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Abstract

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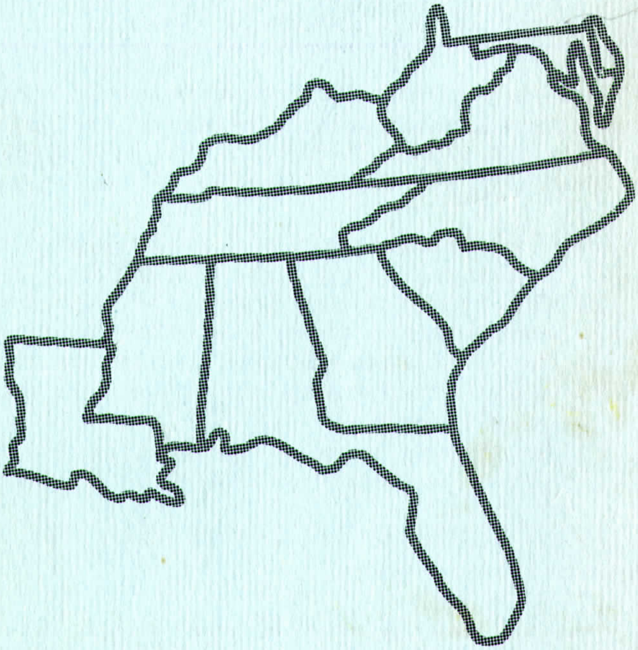
SOUTHEASTERN GEOLOGY



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SOUTHEASTERN GEOLOGY

Table of Contents

Vol. 8, No. 2

1967

CHLORITOID FROM ORANGE COUNTY, NORTH CAROLINA

by

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ABSTRACT

A select chloritoid occurrence from the Duke University quarry, Hillsborough, North Carolina is discussed. Although a continuous gradational deposition sequence exists, it is discussed for clarity as consisting of three types: 1) chloritoid in rocks with no fractures, 2) chloritoid in rocks with fractures, and 3) chloritoid in vein fillings.

The chemical components of some of the metavolcanic-metasedimentary rocks of the Carolina slate belt are plotted against the chemical component concentration field for chloritoid bearing rocks as ascertained by Seki (1954) and Halferdahl (1961). A large percent of the analyzed slate belt rocks are similar in chemical composition to the chloritoid bearing rocks.

Chloritoid forming constituents were considered to have been mobile and formation of chloritoid took place during but probably late in the metamorphic history of the area. Conditions of high aluminum and iron concentration existed and possibly high oxygen pressures.

INTRODUCTION

Chloritoid ($H_2FeAl_2SiO_7$) generally shows little variation in its chemical composition and thereby becomes an important potential mineral in metamorphic considerations. Halferdahl (1961) has given calculated upper variation substitution limits which can occur within the chloritoid structure as:

$Mg \rightarrow Fe^{++}$ 42%
$Mn \rightarrow Fe^{++}$ 17%
$Fe^{+++} \rightarrow Al$ 14%

Seki (1954) considers that chloritoid forms under conditions of low temperature and high pressure though (1957) he feels that temperature and/or chemical composition are more critical than variations of pressure in the formative process. He concurs with other authors, whose work he has summarized, that rocks in which chloritoid forms are generally high in Al_2O_3, FeO, Fe_2O_3, and H_2O and are low in their content of CaO, alkalies and MgO.

Genth (1873), Stuckey (1928), Broadhurst and Council (1953), Zen (1961), Conley (1962), and Butler (1963) are some of the authors who have either noted or discussed the occurrence of chloritoid in rocks from various locations in the Carolina slate belt. This report will not attempt to deal with these occurrences as a whole but will, rather, deal with a single select occurrence from this metamorphosed volcanic and sedimentary sequence of rocks.

GENERAL GEOLOGY

The Carolina slate belt is a northeast-southwest trending belt of metavolcanic, metasedimentary and included igneous intrusive rocks which crop out in south-central Virginia, pass through central North Carolina and South Carolina and end in central Georgia.

