in which polygonal grains of olivine occupy the interstices between the plagioclase laths. Although thin sections of this gabbro suite show a great variety of textures the most common are the hypidiomorphic-granular to subordinate subophitic textures.

Modal Analysis

Eighteen samples were point counted (Table 1) in order to determine the modal variation within the gabbro body. Identity change numbers (Chayes, 1956) ranged from 31 to 114. The chart compiled by Chayes (1956) was used to maintain an average major mineral analytical error of 2.0 or less. Most samples of the Rock Hill gabbro were sufficiently coarse grained to require 3 thin section. The range of sections counted per sample was one to four.

Conclusions

The triangular diagram (Figure 3) shows that the predominant rock types are gabbro and anorthositic gabbro. The terminology is that of Buddington (1939) from his work in the Adirondack Mountains of New York. In this work he set up arbitrary boundaries to delineate the gabbro series as follows: anorthosite 0-10 percent mafic minerals, gabbroic anorthosite 10-22.5 percent mafic minerals, anorthositic gabbro 22.5-35 percent mafic minerals, gabbro 35-65 percent mafic minerals and mafic gabbro greater than 65 percent mafic minerals. The diagram indicates that the plagioclase-rich end of the gabbro series predominates in the Rock Hill suite. This classification system was also used on Figure 2.

The gabbro may have as much as 10.5 percent hornblende or 11.9 percent olivine. The biotite and hornblende content is typically low. When hornblende is found in greater abundance in the gabbros it probably has formed by late magmatic transformation of a pyroxene. RH-44 and RH-72 are amphibolite and metagabbro. Sample RH-26 has a high (17.9 percent) quartz content and therefore is properly termed a tonalite. The origin of sample RH-26 is questionable and may be related to an incorporation of a more felsic rock type by the intruding gabbroic magma.

The chemical alteration that is common throughout the pluton has no regular pattern of distribution and the degree of alteration is random. Mafic minerals alter more readily than the plagioclase. This transformation is accompanied by a lightening of the rock color and a probable change in the chemical composition. The alteration was probably brought about by the penetration of fluids that were formed as

Table 1. Modal Analysis Data in Volume Percent. Sample Locations are Shown on Figure 2.

Sample	plag.	hbld.	augite	hyp.	oliv.	biot.	mag.	epid.	serp.	act.	qtz.	musc.	An.
4	75.1		18.2	0.1	6.6								57
5	66.4		21.5	3.4	7.3		1.4						72
15	59.6	0.5	25.0	2.1	11.9	0.1	0.3		0.3				43
17	57.4	3.6	31.8	3.0		1.3	2.9						43
26	53.4	10.5	8.1	3.4		0.1	3.6			0.9	17.9	2.1	40
28	63.8	1.3	24.7	3.5	5.7		1.0						55
31	66.9	2.3	22.7	4.4	1.1	0.3	2.3						51
32	95.0	0.3	1.8	0.7	0.1	0.3	1.4	0.3		0.1			53
34	71.4	3.5	17.5	3.2	1.1	0.5	2.8						55
37	68.8	tr	23.8	2.8	4.0	tr	0.6						61
39	54.8		35.3	3.3	5.4	tr	1.2						59
41	61.1		33.2	2.3	1.4	0.5	1.2			0.3			65
48	83.4		2.5	0.3	8.2		0.8		2.8	0.1		1.9	51
49	90.4		2.1	0.9	5.6	0.3	0.7	tr					54
54	63.3	2.7	27.4	4.1	1.0	0.6	0.9						59
67	69.1	5.7	11.6	6.4	4.7	1.1	1.0						32
70	66.2	9.3	5.3	8.6	6.5	1.9	2.4						42
75	57.0	0.8	24.5	7.7	9.7	0.1	0.2						61
95	81.5	0.6	6.9	4.9	5.0		0.6	0.2					52
44	36.2	62.1	0.7			0.2	tr	tr			0.5	0.3	
72	42.5	47.9	2.8				6.7		0.1				

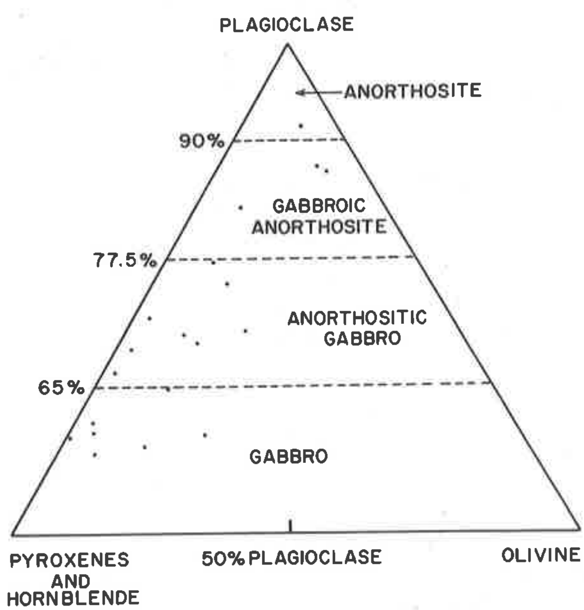

Figure 3. Triangular variation diagram for gabbro varieties.

the products of differentiation of the gabbroic magma.

The predominant minerals from five samples taken from the contact aureole are epidote, augite, plagioclase, actinolite and magnetite. These minerals occur as small grains (less than 1.5 mm in length) that form an interlocking network and account for the rock's hardness. Relict ophitic and subophitic textures are visible with the enclosed plagioclase grains partially converted to epidote.

GRAVITY

In March of 1967, 47 gravity stations were established. The station locations were determined by the superposition of a one-square-mile grid pattern upon the York County highway map. Gravity values were determined with a standard Model 121 Worden Gravimeter and elevation determination for each station was obtained from two Model M-1 Paulin altimeters.

The Bouguer gravity anomalies were plotted on the York County highway map and these anomaly values were then used for reference points in the construction of the Bouguer anomaly contour map (Figure 4), which has a contour interval of 2 milligals. The approximate northeast trend of the Charlotte Belt is reflected in the gravity isogals. The gravity values increase toward the northwest, across the regional trend and this increase can be correlated with the predominant rock type. Rocks lying south and southeast of the gabbro body form a predominantly granitic terrain, whereas mafic rocks predominate north and northwest of the pluton.

The inflections in the isogals can frequently be related to the density contrast between the gabbro and its contact rocks. The contact relationship is easier to define where the gabbro lies adjacent to foliated adamellite and related rock types. In areas where the gabbro adjoins diorite, metagabbro, and hornblende and biotite gneisses, the isogal inflection is absent or inconspicuous.

MAGNETISM

Field Procedure and Data Reduction

In March, 1968, 248 magnetic stations were established in order to generate a magnetic contour map that would provide information on the extent of the gabbro body at depth. Data points were contoured with a 500 gamma interval (Figure 5). A Jalander flux gate magnetometer was used to measure the vertical component of the earth's magnetic field at stations, with a 0.2-0.4 mile separation along roads and creeks in the Rock Hill area.

Rock samples from 68 locations (Figure 6) were examined with a magnetic susceptibility bridge, Model MS-3, manufactured by the Geophysical Specialties Company. The purpose of the susceptibility

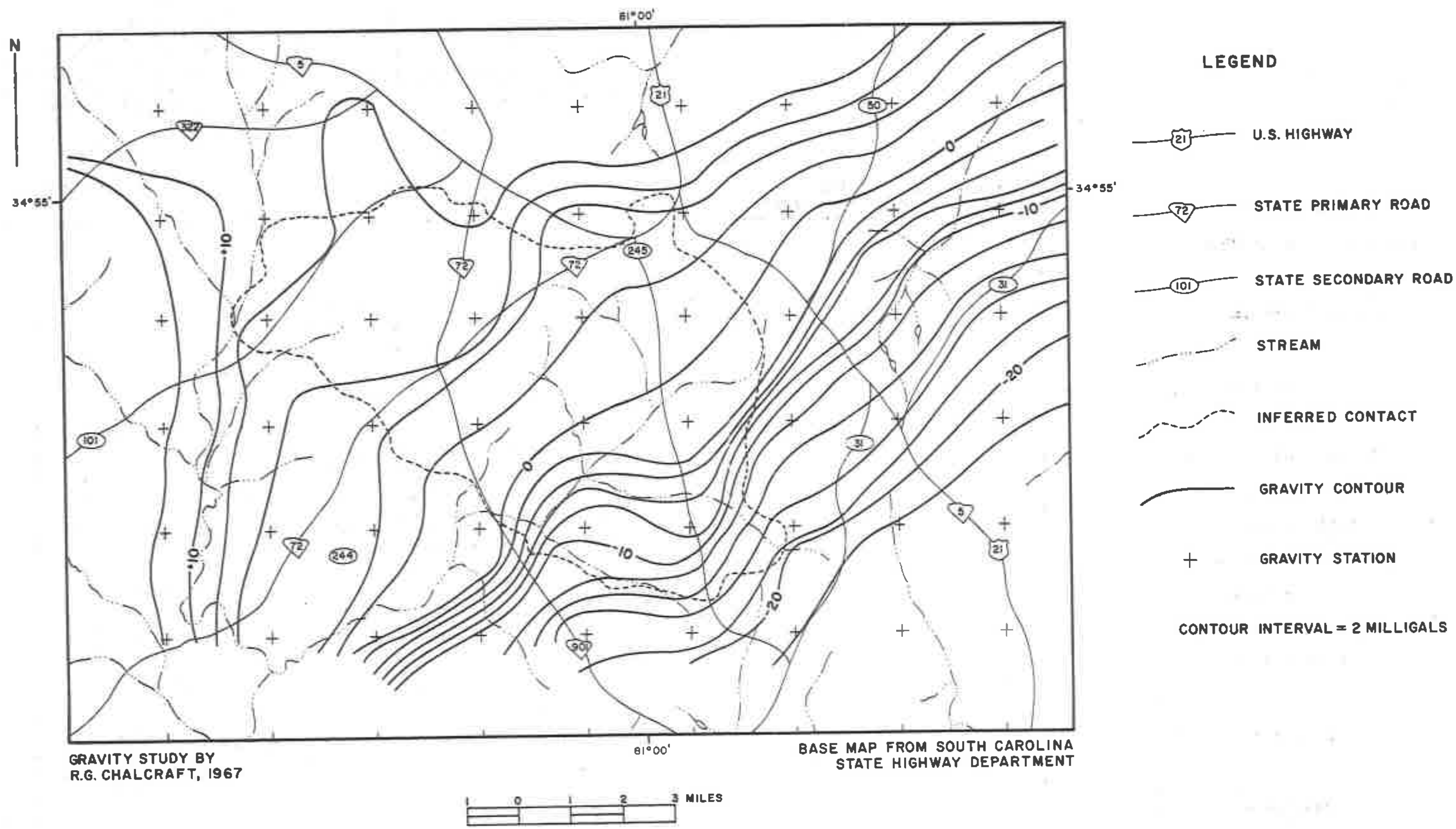

Figure 4. Bouguer anomaly map of the Rock Hill area.

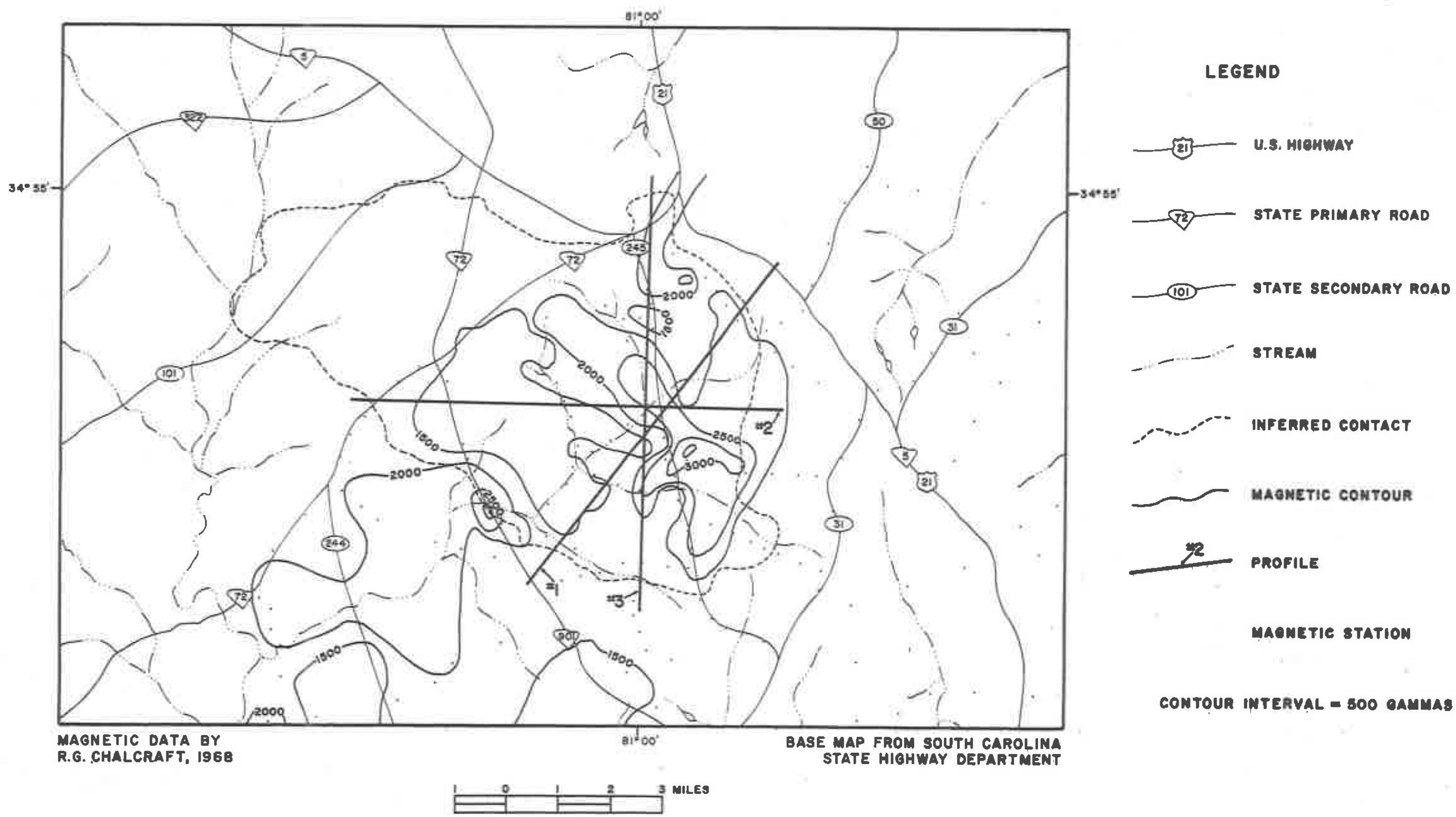

Figure 5. Magnetic map of the Rock Hill area.

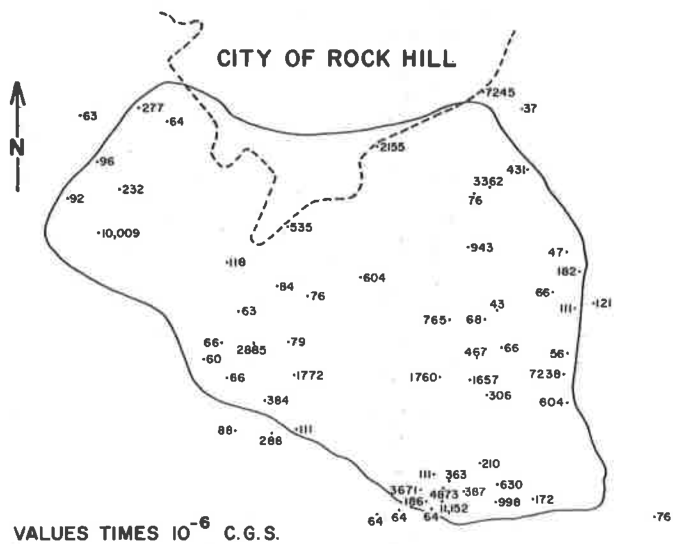

Figure 6. Susceptibility determination locations.

determinations was to discover if the observed magnetic anomalies were due to the variation in magnetic susceptibility. Susceptibility and magnetic anomaly values were averaged and used to calculate a correlation coefficient. A value of 0.220 was obtained, indicating that the observed magnetic anomalies are not due to variation in the magnetic susceptibility.

If it is assumed that the gabbro pluton consists of a vertical cylinder magnetized vertically, a susceptibility value of approximately $3,200 \times 10^{-6}$ c.g.s. would be necessary to account for a pluton whose thickness is approaching infinity. This susceptibility value is abnormally high for a gabbro, thereby indicating that susceptibility is not the major contributor to the magnetism of the gabbro. The susceptibility value necessary to account for the anomaly increases as the thickness of the cylinder decreases.

Four magnetic models were constructed by:

1. Assuming that the pluton was best represented by a vertical cylinder magnetized vertically with its top at the surface of the ground.
2. Varying the thickness of the pluton with respect to the radius.

The four models (Figure 7) are plotted with anomaly profiles in Figures 8, 9, 10 and 11.

An observed magnetic anomaly may be due to:

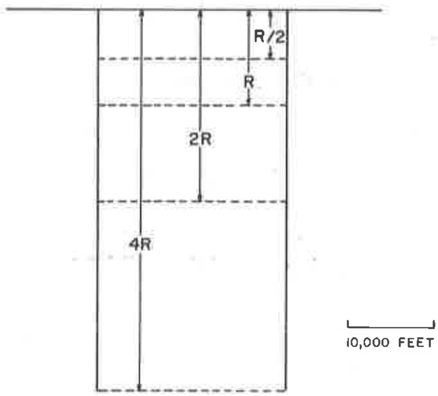

Figure 7. Model representations.

1. Variation in magnetic susceptibility.
2. Change in shape of the body at depth.
3. Natural remanent magnetization.

Calculations have indicated that the measured magnetic susceptibility is not large enough to cause the magnetic anomaly observed. This observation is by no means uncommon, for natural remanent magnetization often accounts for a high percentage of the rock magnetization. The anorthositic gabbro probably obtained an undetermined amount of thermoremanent magnetization when the magnetic material cooled through its Curie Point in the presence of a magnetic field.

Evaluation of the approximate thickness of the gabbro body was obtained by comparison of model profiles with anomaly profiles. Figures 8, 9, 10 and 11 show the anomaly profiles, with profiles exhibiting model thickness of 5,500 feet, 11,000 feet, 22,000 feet and 44,000 feet respectively. Comparison of observed profiles and profiles constructed from the models indicate that the assumed models greatly oversimplify the magnetic interpretation. Close scrutinization of the profiles indicates that the slope of the model profile representing a thickness of 11,000 feet most closely resembles the observed profiles. This fact suggests that the pluton is of the order of 10,000 feet thick.

Examination of Figure 5 reveals two magnetic maxima related to basic rocks in the Rock Hill area. The larger anomaly lying near the southwestern contact is probably coincident with an older metagabbro body.

The Rock Hill anorthositic gabbro pluton exhibits an anomaly of approximately 3,800 gammas in the southeastern quarter. This maximum may be related to a variation in shape within the pluton or perhaps a magnetic neck or feeder.