In North Carolina the Carolina slate belt contacts the gneisses, schists and igneous plutons of the Charlotte belt and Inner Piedmont belt to the west. To the east it is overlain by the Coastal Plain sediments and is itself split into an eastern and western section by the Deep River-Wadesboro Triassic Basin (Figure 1, and North Carolina Geologic Map, 1958). Plutonic intrusive bodies, ranging from granitic to gabbroic in composition are scattered throughout the slate belt. The northeastern edge of the belt has been intruded by large acidic plutons (Parker and Broadhurst, 1959), but otherwise consists of schist and gneiss. Most of the rocks of the western portion of the belt in which the chloritoid occurrence described here is located show low-range regional metamorphic effects with mineral assemblages typical of the chlorite zone. Locally the assemblages may change at the contact of intrusive plutonic bodies.

The term Carolina "slate belt" was first used by Nitze and Hanna (1896) and though they found rocks in the sequence with true slaty cleavage they also recognized volcanic origin for some of the materials. The major rock types of the Carolina slate belt are not slates and the term is, therefore, a misnomer perpetrated by continued common usage.

Conley and Bain (1965) recognized both felsic and mafic rocks, including lithic and crystal tuffs, welded flow tuffs, volcanic breccias, volcanic conglomerates, argillites, slates and phyllites as the rock types in the Carolina slate belt west of the Deep River-Wadesboro Triassic Basin in North Carolina.

Major structural trends for the belt are generally aligned in a northeast-southwest direction. In the study area under consideration the rocks assume an attitude that is nearly vertical and strike approximately north 35° east. Cleavage parallel to the bedding plane is predominant and quite prominent especially on bedding planes and in the schist and phyllite rock contingent.

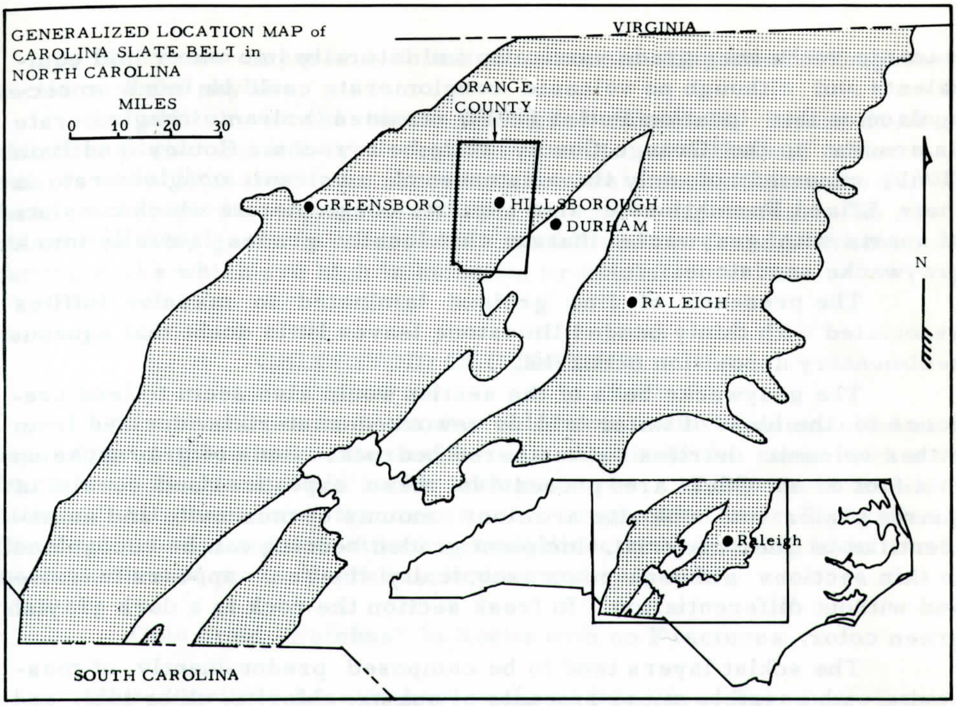


Figure 1. Index and location map showing the Carolina slate belt and the area under consideration in this report.

Removal of building stone from the Duke University quarries (Figure 1), which are located directly west of Hillsborough, North Carolina, has exposed what Conley and Bain (1965) consider to be the most typical part of their Efland Formation. A stratigraphic section of relatively unweathered rock material up to one hundred feet thick has been discontinuously exposed for a quarter of a mile along strike in the quarries. The third dimensional component of exposure on one quarry face is as much as one hundred feet.

Rock units in these exposures consist predominantly of tuff breccia and volcanic breccia with interbedded layers of muscovite schist, chlorite or muscovite quartz phyllite and slate, graywacke, and laminated to semi-massive tuffite.

Although tuff breccia and volcanic breccia probably predominate as the abundant rock type of these exposures, all gradations of tuffaceous rocks are present and in many cases one will grade imperceptibly into another. The variations include the lapilli tuffs, lithic tuffs and lithic-crystal tuffs. They are usually light gray to dark gray, green, or blue in color. Lithic clasts may be lighter in color than the groundmass but are usually darker and are composed of fine grained volcanic rock fragments which are usually flattened and sheared giving the parent rock a semi-planar structural appearance. These tuffaceous volcanic rocks may grade vertically and laterally into water laid equivalents and although no volcanic conglomerate could be found or recognized at this location Butler (1963) reported volcanic conglomerate as present in the Orange County slate belt rocks. Conley and Bain (1965) reported not only the existence of a volcanic conglomerate in their Efland Formation but also a quartz conglomerate which consists of quartz with graywacke matrix that locally grades laterally into a graywacke sandstone.

The presence of fine grained laminated to massive tuffites associated with thinly bedded limestone leaves little doubt that aqueous sedimentary deposition occurred.

The graywacke beds of the section would also seem to lend credence to the idea of water laid or reworked materials derived from either volcanic detritus or weathered bedrock. Beds of graywacke up to a foot or so thick are present in these exposures and consist of quartz grains and chlorite or minor amounts of muscovite and an unidentifiable finer material. Incipient graded bedding can be recognized in thin sections although macroscopically the rock appears massive and without differentiation. In fresh section the rock is a dark grayish green color.

The schist layers tend to be composed predominantly of muscovite with possibly minor amounts of quartz, chlorite, chloritoid, and hematite. The color is usually light gray and the grain size of the mica is from microscopic to nearly a quarter inch across, the last creating an extremely coarse texture in the rock.

X-radiation and thin section studies of the phyllite show the rock to be composed of a chlorite-quartz or chlorite-quartz-muscovite mineral assemblage with or without hematite, chloritoid, and leucoxene. In thin section the host rock is composed of irregular lenses and layers of various sizes of angular, elongate quartz grains aligned parallel to the rock bedding cleavage. These lenses or layers alternate with and may be enclosed in a felty mosaic of fine grained muscovite and chlorite. The muscovite is dispersed rather evenly throughout but the chlorite appears in lenses parallel to the rock bedding. Up to one eighth or more of the rock may consist of large irregular blebs and/or fine flakes of hematite rather evenly but very definitely aligned with the bedding. The contrasting dark color of the hematite darkens the rock and accentuates the visible schistose texture of the rock. Butler (1963) reported opaque minerals which included hematite, magnetite, pyrite, and leucoxene, in the mineral assemblages of the argillite, slate, and phyllite from the Orange County, North Carolina slate belt rocks. McCauley (1961) reported a rather extensively developed quartz-sericite phyllite (Sample 4 [65-N-1]) in the southeastern part of Newberry County, South Carolina slate belt which had a 17.64% Fe_2O_3 content. He reports that the hematite shows no replacement relationships with the quartz or mica and considers it to be primary.

Leucoxene is present in minor amounts dispersed throughout sections of the phyllite.

In addition to the above discussed minerals, layers of Al-chlorite, planar rosette-type crystals of white wavellite and minor amounts of zeolites occur as secondary minerals deposited on the surfaces of the fracture systems within the rock exposures. The Al-chlorite is deep green in color, well crystallized and found in fractures cutting rocks which are high in hematite or chlorite.

CHLORITOID OCCURRENCE

For convenience and clarity chloritoid will be considered as occurring in three separate modes: 1) chloritoid in rocks with no fractures, 2) chloritoid in rocks with fractures, 3) chloritoid in vein fillings. All three are the result of a single total overall physicochemical process and are either related directly by grading from one into the other of more discretely in time and space.

Chloritoid "Patches" in Rocks with no Fractures

The chloritoid-bearing host rocks of this group have been least affected by regional or local movements so that schistosity and slaty cleavage or shearing has not occurred in them as it has in the associated schists and phyllites. These host rocks are the more competent local portions of the volcanic breccias, lithic tuffs, and lithic crystal tuffs.