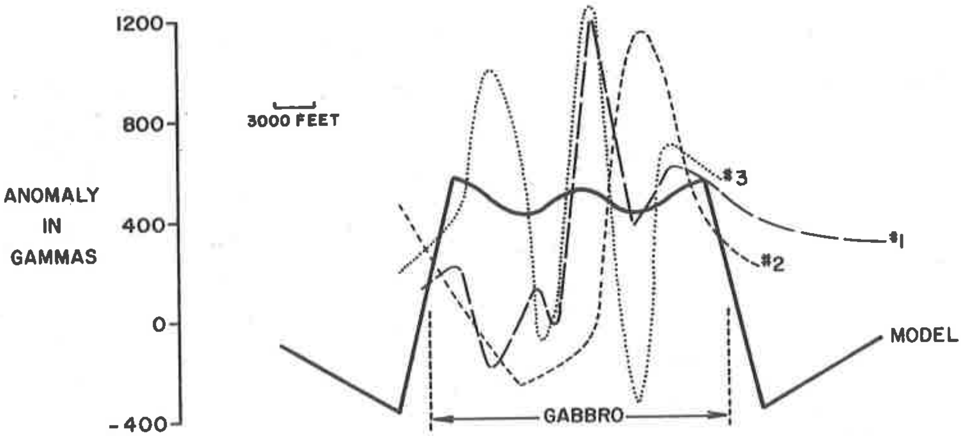

Figure 8. Profile comparison with model thickness equal to one-half the radius of the pluton. Profiles 1, 2, and 3 are located on Figure 5.

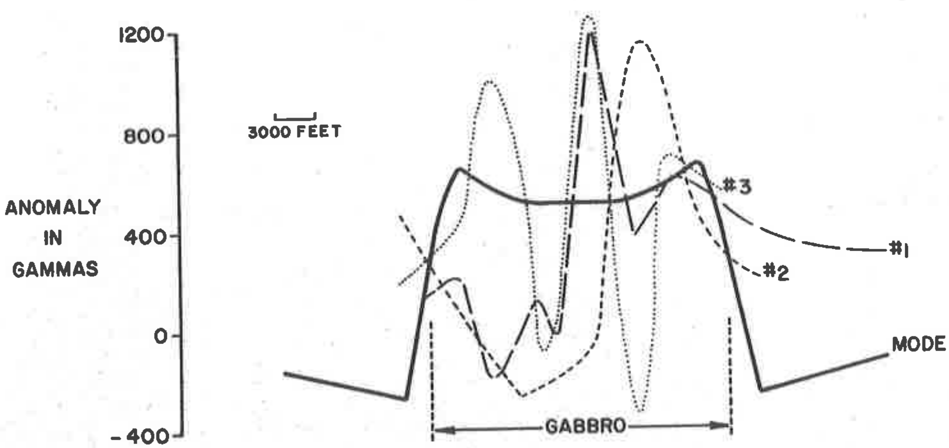

Figure 9. Profile comparison with model thickness equal to radius of the pluton.

DISCUSSION OF RESULTS

Differentiation

Fractionation of the mafic magma was accomplished by fractional crystallization and relative movement of crystals and liquid. Inspection of the geologic map (Figure 2), which was drawn by plotting the plagioclase percentage of each sample taken, indicates that the more mafic rocks form the outer margins of the pluton. As differentiation proceeded, the plagioclase content of the rocks increased toward

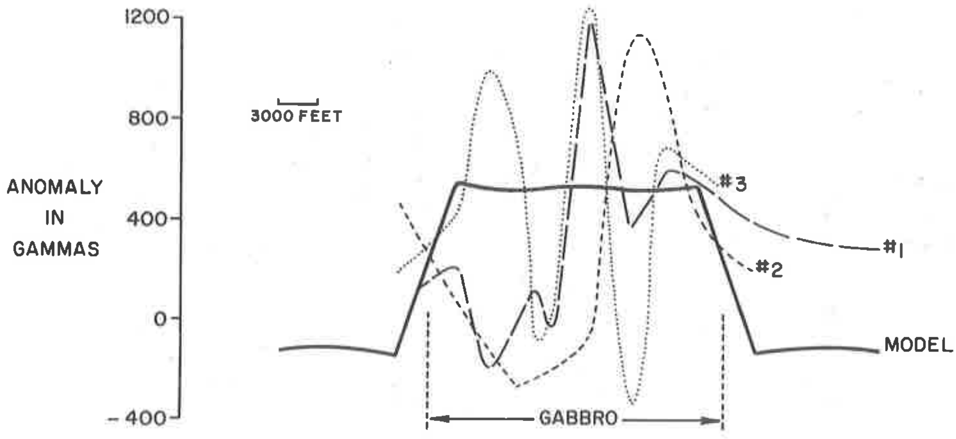

Figure 10. Profile comparison with model thickness equal to twice the radius of the pluton.

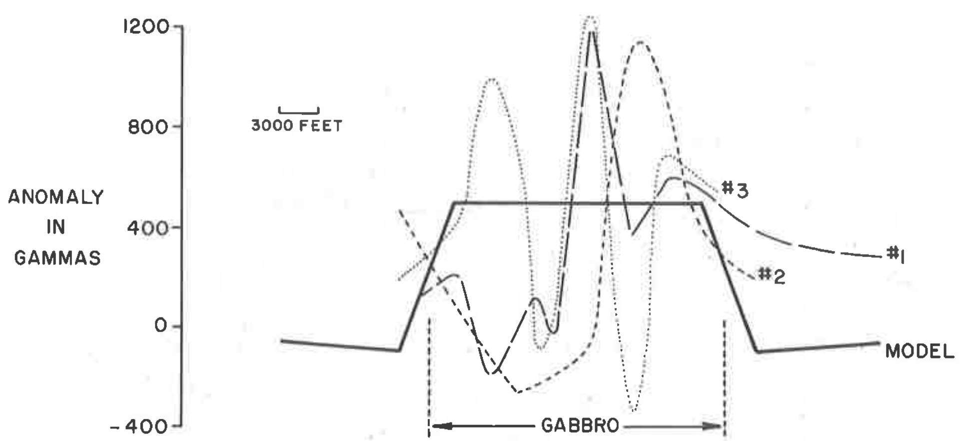

Figure 11. Profile comparison with model thickness equal to four times the radius of the pluton.

the center of the intrusion. Crystal settling, which affects the percentages of minerals, and magmatic flow are responsible for the layering within the pluton. Thin section examination reveals incomplete development of Bowen's discontinuous reaction series. Plagioclase feldspars of unaltered gabbros display a wide range of composition, an absence of resorption effects, and also lack zoning of any type. Poikilitic textures commonly occur. The poikilitic occurrence indicates that plagioclase crystallization began relatively early. It is likely that the labradorite either proceeded olivine crystallization or crystallized simultaneously. These observations indicate the progressive trend of crystallization formulated by Bowen. Locally the volume percentage of hornblende or biotite increases. Subhedral crystals of augite have been partially converted along irregular boundaries within the grain to hornblende. This conversion may be an indication of a locally higher water content of the magma. In some gabbro samples the content of hornblende may reach 10. 5 percent by volume.

Emplacement

The foliated structures, discordancy, size and tectonic environment indicate a mesozonal emplacement (Buddington, 1959). Foliated structures, which are common to mesozonal intrusions, are well developed in the Rock Hill gabbro. Petrologic data suggest a late or post-tectonic emplacement. Mesozonal plutons are most frequently intruded during the late stages of the development of a tectonically active belt (Badgley, 1965).

Age Relations and Metamorphism

The Rock Hill gabbro pluton is believed to be a member of the younger series of gabbro plutons that have intruded the Charlotte Belt throughout North and South Carolina. The younger gabbros are usually circular or kidney shaped in outline and typically form topographic lows (Overstreet and Bell, 1965). The most significant feature of the intrusions is the absence of a later metamorphic effect (Overstreet and Bell, 1965).

McCauley (1960) recognized three intrusive gabbro plutons in Newberry County, South Carolina, which he believes to be post-metamorphic intrusions. The paragenesis described by McCauley corresponds closely to the mineral suite of the Rock Hill gabbro.

Morgan and Mann (1964) concluded from gravity data that the gabbro body, which lies within the ring dike near Concord, North Carolina is a post orogenic intrusion. The Concord pluton shows no metamorphic effect and therefore probably represents a younger gabbroic intrusion.

The gabbro of the gabbro-meta-gabbro complex in the Charlotte Belt of Mecklenburg County, North Carolina, shows no indication of metamorphism (Hermes, 1966). It is likely that this intrusion is also a younger gabbro.

Potassium-argon radioactive dates from micas indicate that the last major metamorphic period in the southeast may have occurred approximately 250 million years ago. Hadley (1964) points out that the potassium-argon dates are easily reset. The presence of a metamorphic epoch 250 million years ago is not thoroughly documented at this time. Therefore one may only conclude that the gabbros are younger than the well documented 360 million year age date (Hadley, 1964).

The intrusion of a gabbroic magma is almost always accompanied by the formation of a hornfels zone (Winkler, 1965). The Rock Hill gabbro has an aureole of pyroxene hornfels. The nature of this hornfels zone has been discussed earlier.

Summary of the Geologic History

The Rock Hill gabbro pluton is believed to have been intruded into the metamorphosed rocks of the Charlotte Belt less than 360 million years ago. The country rocks were metamorphosed or intruded during a previous metamorphic period that most likely occurred approximately 360 million years ago (Long, Kulp, and Eckelmann, 1959).

Alteration of the primary minerals was induced by the introduction of hydrothermal fluids derived from the parental magma. The pluton was not subjected to post-consolidation metamorphism of a regional nature.

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VIRGINIA METAMICT MINERALS:

PYROCHLORE-MICROLITE SERIES

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ABSTRACT

Members of the pyrochlore-microlite series are relatively common in Amelia County, Virginia. Contrary to common belief spectrochemical studies indicate most specimens are not typical microlite (the Ta end member), but varieties in which Ta and Nb are about equal. This is especially true for Champion and Rutherford pegmatite specimens; a Morefield pegmatite specimen showed Ta > Nb and abnormally high Pb, U, Th, and TCe. A good correlation between chemical composition, degree of metamictization, and unit cell size (ranging from 10.38 A to 10.51 A) was not found. Metamict specimens recrystallize between 400° and 500°. Heat-treated samples are essentially pure pyrochlore-microlite through 700°. A hexagonal $Ca(Ta,Nb)_4O_{11}$ phase appears at 800° but disappears before 1000°. Orthorhombic $Ca(Ta,Nb)_2O_6$ and a phase close to cubic $(Ta,Nb)O_2F$ develop between 900° and 1000°. In Morefield specimens a tetragonal $(Ta,Nb)O_2$ phase forms at 900° and above. Specimens from Amelia, Powhatan, and Prince Edward Counties are described.

INTRODUCTION

Microlite has been recognized in Amelia County, Virginia, for nearly 90 years (Dunnington, 1881). This study shows that contrary to common belief most specimens from the area are not typical microlite, but have compositions near the middle of the series whose end members are pyrochlore (Nb rich) and microlite (Ta rich). The results of heat-treatment of metamict pyrochlore-microlite have been studied to see at which temperatures recrystallization occurs and what phases develop by the decomposition of the mineral. Although data on decomposition phases have been reported before, a better identification of these has been made, owing to the availability of newly published X-ray data for Ca-Ta-Nb oxide compounds.

CHEMICAL DATA

The pyrochlore group consists of three minerals, pyrochlore, microlite, and betafite, having the general formula $A_{2-x}B_2O_{6-x}X_{1+x}$. In this formula the major A cations are Na, Ca, TR(TCe and TY), U, Th (more rarely K, Sr, Ba, Mg, Fe^2, Mn, Pb, Sb, Bi). The B cations are Nb, Ta, Ti (more rarely Zr, Fe^3, Al, Sn). X is generally F and OH, but also may be O.

The theoretical composition for pyrochlore is $NaCaNb_2O_6F$. Usually there is a deficiency in the A cations, especially in Na, and there is a decrease in F accompanied by an increase in the amount of OH. There is much heterovalent isomorphous substitution in groups A and B. Na is often replaced by TR, U, Th, etc., and Nb may be replaced by Ta, Ti, Zr, etc. The theoretical composition for microlite is $(Ca,Na)_2Ta_2(O,OH,F)_7$. Again the chemical composition is variable, especially in the A cation positions, and in the anion positions. Because of the large scale isomorphism between Ca and Na, valence compensation is brought about by simultaneous replacement of part of the O by F and OH groups. A complete series exists between pyrochlore and microlite, although most analyses are close to the end members (Palache and others, 1944). Betafite, in which the major B cation is Ti, has not been found in Virginia and is not included in this study.

Semiquantitative spectrographic analyses (Table 1) were made of seven specimens from Amelia County (described below with the occurrences of Virginia pyrochlore-microlite). These include one from the Champion pegmatite, one from the Morefield pegmatite, and five from the Rutherford pegmatite. A study of the results verify they are in the pyrochlore group, but in almost every case near the boundary line separating pyrochlore from microlite where the ratio of Nb:Ta is near 1:1[1/]. This differs from the original analysis of Rutherford microlite given by Dunnington (1881), which has been quoted in numerous publications (e.g., Palache and others, 1944, p. 751; Vlasov, 1966, p. 521). This also departs from the generally accepted notion that compositions of minerals in this series are usually near the ends of the series. Dunnington (1881) indicated that he had difficulty separating Ta_2O_5 and Nb_2O_5 in his analysis, and that the specific gravities of his separations showed them to be mixtures rather than pure oxides. However, he was certain Ta exceeded Nb.

The remaining chemical constituents for the specimens present no special problems. The relatively high Na, Ca, TR, U, Th, and Pb

1/ The percentages quoted for Nb and Ta oxides in Table 1 are not precise. Because of the difficulty in estimating major amounts of highly refractive elements like Nb and Ta, and the complex natures of the spectra of these elements, the percentage values quoted for them are not as accurate as one would like.

Table 1. Semiquantitative Spectrographic Data on Amelia County, Virginia, Pyrochlore-Microlite. Elements Reported as Oxides. Analyst: F. W. Barley, American Spectrographic Laboratories, San Francisco.

Element	V3027	V406	V3048	V3218	V407	V410	V408
Na	0.7	1.5	1.25	1.5	3.	3.	1.5
Al	0.12	0.2	0.1	0.08	0.12	0.06	0.15
Si	0.4	1.5	0.25	0.2	0.12	0.25	1.25
Ca	12.5	8.	10.	15.	12.	10.	7.5
Ti	1.25	1.75	1.5	0.85	1.75	0.6	1.
Mn	0.5	0.5	0.25	0.12	0.15	1.	0.15
Fe	0.8	0.6	0.5	0.2	0.15	0.3	0.05
Cu	0.015	0.01	0.01	0.01	0.006	0.01	0.01
Sr	0.04	0.15	0.08	0.15	0.02	0.07	0.05
Y	0.08	0.03	0.06	0.04	0.1	0.05	0.03
Zr	0.025	0.06	0.025	0.025	0.025	0.02	0.015
Nb	35.	15.	30.	30.	35.	35.	40.
Mo	0.015	0.01	0.01	0.01	-	0.01	0.01
Sn	1.25	0.6	0.3	0.4	1.75	0.85	1.
Ba	0.01	0.06	0.02	-	0.02	-	-
La	0.025	0.3	0.06	0.05	0.07	0.07	0.02
Ce	0.2(?)	2.	0.2(?)	0.2(?)	1.	0.35	0.4
Nd	0.2	0.35	0.35	0.25	0.1	0.2	0.12
Sm	0.04	0.12	0.06	0.06	0.05	0.12	0.05
Yb	0.015	0.03	0.02	0.01	0.01	0.025	0.02
Ta	35.	40.	35.	30.	30.	30.	25.
Pb	0.4	7.	0.25	0.75	0.85	0.6	0.2
Bi	0.015	0.15	0.04	0.1	0.03	0.06	0.02
Th	-	3.	0.6	0.5	1.25	0.3	0.2
U	0.85	3.	1.75	0.2	1.5	-	0.3

Mg, less than .01 in each
Cr, less than .02 in each
Dy, Nb interference

V3027, Champion pegmatite
V406, Morefield pegmatite
V3048, V3218, V407, V410, V408, Rutherford pegmatite

are typical of the A cations, and the high Ti and Sn are typical of B cations. As in other localities TCe exceeds TY (a ratio of about 10:1). The other trace elements can likewise be separated into the two cation groups, perhaps with the exception of Si which may be an impurity.

The Morefield microlite (V406) differs considerably from the other specimens in that it has much Pb and more U, Th, and TCe. The heating characteristics of this mineral are also somewhat different from the other specimens, a fact to be discussed below. A lead-rich microlite has also been reported by Safiannikoff and Van Wambeke (1961).

LITERATURE SURVEY OF X-RAY DATA

The general nature of the structure of the pyrochlore-microlite series (space group Fd3m, Z = 8) has been known about 40 years (Gaertner, 1930). Tables of X-ray powder data have been included in numerous publications. Indexed powder data for pyrochlore have been given by Gaertner (1930), Lima-de-Faria (1958), Hogarth (1961), Van der Veen (1963), Perrault and Guy (1964, ASTM X-ray powder diffraction data cards 17-746, 17-747), and Vlasov (1966, p. 505). Indexed powder data for microlite and its varieties are included in papers by Kerr and Holmes (1945), Arnott (1950), Safiannikoff and Van Wambeke (1961), Matias (1963), and Haapala (1966, p. 62). Numerous values for the unit cell constant $\underline{a}$ have been published. Strunz (1966, p. 174) has summarized the range: for pyrochlore $\underline{a}$ = 10.34 to 10.43 Å; for microlite $\underline{a}$ = 10.37 to 10.43 Å. Presumably composition variability and degree of metamictization effect the cell size.

The results of heat-treatment of metamict members of the series have been reported by several investigators. Gaertner (1930) observed pyrochlore and a perovskite phase in metamict pyrochlore heated at 1400° for 24 hours. Upon heating pyrochlore (koppite) Sorum (1955) found a phase close to loparite (itself close to $Na_2Nb_2O_6$); he also reported columbite lines in his X-ray pattern, but these were also on the unheated samples and are probably an impurity. Van der Veen (1963) found pyrochlore, perovskite, and fersmite in heated pyrochlore. Lima-de-Faria (1964) observed that heating of metamict pyrochore forms pyrochlore with a cubic phase close to NbO_2F and a fersmite-like phase. Upon heating to very high temperatures, Harris (1965) obtained a barium niobate phase from the decomposition of hydrated barium pyrochlore (pandaite). Gasperin (1958, 1960) found pyrochlore and strüverite phases after heating a titanium-rich uranium pyrochlore (ellsworthite). Gorzhevskaya and Sidorenko (1962) reported the following decomposition products of heat-treated pyrochlore: fersmite, perovskite, lueshite, fergusonite, rutile (strüverite), and samarskite. Hogarth (1961), however, heated pyrochlore at 900° for 20 hours in a vacuum without forming new phases.

Microlite, when heated to 800° by Reuning (1933), yielded microlite with traces of perovskite and unknown phases. Gorzhevskaya and Sidorenko (1962) reported a lueshite-like phase ($NaTaO_3$?) after heating microlite. Lima-de-Faria (1964) found a mixture of microlite with a cubic phase close to TaO_2F and other X-ray reflections (especially at 1300°).

X-RAY DATA FOR VIRGINIA SPECIMENS

Prior to this study of Virginia pyrochlore-microlite, some X-ray diffraction data have appeared in the literature on unanalyzed "microlite" from the Rutherford pegmatite. Glass (1935) observed that an olive-colored variety gave a poorer X-ray pattern than an amber variety, although the patterns were essentially the same. Kerr and Holmes (1945) obtained films without lines, but with diffused bands, from an Amelia specimen. Excellent indexed powder data are given by Arnott (1950) for ignited "microlite" from Amelia. Unheated, the mineral was partially metamict. Before ignition $\underline{a}$ = 10.441 A, after ignition $\underline{a}$ = 10.424 A (temperature not given). Photographs of X-ray powder films of Amelia material are given by Kulp and others (1952) including a metamict pattern and a well-crystallized pattern obtained by heating at 1000° for 10 minutes. Reuning (1933) found traces of perovskite and unknown phases in heat-treated "microlite" from Amelia. Lima-de-Faria (1958) found only the "microlite" phase in his Amelia specimen heated at 1000° for one hour.