When chloritoid is found in these rocks the crystals are unoriented, very small single crystals or rosettes and constitute only an extremely small fraction of the overall rock volume. The chloritoid tends to be localized in small "patches" within the host rock and is apparently not related to fractures. The general pattern for these "patch" occurrences is random and seems to follow no recognizable system of placement within the host rock, although they are probably associated with local differences in composition or competence which previously existed or became significant during orogeny of the host rock.

The mineral assemblage of the "patch" may be only quartz, mica and chloritoid or it may also contain chlorite. The area containing chloritoid is usually lighter than surrounding rock and as the chloritoid to chlorite ratio increases the difference becomes quite marked, especially at the center of the "patch" where the chloritoid crystals or crystal aggregates become relatively large. Thin section study does not reveal replacement textures between chlorite and chloritoid. Under these circumstances it is difficult, if not impossible, to say whether the chloritoid formed from the chlorite or whether it formed contemporaneously because of a local variation of physicochemical condition which existed in the host rock at the time of formation. These "patches" of chloritoid have not been found in rocks with a high hematite content.

Chloritoid Segregations Related to Fractures

This group includes the bulk of the chloritoid deposition. Within this group, chloritoid occurs in schists, phyllites, "slates" and next to fractures or fracture systems in these or other rock types.

Chloritoid porphyroblasts occur in complete random orientation within the rocks of this group and though in most instances the position of the chloritoid crystal could be seen to be related to the fractures created by the dynamic displacement in the host rock, its crystallographic orientation appeared completely unrelated to lineation and foliation of associated minerals.

Size of the chloritoid crystal or crystal rosette porphyoblasts appears to be directly related to crystal size of the surrounding minerals and an indirect function of the dynamometamorphism which occurred in the host rock. In exceptionally coarse-grained mica schists individual chloritoid porphyroblasts up to a quarter inch in size were found. In these coarse grained schists the effect of rotation on chloritoid porphyroblasts was quite evident. It became less evident, however, as grain size and the effect of dynamic displacement decreased. In chloritoid containing phyllites and "slates" rotation appeared to be almost or wholly lacking.

The above discussion holds only for rocks of this study where displacement is an intregal part of the rock fabric. Where a secondary fracture or a fracture that is unrelated to the schistose rock fabric occurs the situation is quite different. As these fractures are scrutinized in cross-section a distinct weakening or complete loss of parallel alignment of minerals may occur as the fracture is approached. This zone of obscure schistose texture was noted to be as much as six inches wide on either side of the fracture and in this zone gave the rock a massive appearance. Where chlorite is normally a constituent of the host rock it tends to be relatively less abundant in this zone and its place is taken by quartz, sericite and chloritoid. This change in the mineral assemblage tends to lighten the background color of the host rock next to the vein. There is also a general increase in chloritoid crystal size near the fracture that would indicate mobility and migration of chloritoid constituents but it does not necessarily reflect an increase in the number of nuclei.

Chloritoid-bearing Vein Fillings Formed by Lateral Secretion

Veins constituting this classification contain chloritoid and fill the dilatantly formed openings on bedding plane cleavages. Those observed are sheet-like in form. They are up to two inches thick and tens of feet in lateral extent, forming from and grading gradually into the previously discussed type. Away from these gradational zones their contact with the host rock is sharp and there is evidence for little if any reaction between vein and host rock during the time of emplacement and crystallization of the vein-filling material. Although the