The degree of metamictization of Virginia pyrochlore-microlite is quite variable, even within a single deposit. Recrystallization of metamict specimens was achieved by heating them in air for an hour between 400° and 500°. No good correlation was found between the amount of metamictization and the radioactive-element content (Table 2). However, if one only considers the Rutherford specimens a general relationship is suggested. Neither is there a good correlation between unit cell size and chemical composition or degree of metamictization. Accurate cell parameters for specimens from the Amelia area vary from $\underline{a}$ = 10.40$\pm$.01 to 10.45$\pm$.01 A for unheated specimens. Values as low as 10.38$\pm$.01 A were determined for some samples heated in the range 500° to 700°. After heating at 1000° the maximum value observed was 10.41$\pm$.01 A. In an earlier preliminary study an unheated Morefield specimen gave 10.51$\pm$.01 A. Material from Prince Edward County (V3038) gave a value of 10.34$\pm$.01 A when heated at 700° for an hour. An unheated pyrochlore formed by alteration of a probable samarskite from Powhatan County (Mitchell, 1965) had $\underline{a}$ = 10.52$\pm$.05 A.

The variation of the unit cell size with treatment at different temperatures was studied for several specimens. The maximum value is nearly always found on unheated specimens. Generally, for metamict specimens the parameter has a minimum value after heating from about 500° to 600°. The value increases with increasing temperature up to about 1000° to a value slightly smaller or approximately equal to the relatively large unheated value (Table 2). An obvious exception to this was on unmetamict Rutherford specimen V3218 where $\underline{a}$ varied the following way: unheated, 10.41$\pm$.01; 600°, 700°, 10.42$\pm$.01; 800°, 900°, 1000°, 10.41$\pm$.01 A. The tendency for the cell size to decrease upon heating has been pointed out by Hogarth (1961) for pyrochlore, and by Arnott (1950) for Amelia "microlite." Harris (1965), for barium

Table 2. Metamictization, Radioactive Element Content, and Unit Cell Parameters for Amelia County, Virginia, Pyrochlore-Microlite.

Pegmatite and specimen no.	Nb:Ta	Degree of metamictization	r. a. element as oxide		Unit cell parameter, a		
			U	Th	Unheated	600° C	1000° C
Champion							
V3027	Nb=Ta	very meta.	0.85	0	-	10.38	10.41
Morefield							
V406	Nb<Ta	weakly meta.	3.0	3.0	10.45	10.42	10.41
Rutherford							
V3048	Nb<Ta	very meta.	1.75	0.6	10.40	10.38	10.38
V3218	Nb=Ta	not meta.	0.2	0.5	10.40	10.42	10.41
V407	Nb>Ta	very meta.	1.5	1.25	-	-	-
V410	Nb>Ta	not meta.	0	0.3	10.43	-	-
V408	Nb>Ta	weakly meta.	0.3	0.2	10.44	-	-

pyrochlore, found a to decrease in size until recrystallization was complete, with a constant value after that until decomposition. For uranium microlite Matias (1963) determined a constant cell parameter in the range 600° to 1200°.

In addition to changing the unit cell parameter, heating in air also brings about decomposition of the mineral to form other chemical phases. Samples representing three pegmatites (Champion, Morefield, Rutherford) were heated in air an hour each at various temperatures, generally at 100° intervals from 400° through 1000°. Although there were slight variations between deposits the following general results were observed. Samples heated at 400° were still metamict. Recrystallization occurred at 500°, and relatively pure pyrochlore-microlite persisted in the 500°, 600°, and 700° samples. In all samples heated at 800° hexagonal $Ca(Ta, Nb)_4O_{11}$ was present along with pyrochlore-microlite. With longer periods of heating at this temperature the new phase is better developed. This same compound was found in the 900° samples, but disappeared at higher temperatures. At 1000° pyrochlore-microlite is still present, but the additional phases include a definite cubic substance, probably $(Ta, Nb)O_2F$, and orthorhombic $Ca(Ta, Nb)_2O_6$. One or both of these new phases were also in samples heated 7 hours at 1080° The heat-treatment of Morefield microlite showed some variations from the above general results. Traces of cubic $(Ta, Nb)O_2F$ appeared at temperatures as low as 700°. Instead of $Ca(Ta, Nb)_2O_6$, tetragonal $(Ta, Nb)O_2$ began to form at 900°, and at 1000° the sample was a mixture of microlite, tetragonal (Ta, Nb)O_2, and a major amount of (Ta, Nb)O_2F. As mentioned above the chemical composition of Morefield microlite was somewhat different from the other specimens of this study. In addition to the major phases discussed here, other very minor unidentified phases were present in some of the pyrochlore-microlite samples. Quartz lines (especially the intense value at 3.34 A) were sometimes observed.

The phases formed by heating pyrochlore-microlite were identified by comparing them to ASTM X-ray powder diffraction data cards of pure compounds. The correlation was excellent in each case in spite of much isomorphism in the Virginia specimens. The presence of both Ta and Nb in these new phases is assumed on the basis of the chemical data of Table 1. Gasperin (1963) synthesized $CaTa_4O_{11}$, and indexed the powder data using a hexagonal cell in which $\underline{a}$ = 6.22 A and $\underline{c}$ = 12.25 A (ASTM data card 15-679). Values determined in this study vary somewhat from sample to sample. Frevel and Rinn (1956), using $\underline{a}$ = 3.896 A, indexed cubic TaO_2F (ASTM data card 9-21). The cell value in the present study varies some, but was commonly 3.88 ± .01 A. The possibility that this compound with $\underline{a}$ = 3.88 A might be some other cubic phase should be pointed out. Without a chemical analysis for it, the composition only can be inferred, since for cubic compounds with one cell parameter several possibilities often exist. According to Jahnberg and others (1959) there is a cubic form of $CaTa_2O_6$ having $\underline{a}$ = 3.886 A. Also $NaNbO_3$ (ASTM data card 14-603) has similar powder data, if one only considers its more intense reflections. Other compounds with similar X-ray data have been outlined by Van der Veen (1963, p. 123). The structure of orthorhombic $CaTa_2O_6$ was determined by Jahnberg (1963). Her indexed X-ray powder data (ASTM data card 17-550), with $\underline{a}$ = 11.07 A, $\underline{b}$ = 7.505 A, and $\underline{c}$ = 5.378 A, are very close to values observed in this study. The cell size for the Virginia material is approximately $\underline{a}$ = 11.00 A, $\underline{b}$ = 7.48 A, and $\underline{c}$ = 5.38 A. In their paper Magnéli and others (1955) described tetragonal NbO_2 with a structure having a subcell close to rutile. The lines were indexed (ASTM data card 9-235) using the parameters $\underline{a}$ = 13.71 A and $\underline{c}$ = 5.985 A. The Virginia parameters are similar.

The observed recrystallization of metamict pyrochlore-microlite between 400° and 500° correlates well with observations made by other investigators. Matias (1963) recrystallized uranium microlite at 470°. Numerous others, in differential thermal analyses of pyrochlore and microlite, have found a strong exothermic peak from a little below 500° to 600° (Kerr and Holland, 1951; Kulp and others, 1952; Adler and Puig, 1961; Hogarth, 1961; Soboleva and Pudovkina, 1961, p. 340). A correlation between the high-temperature phases formed by heating Virginia specimens with phases reported in the literature is difficult, but the fersmite phase of Gorzhevskaya and Sidorenko (1962), Van der Veen (1963), and Lima-de-Faria (1964) and the loparite phase of Sorum (1955) are essentially the same as the orthorhombic

$Ca(Ta, Nb)_2O_6$ phase reported here. The perovskite-like structure reported by Gaertner (1930), Reuning (1933), and Van der Veen (1963), and the lueshite-like phase of Gorzhevskaya and Sidorenko (1962), are nearly identical to $(Ta, Nb)O_2F$. The earlier reported rutile and strüverite may possibly be related to the tetragonal $(Ta, Nb)O_2$ found in the Virginia specimens.

OCCURRENCES OF VIRGINIA PYROCHLORE-MICROLITE

Of the six occurrences of pyrochlore-microlite described here, only specimens from the Morefield and Rutherford pegmatites have been described in detail before. Some of the others have been mentioned only in the literature. In the following outline original references are given with pertinent data where they apply. Then a description is given of the specimens used in this study.

Amelia County, Champion Pegmatite

Lemke and others (1952, p. 116) mentioned that octahedral crystals of microlite 1/64 to 1 inch or more in diameter occur in albite at the Champion pegmatite. Pyrochlore, associated with quartz and feldspar in the wall zone, was also reported. How the minerals were identified was not given. Fitzgerald and Mitchell (1961) indicated an X-ray analysis verified the pyrochlore from the pegmatite, but further studies by them showed the determination was not unambiguous.

Crystals collected and analyzed by the writers are reddish to greenish-brown subhedral octahedrons (up to 1 cm across) having a resinous luster. They are associated with iron-oxide stained massive white plagioclase along with spessartine and muscovite. Spectrochemical analyses of one crystal (V3027) indicate a variety on the borderline between pyrochlore and microlite. The specimen is nearly completely metamict.

Amelia County, Morefield Pegmatite

Microlite was first reported at the Morefield pegmatite by Pegau (1932, p. 60). It was verified by Glass (1934) who later described it (1935). She reported dark olive-colored octahedrons up to 1.5 cm in diameter embedded in bluish albite. A study showed the mineral to be radioactive. Small granular masses and flattened crystals of pale honey-colored microlite in albite were also observed by Glass (1935). In a study of the morphology of fifty crystals from the pegmatite, Donnay (1941) found the following forms in their order of decreasing importance: {111}, {110}, {311}.

Dark olive-green subhedral resinous crystals (V406), smaller than 0.5 cm across, and embedded in pale blue-green cleavelandite albite, were analyzed in this study. Table 1 shows that Ta exceeds Nb so microlite is the correct name. Of all specimens analyzed Pb, U, Th, and TCe are higher than in the others. In spite of the high radioactivity of the mineral X-ray studies show it is only weakly metamict.

Amelia County, Rutherford Pegmatites

The earliest known occurrence of microlite in Virginia is the Rutherford pegmatites. In fact, most specimens in museums or in the literature from "Amelia, Virginia" are from this deposit. Dunnington (1881) published a chemical analysis and gave a description of the mineral. Fontaine (1883) reported that masses weighing at least 8 pounds occur in pegmatite No. 1, and well-formed octahedrons up to 4 cm across occur especially in pegmatite No. 2. He described the mineral and considered its paragenesis. A transparent hyacinth-red variety, cut as a gem, was described by Hidden (1885). The morphology of Rutherford microlite crystals was studied by Feist (1885) who found the following forms: {111}, {110}, {311}, {100}, {221}.

Further studies of Rutherford microlite by Glass (1935) classified the mineral into two distinct types. The first is dark olive-buff to dark olive and occurs in single octahedral crystals having lusterous faces. The specific gravity varies from 5.5 to 5.7. It is strongly radioactive and gives poor X-ray patterns (presumably because of metamictization). Dunnington (1881) reported 1.59% UO_3 in this microlite. It is associated with massive chatoyant albite. The second type is reddish-yellow to dark amber, occurring as distorted modified octahedral crystals and fractured crystalline masses. The specific gravity varies from 5.9 to 6.0. It is weakly radioactive and gives a good X-ray pattern according to Glass (1935). It is found in reticulated cleavelandite albite. This type resembles that described by Hidden (1885) both in color and specific gravity.

The possibility that pyrochlore might occur in the Rutherford pegmatites was mentioned by Fontaine (1883). This was not verified until the present study. Uranium pyrochlore (hatchettolite) has erroneously been listed as occurring in the deposit by Gordon (1918) who wrote, "the species hatchettolite being found only there...". At that time the well-known locality for that mineral was in Mitchell County, North Carolina.

Because of the highly variable nature of specimens from the Rutherford pegmatites five different types were selected as representative of the deposits. Generally no distinction is made here between the No. 1 and No. 2 pegmatites, since this information is not known for most of the museum specimens studied. Specimen V3048 is a submetallic dark-olive subhedral mass, measuring 5 by 6 by 6 cm. It is very metamict. Although it is nearly on the borderline of pyrochlore-microlite, it falls into the microlite category. Another specimen (V3218) right on the compositional borderline occurs in platy cleavelandite as waxy to resinous bright olive-green, highly modified octahedral crystals about 1 cm across. The mineral is nearly unmetamict, and is from the No. 2 pegmatite. Specimen V407 is a resinous, dark reddish-brown octahedral crystal about 6 cm across. It is attached to cleavelandite which also contains columbite-tantalite. The specimen is very metamict, and has a composition near the border, but falls in the pyrochlore range. Another specimen (V410) which falls in the pyrochlore range is a vitreous dark-brown to amber-yellow octahedron over 2.5 cm across. It occurs between well-developed plates of bluish cleavelandite and is essentially not metamict. Specimen V408 consists of vitreous transparent honey-yellow octahedral crystals (less than 0.5 cm) embedded in a quartz-feldspar-muscovite matrix rock. Studies show this to be weakly metamict pyrochlore.

The occurrence of both microlite and pyrochlore at the Rutherford pegmatites has thus been verified, although each seems to have a composition near the borderline between the two minerals. On the basis of the above discussion it would be premature to draw relationships between compositional type and color, size, or habit.

Powhatan County, Herbb No. 2 Pegmatite

Microlite, associated with cleavelandite, tantalite-columbite, beryl, and other minerals, was reported at the Herbb No. 2 pegmatite by Griffitts and others (1953, p. 175). The mineral was not described and the basis for its identification was not given. Several trips to the pegmatite failed to verify this occurrence, although monazite (V4310) resembling microlite was collected, as well as tantalite-columbite and other minerals mentioned in the literature.

Powhatan County, White Peak No. 1 Pegmatite

X-ray studies of moderate orange-pink alteration rinds on a metamict uranium-niobium oxide mineral (possibly samarskite) from the White Peak No. 1 mine, indicated they are composed of a single substance with the pyrochlore structure (Mitchell, 1965). The three most intense X-ray reflections were at 3.03 A, 1.86 A, and 1.59 A, and $\underline{a} = 10.52 \pm .05$ A. Chemical data are not available. The alteration of samarskite to form pyrochlore has been studied by Kalita (1964).

Prince Edward County, Darlington Heights

On the northern side of State Road 665, about three miles east of Darlington Heights, small fragments of columbite-tantalite occur in an unmined weathered pegmatite. An X-ray study of some of these crystals (V3038) showed associated pyrochlore-microlite. The mineral occurs in very small quantity and is reddish-brown, vitreous, and highly fractured. Although quite metamict it was verified by a study of X-ray films of unheated material. Another sample heated at 700° for an hour gave excellent data with $\underline{a}$ = 10.34$\pm$.01 A. Unfortunately not enough pure material is available for a determination of chemical data. This would be of interest since the cell constant is so different from other localities reported here. In his study of radioactive columbite Heinrich (1962) thought this mineral might be euxenite, and observed strong, minute fractures and yellow-brown internal reflections in polished sections.

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GEOLOGY OF DAVIE COUNTY TRIASSIC BASIN,

NORTH CAROLINA

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ABSTRACT

Davie County Basin, a small irregular-shaped graben located in the west-central Carolina Piedmont, is believed to be an outlier of Dan River Basin.

Upper Triassic nonmarine strata within the basin can be divided into two mappable intertonguing facies: (1) basin margin polymictic conglomerate, and (2) basin center sandstone-siltrock-clayrock.

The conglomerate facies, of interpreted alluvial fan origin, consists of subrounded and rounded, very poorly sorted boulder- to granule-size clasts that float in a very coarse-grained plagioclase lithic arkose matrix. It is very thick-bedded, crudely stratified, and contains sandstone lenses that display cut-and-fill, and large scale trough cross-stratification. Rocks assigned to the sandstone-siltrock-clayrock facies exhibit fining-upwards cycles and are considered to be of fluvial origin. Medium- and coarse-grained, poorly sorted plagioclase arkoses and lithic arkoses grade upward through a decrease in grain size and bed thickness into massive-appearing maroon mudrocks.

Sandstone mineralogy, clast petrology, and lithofacies distribution indicate the basin was filled from the eastern and western sides, and that the sediment was dominantly derived from nearby medium and high rank metamorphic sources. Lesser amounts of detritus was derived from low rank metavolcanic rocks of the Carolina Slate Belt.

INTRODUCTION

General Statement

Davie County Basin is an irregular-shaped asymmetric graben that occupies 14 square miles in northwestern Davie and southern Yadkin Counties, North Carolina (Figure 1). The basin, located in the Piedmont Physiographic Province, forms a slight topographic lowland approximately 50 feet below the surrounding Piedmont surface.

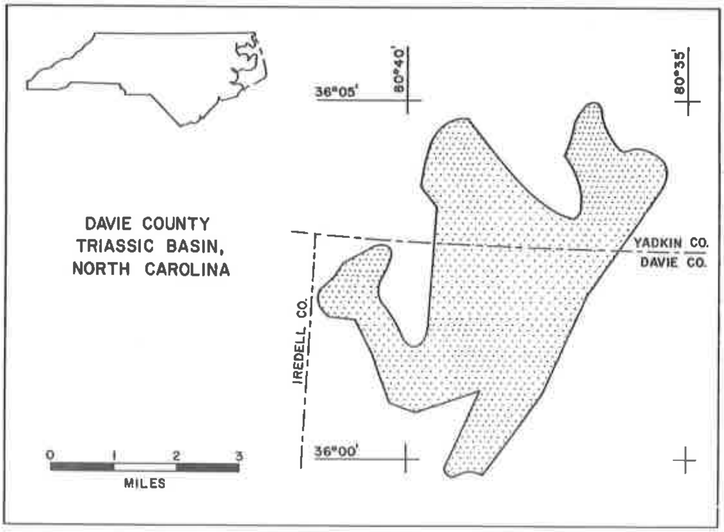

Figure 1. Index map of Davie County Basin, North Carolina.

Sedimentary rocks within the basin were assigned a probable Late Triassic age by Jurney and Bacon (1927, p. 5), Brown (1932, p. 525), LeGrand (1954, p. 16), and Stuckey and Conrad (1958, p. 40) on the basis of lithic similarity and analogous structural setting with the Dan River Triassic Basin, located 25 miles to the northeast (see Geologic Map of North Carolina).

Field work for this project took place at scattered intervals during the spring and summer of 1966. Geologic mapping was done on U. S. Department of Agriculture aerial photographs (scale 1:20,000) and data then transferred to North Carolina county road maps (scale 1:63, 360) since no large-scale topographic base was available for the area.

Previous Investigations

The presence of Upper Triassic sedimentary rocks in northern Davie County was first recognized by Jurney and Bacon (1927, p. 5) in their county soil report. Brown (1932), using the soil report as a guide, made a reconnaissance geologic map of the area but failed to map the continuation of Triassic rocks into Yadkin County. LeGrand (1954) published a generalized geologic map of Davie County as part of a ground water study, and mapped a small Triassic outlier southeast of Harmony in Iredell County.

The geologic map of North Carolina (Stuckey and Conrad, 1958) shows Triassic rocks extending into Yadkin County for almost a mile. It does not differentiate units within the Triassic nor does it show the distribution of diabase dikes.

G. H. Espenshade of the U. S. Geological Survey is currently mapping the Winston-Salem Two-degree sheet that includes the northern part of Davie County Basin.

Regional Geology

Davie County Basin is situated in the northeast-southwest trending Inner Piedmont belt of King (1955). Rocks within this belt are dominantly medium- and high-grade regionally metamorphosed gneisses and schists that have been intruded by small bodies of synorogenic to late orogenic granitic rocks. Stratigraphic relations and ages of these rocks (whether Precambrian and/or Paleozoic) are unknown at this time.

Rock units in the vicinity of Davie County Basin include granite gneiss, metagabbro, metadiorite, amphibolite, biotite gneiss, kyanite-garnet schist, biotite-amphibole gneiss, augen gneiss, and sillimanite schist (Espenshade, 1967, Open-file map). Small bodies of granitic rock, pegmatite, and quartz veins are intrusive into this metamorphic complex. Rock types immediately surrounding the basin are shown on the geologic map (Figure 2).