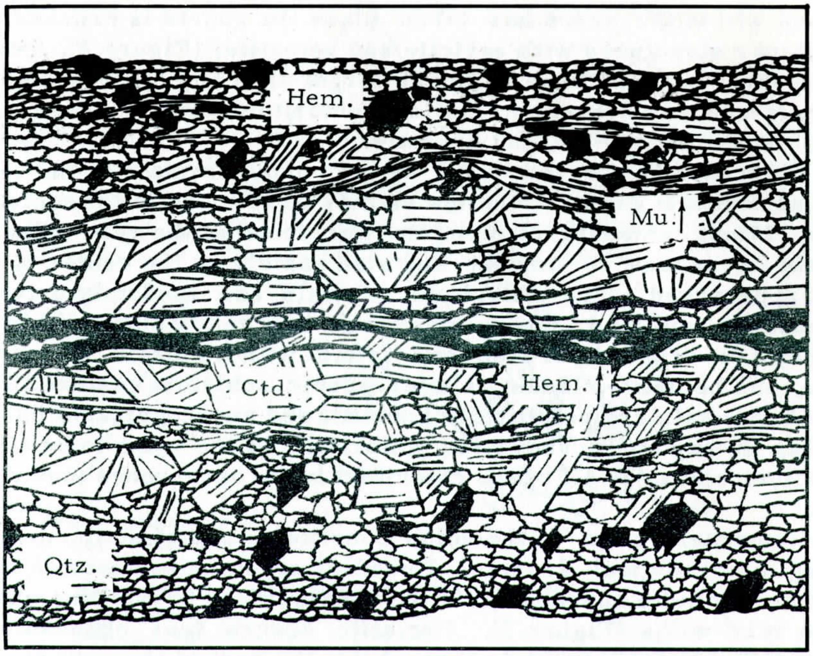


Figure 2. Drawing of section across quartz-chloritoid-hematite-muscovite bearing vein. Ctd-chloritoid, Qtz-quartz, Hem-hematite, Mu-muscovite, ---- trend of cleavage.

chloritoid-bearing secretion veins were observed to occur only in phyllites, it is possible that they may occur in other types of rocks with which the phyllites are interlayered.

The crystalline nature, superimposed dynamometamorphic cleavage, parallel to the bedding plane cleavage and high content of opaque mineral near the center of the vein are immediately visible in hand specimen. A hand lens further reveals a general gradation of crystal size from edge to center and colliform banding of hematite at the vein center (Figure 2).

Microscopic and X-ray analysis show a mineral assemblage of quartz, chloritoid, and hematite with minor amounts of sericite and chlorite. This assemblage, except for changed ratios and presence of chloritoid in the vein, is the same as that of the enclosing phyllite.

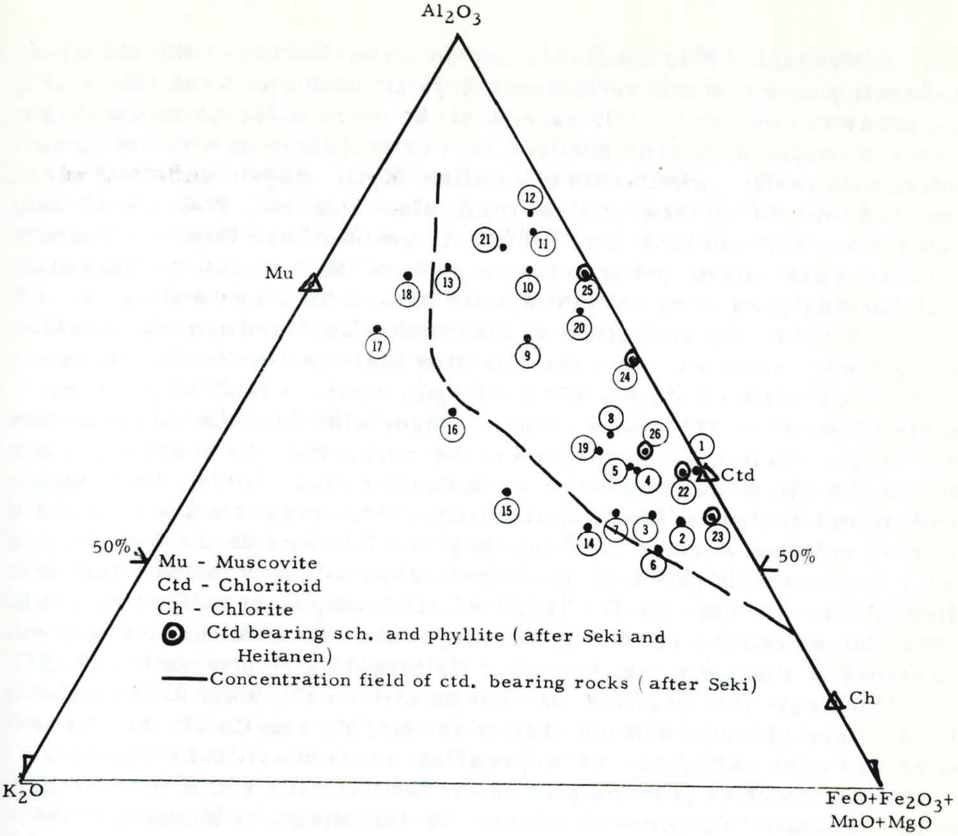
Mineralogy of the Chloritoid-bearing Veins: Quartz is the matrix-forming mineral of the chloritoid-bearing veins and occurs throughout them as probably the predominent mineral in most sections. Size of the quartz grains varies considerably across and throughout the veins, although a generalization can be made that the vein is fine-grained at the edge and coarsens toward the center. Individual grains are elongate parallel to vein walls and of xenoblastic structure. Where movement within the veins has taken place the quartz is crushed and forms planar structures with sericite and hematite, (Figure 2). Grains of strain-free quartz that are much larger than any other within the field occur in areas between crystals of chloritoid that are close together and in strain shadows of the crystals.