Gross structure of the Inner Piedmont in this area has not been determined with any degree of certainty. King (1955, p. 353) believes the structure is an anticlinorium characterized by low and undulatory foliation. Overstreet and Griffitts (1955, p. 551) speculate that the structure is a geanticline overturned to the west, and Espenshade (1968, p. 34) suggests a belt of large-scale gentle folding.

Acknowledgments

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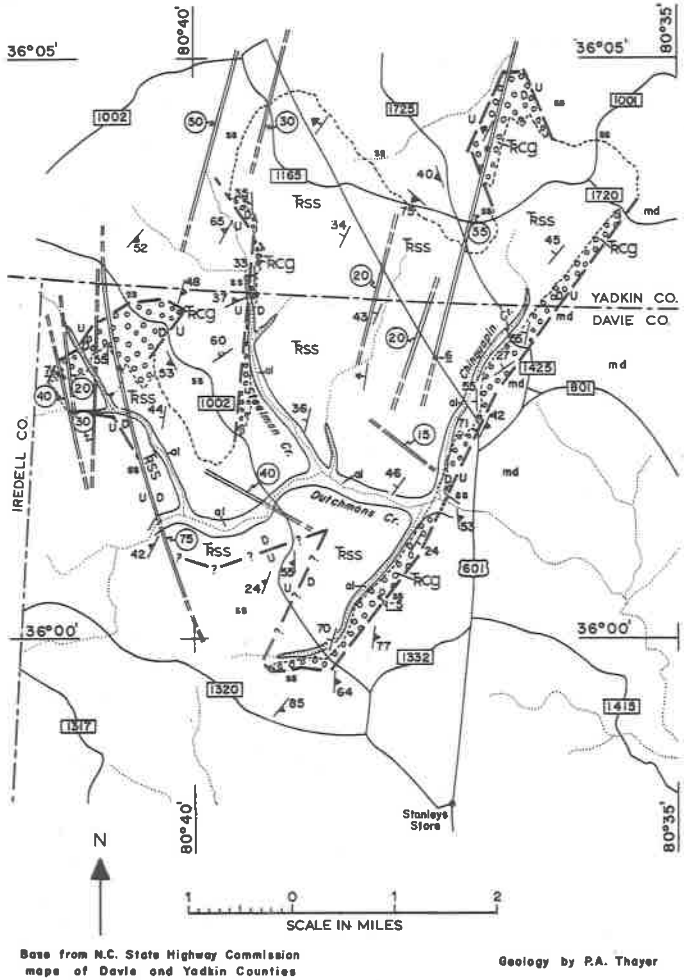
This paper is extracted from a previously unpublished Ph. D. dissertation submitted to the University of North Carolina at Chapel Hill in 1967.

TRIASSIC SEDIMENTARY ROCKS

General Statement

Upper Triassic sedimentary strata within Davie County Basin

Figure 2. Geologic map of the Davie County Triassic Basin, North Carolina.



EXPLANATION

POST-TRIASSIC	al	ALLUVIUM
TRIASSIC (?)		DIABASE DIKES. NUMBERS IN CIRCLES INDICATE AVERAGE THICKNESS IN FEET.
	Ꞧcg / Ꞧss	Ꞧcg, CONGLOMERATE FACIES; Ꞧss, SANDSTONE-SILTROCK-CLAYROCK FACIES.
PRE-TRIASSIC		METAMORPHIC ROCKS
	ss	SERICITE SCHIST
	md	MEDIUM- TO COARSE-GRAINED METADIORITE

CONTACTS

- APPROXIMATE
- COVERED

FAULTS

- U/D NORMAL FAULT, APPROXIMATE
- U/D NORMAL FAULT, COVERED
- ? U/D ? NORMAL FAULT, QUESTIONABLE

ATTITUDE OF ROCKS

- 60 STRIKE AND DIP OF BEDS
- APPROXIMATE STRIKE AND DIP OF BEDS
- 30 STRIKE AND DIP OF FOLIATION

CULTURE

- 1725 ROAD
- STREAM

are nonmarine clastic deposits consisting of interbedded maroon-, tan-, and green-colored conglomerates, sandstones, siltrocks, and clayrocks (mudrock terminology from Ingram, 1953). Abrupt lateral and vertical changes in stratification, texture, composition, and color are characteristic of the sequence. Coarse-grained rocks are dominant along the basin margins; finer-grained clastics occur in the central part of the basin.

Extremely poor exposures, owing to low relief, deep weathering, and abundant vegetation, have prevented formal subdivision of the sedimentary sequence. Instead, two laterally interfingering facies have been mapped: (1) basin-margin conglomerate facies, and (2) basin center sandstone-siltrock-clayrock facies. The contact between the two is gradational and is approximately shown on the geologic map (Figure 2).

Conglomerate Facies

Good exposures of well indurated maroon- and green-colored polymictic conglomerates interbedded with coarse-grained, very poorly sorted sandstones are found on the eastern margin of the basin along Chinquapin and Howard Creeks. Highly weathered conglomerate crops out in scattered areas along the northwest boundary (Figures 2 and 3).

Extreme variability in grain size and clast composition within short distances is the dominant feature of this facies. Maximum observed particle sizes are found along the basin margins; size decreases rapidly towards the basin center. This indicates that the basin was filled from both sides since it is generally recognized that the size of clastic particles decreases in a downcurrent direction (Pettijohn, 1957, p. 529).

The majority of clasts "float" in a matrix of very coarse-grained plagioclase lithic arkose (sandstone terminology after Folk, 1968) with the number of grain-to-grain contacts generally low. Where grains are in contact they are usually either long or tangential (Taylor, 1950, p. 707). Indented concavo-convex contacts are rare and have formed where durable rock fragments have been pressed into softer argillaceous ones.

Clasts are dominantly pebble and cobble size (Figures 3 and 4) although boulder size ones are not uncommon along the basin margins. Most are subround and round (Pettijohn, 1957, p. 59), but angular and subangular clasts also occur. Rounding is a function of size and composition. Boulder and cobble size particles display the highest rounding values. Softer rock fragments (regardless of size) like schist and phyllite are more rounded than durable fragments in the same size class.

Most clasts are spheroidal or disc-shaped. Particle shape is determined by the degree of foliation in source area rocks; spherical particles are non-foliated, whereas disc-shaped ones are foliated.

Figure 3. Weathered boulder, cobble, and pebble conglomerate along northwestern margin of basin. Note disc-and bladed shaped schist fragments breaking down along cleavage planes leaving more durable quartz clasts (upper left).

Figure 4. Poorly sorted, well indurated cobble, pebble, and granule conglomerate along eastern side of basin. Note high degree of rounding displayed by most clasts.

Clasts consist of Precambrian and/or Paleozoic metadiorite, metagabbro, sericite schist, quartz, hornblende gneiss, pink granite, microcline, augen gneiss, quartz-epidote rock, epidotized aphanitic rock (metamorphosed acid volcanic rock?), and quartz-feldspar porphyry. Clast petrology indicates the basin was filled from both sides and that most clasts were derived from local source areas marginal to the basin. However, the quartz-epidote rock, epidotized aphanitic rock, and quartz-feldspar porphyry fragments were probably derived from metavolcanic rocks of the Carolina Slate Belt located southeast of the basin (G. H. Espenshade, personal communication, 1968). Klein (1969, p. 1829) believes all Triassic basins of eastern North America were characterized by a dispersal pattern of sedimentation from all sources marginal to the basin. Davie County Basin certainly fits this model.

The conglomerate is very thick-bedded, crudely stratified, and extremely poorly sorted. Large scale trough cross-stratification, cut-and-fill, and particle imbrication are rare accessory features. Fragments of Araucarian conifers (Araucarioxylan?) up to 60 cm long are locally abundant. Combinations of sedimentary structures such as these are generally believed to be characteristic of alluvial fan deposition (Krynine, 1950; Dunbar and Rodgers, 1957; Klein, 1962; and Allen, 1965).

Sandstone-Siltrock-Clayrock Facies

Rocks assigned to this facies are medium- and coarse-grained, poorly sorted, grayish-green and yellowish-brown sandstones with a subordinate proportion of reddish-brown siltrocks and clayrocks.

Modal analyses of five sandstones from this facies, using the pointcounting technique are presented in Table 1. Thin section samples were cut perpendicular to bedding and 1000 points counted per slide. A square grid was chosen according to grain size of the sample; the point distance being larger than the largest grain on the slide. At the 95 percent confidence level maximum error is probably less than 1.5 percent of the real value of any mineral (Van der Plas and Tobi, 1965, p. 87). Feldspars were stained using the method of Bailey and Stevens (1960) in order to facilitate their identification.

Samples 1, 2, and 4 plot as arkoses, and samples 3 and 5 as lithic arkoses on Folk's (1968, p. 124) triangular composition diagram. Micas were tabulated under rock fragments in contrast to Folk's opinion (1968, p. 123) that they should be ignored in determining the rock name. Since plagioclase is the dominant feldspar in 4 of 5 samples (Table 1) these rocks might be aptly termed plagioclase arenities. Sandstone composition corresponds to the findings of Thayer (1967) for the Dan River Basin and indicates that in both areas the sandstones were mostly derived from nearby medium- and high-grade metamorphic source areas.

Davie County sandstones are typical examples of tectonic arkoses formed in response to vertical deformation and block faulting. Sorting is generally very poor (σ_I 2-4Φ) and ranges from moderate (σ_I 0.71-1.0ϕ) to extremely poor ($\sigma_I > 4.0\Phi$). Most sandstones are texturally immature although a few are submature. Grains are always angular or subangular.

Mineralogically these sandstones are extremely immature. All are characterized by an abundant suite of unstable major and accessory minerals, and a low quartz-feldspar ratio. Feldspars of the same

Table 1. Modal Analyses of Sandstones from Davie County Triassic Basin.

Sample No.	1	2	3	4	5
Mineral					
common quartz	31.8%	15.2%	28.6%	24.3%	15.4%
composite quartz	15.2	27.9	18.1	8.6	20.1
sutured quartz	10.5	5.2	7.9	11.2	9.6
untwinned K-feldspar	4.8	4.5	3.2	4.1	7.3
twinned K-feldspar	8.9	3.5	4.7	3.8	1.1
plagioclase	13.7	24.2	9.3	17.3	15.4
aphanitic rock fragments	-	-	1.2	1.4	2.8
gneiss fragments	4.1	2.9	3.6	2.1	3.1
schist fragments	6.3	3.5	4.7	1.6	7.4
micas	1.7	5.2	3.6	2.1	2.7
matrix	1.8	5.5	7.4	15.6	9.4
cement	-	1.0	4.3	3.8	1.9
opaque minerals	0.2	1.3	1.7	2.1	1.3
others	-	-	1.8	2.0	2.6
Total	100.0	99.9	100.1	100.0	100.1

species are usually a mixture of fresh and altered types. This, according to Folk (1968, p. 127), is the diagnostic feature of a tectonic arkose and indicates that soil formation as well as rapid erosion was taking place in the source area (Krynine, 1950, p. 150).

Accessory heavy minerals include many unstable types such as green hornblende, epidote, pyroxene, staurolite, sillimanite, sphene, monazite, kyanite, garnet, and apatite. Stable heavy minerals that dominate the suite include zircon, tourmaline, and rutile. Most heavy mineral grains are euhedral and show little evidence of wear. This indicates a relatively short transportation history for these sediments.

Siltrocks and clayrocks are reddish-brown and typically weather quite rapidly to a reddish-brown or yellowish-brown saprolite. These have not been studied petrographically, but field evidence shows they consist of varying proportions of siltstone, claystone, silt shale, and clay shale.

Sandstones are thick- and medium-bedded, and display both uniformly parallel stratification, and medium and large scale trough cross-stratification. Siltrocks and clayrocks are medium-bedded, massive, and show uniformly even stratification. Rhythmic alternations (fining-upwards cycles) of sandstone and mudrock are common with lighter-colored sandstone grading upward through a decrease in grain size and bed thickness into darker-colored mudrock. Channel lag conglomerates consisting of rock fragments, intraformational mudrock breccia, and chips of silicified wood are occasionally found at the base of the sand units. Green reduction spots, mudcracks, and carbonate concretions are locally present in the maroon mudrocks. This combination of sedimentary structures is characteristic of fluvial deposition (Dunbar and Rodgers, 1957; Klein, 1962; Allen, 1965, 1965a). Sandstones are interpreted as channel deposits (channel-lag, point bar, and channel bar) and the mudrocks are believed to be overbank deposits (swale fill and floodbasin).

TRIASSIC IGNEOUS ROCKS

Eleven dolerite dikes, ranging from 15 to 75 feet wide, occur within the map area (Figure 2). Dikes less than 5 feet wide were not mapped. They display a linear to curvilinear map pattern and are up to 4 miles long; most are less than 2 miles long. Dikes were easily traced on the ground by following boulder trains of spheroidally weathered dolerite. Three sets of dikes are present, striking N10° - 15° W., N20°E., and N40°-50°W. Although dips could not be measured in the field most dikes are believed to be vertical because of their straight map pattern. Also, by analogy, dikes thought to be of similar age in the Dan River and Deep River Basins are generally vertical (Reinemund, 1955, p. 55; Thayer, 1967, p. 112). Purplish-gray contact aureoles are almost always present near dike contacts.

Modal data for two dikes (sample localities 6 and 7, Figure 2) are presented in Table 2. Sample 6 displays isogranular texture and sample 7 is intergranular. These modes are very similar to ones reported for Dan River Basin (Thayer, 1967, p. 116) and Deep River Basin (Reinemund, 1955, p. 59; Hermes, 1964, p. 1723), and strongly suggest a common origin for all North Carolina Triassic dolerite dikes.

STRUCTURE

Triassic strata strike from north to northeast and dip to the northwest at approximately 40°. The Triassic-Piedmont contact on the east side of the basin is a normal fault striking northeast and dipping approximately 70° to the northwest (Figure 2). Other normal faults have been mapped along the basin margin where the contact is extremely straight, where conglomerate is present, and where bedding is abruptly truncated. Dolerite dikes are believed to follow cross faults and possibly joints.

Total thickness of Triassic strata is impossible to determine accurately because of poor exposures, lack of key beds, and lack of precise knowledge concerning the effects of cross faulting. LeGrand (1954, p. 46) guessed that they might be less than 200 feet thick because of the small size and irregularity of the area. A gravity survey of the

Table 2. Modal Analyses of Dolerite from Davie County Basin.

Sample No.	6	7
Mineral		
clinopyroxene	29.4%	29.8%
plagioclase	39.7	42.0
olivine	22.4	16.8
opaque minerals	4.1	6.6
alteration	4.4	4.8
Total	100.0	100.0

area by the writer and Wilburt Geddes of the University of North Carolina may help resolve this problem.

RELATION WITH DAN RIVER BASIN

Brown (1932, p. 528) and LeGrand (1954, p. 46) suggested that Davie County Basin and Dan River Basin were once connected and that postdepositional faulting and erosion account for the present configuration of two separate basins. Field relations that support this concept are (1) bedding in each basin dips steeply to the northwest at approximately 40°, (2) similar distribution of lithofacies away from border faults in both basins, (3) Davie County Basin lies in a direct line of strike with Dan River Basin, and (4) the presence of a small outlier of Triassic rocks in Iredell County along the same strike line formed by connecting Dan River and Davie County Basins.

The absence of lacustrine facies in Davie County Basin, even though present in Dan River Basin (Thayer, 1967), could be the result of postdepositional erosion or lack of favorable depositional sites during time of sediment accumulation. Lithofacies distribution and paleocurrent indicators in Dan River Basin show a northeast trending paleoslope with the deepest part of the basin near the Virginia-North Carolina border (Thayer, 1967, p. 111). If both basins were once connected this would indicate that Davie County area was at a higher elevation (with drainage in the central part of the basin flowing towards the northeast) than Dan River Basin so that sites favorable for lake formation possibly did not exist.

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RADIOCARBON DATES OF PEAT FROM OKEFENOKEE SWAMP, GEORGIA

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ABSTRACT

Peat samples collected from the eastern margin of Okefenokee Swamp, Georgia, yielded radiocarbon dates of 5140 $\pm$ 417 and 5260 $\pm$ 360 years B. P. The dates are the first reported for this area of the swamp and are interpreted as marking the beginning of deposition of organic matter in the eastern portion of the swamp.

INTRODUCTION

In the course of pollen investigations in the Okefenokee Swamp of southeast Georgia by the author, peat samples were routinely dated using radiocarbon analysis. The dates herein reported are from peat samples collected in the eastern portion of the swamp and are the first such dates for the area. Although pollen analysis of the sediments is incomplete, analysis of the dated levels indicates a vegetation composed predominantly of elements of the Pine - Oak - Hickory Association.

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DISCUSSION

The peat material submitted for radiocarbon dating was collected from two localities by the author with a Davis peat sampler from the bottom level of Okefenokee Swamp in Chase Prairie, Charlton County, Georgia, near the eastern margin of the swamp. In both localities there was a sharp contrast between the earliest organic deposits and the sandy bottom sediments. The core is described in Table 1.

Table 1. Okefenokee Swamp Peat Core.

Peat cores GSPC 11 and GSPC 12 are nearly identical, therefore, only GSPC 12 is described in detail.

Depth from surface (cm)	Thickness (cm)	Sample Description
0-60	60	Fresh, greenish white organic plant debris composed predominantly of Nymphaeaceae and Pontideriaceae remains, abundant sponge spicules and diatoms.
120	60	Fibrous decomposed organic remains, brown organic mud matrix.
180	60	Peat and organic black mud, some silt, abundant sponge spicules.
332	152	Peat and organic black mud, no microfossils, abundant poorly preserved seed types.
362*	30	Organic black clay, peaty near the top, abundant sponge spicules and Trachelomonas tests.
---------------------DISCONFORMITY-----------------------		
362+	?	Clean, gray-white, well-rounded quartz sand, unfossiliferous.

*Radiocarbon dated sample.

The samples thus obtained were submitted to the Research Laboratory of Mobil Oil Company, Dallas, Texas, for Carbon -14 determination. The samples were treated with HCl to dissolve all carbonate particles and thus eliminate carbon contamination from that source. The organics were then converted to CO_2 by combustion and the gas passed through a microcounter. Results are shown in Table 2. The dates are enterpreted as marking the beginning of deposition of organic matter in the area. Pollen analysis presently underway on these peat cores indicates that during this 5200 - yr. period in this part of the swamp the water level fluctuated periodically with the present mean water depth approximately 2 m. Also, during this period,

Table 2

Radiocarbon No.	Sample Location*	Depth (cm.)	Radiocarbon Age (Years B. P.)
GSPC 11 (SM 1179)**	290, 500 East 487, 100 North	300-330	5140 ± 417
GSPC 12 (SM 1180)**	300, 000 East 480, 300 North	330-360	5260 ± 360

*Locations based on Georgia Coordinate System, East Zone.
**SM = Mobil Oil Laboratory Number.

the pollen succession shows a change from an oak dominated to a pine dominated vegetation. This change appears to be a regional phenomenon in southern Georgia and Florida and perhaps more widely in the southeast. According to Watts (1969) its climatic significance is unknown. At the present time the forest distribution is predominantly pine - oak - hickory - with bald cypress (Taxodium distichum) in areas of standing water.

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STRUCTURAL FEATURES OF THE COASTAL PLAIN OF GEORGIA:

A CORRECTION

By

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Robert Vorhis has kindly called my attention to an omission from the paper of the above title published in a recent issue of this journal; it is important that a correction be made, as one intent of the original paper was to gather together all of the terminology used for various features and use those which applied and held priority.

The name Charlton High was proposed for an area in southeastern Georgia where Miocene rocks rested upon Eocene rocks, the Oligocene rocks being absent. It now seems wise to abandon this name in favor of one proposed earlier, this being Orange Island--Vaughan, 1910. The name appeared as shown, but the name was not placed on a map until 1966 by MacNeil. The name, while a geomorphic term, was clearly used in a structural sense by Vaughan and later workers.

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