Chloritoid has crystallized as individual crystals in the veins and not as multiple crystal rosettes as may be the case in the schist or phyllite. Minor amounts of chlorite are found at the vein walls but chloritoid increases in amount and size rapidly as the center of the vein is approached. The chloritoid crystals are randomly oriented throughout the vein and, where movement has taken place within the vein, they are crushed, broken and felted with a mosaic of quartz, sericite, and hematite, lending a microscopic schistose texture to the vein. All crystals are poikilitic of hematite and unrecognizable opaque minerals. X-ray study shows the chloritoid of the veins to be a triclinic polymorph as is all of the chloritoid in the occurrence under discussion.

Hematite content of the veins constitutes one of their more interesting mineralogical aspects. Euhedral crystals of rhombic morphology occur quite abundantly in the fine-grained crystalline quartz near the vein walls (Figure 2). Hematite occurs less abundantly in euhedral to subhedral crystals throughout the rest of the vein section and much of this has been crushed and spread out by dynamic processes. Hematite also occurs as minute inclusions in all of the chloritoid crystals. At the center of most veins (Figure 2), in either continuous or lens shaped bodies, hematite also occurs as a crustification of colliform structure over the quartz and chloritoid. The colliform structure outlines or fills the openings formed either by movement within the vein or unfilled sections of a continuous vein-filling process. The crystals are radially arranged with their long axis normal to the free surface of crystallization and the colliform structure outlines the crystal form of quartz or chloritoid over which is is deposited. In most cases there has been no destruction of this hematite layer by movement and it probably was the last depositional layer to be formed within the chloritoid-bearing veins.

CHEMICAL COMPOSITION OF SOME CHLORITOID-BEARING AND CAROLINA SLATE BELT ROCKS

Seki (1954) concluded from his own and previous work on chloritoid-bearing rocks that they are restricted in chemical composition and are characterized by a richness in Al_2O_3, FeO, Fe_2O_3, and H_2O, and a poverty in CaO, Na_2O, K_2O, and MgO. Further, they vary widely in their range of SiO_2 content. This variation may range from 40 to 90 percent.



Sample Identification

1. Vitric tuff, Albermarle Co., N. C., (Conley, 1962).
2. Silicified argillite, Orange Co., N. C., (Butler, 1964).
3. Amygdaloidal greenstone, Orange Co., N. C., (Butler, 1964).
4. Argillite, Albermarle Co., N. C., (Conley, 1962).
5. Amygdaloidal basalt, Albermarle Co., N. C., (Conley, 1962).
6. Phyllite schist, Newberry Co., S. C., (McCauley, 1961).
7. Volcanic breccia, Orange Co., N. C., (Butler, 1964).
8. Mafic lithic-crystal tuff, Albermarle Co., N. C., (Conley, 1962).
9. Felsic tuffaceous argillite, Albermarle Co., N. C., (Conley, 1962).
10. Mafic tuffaceous argillite, Albermarle Co., N. C., (Conley, 1962).
11. Felsic tuff, Albermarle Co., N. C., (Conley, 1962).
12. Graywacke, Albermarle Co., N. C., (Conley, 1962).
13. Welded felsic flow tuff, Albermarle Co., N. C., (Conley, 1962).
14. Vitric-crystal tuff, Orange Co., N. C., (Butler, 1964).
15. Quartz-mica schist, Newberry Co., S. C., (McCauley, 1961).
16. Phyllite, Orange Co., N. C., (Butler, 1964).
17. Porphyritic rhyolite, Albermarle Co., N. C., (Conley, 1962).
18. Porphyritic rhyolite, Albermarle Co., N. C., (Conley, 1962).
19. Avg. of 23 analyses of graywacke, (Pettijohn, 1957).
20. "Typical, banded, dark slate," Wet Creek, Moore Co., N. C., (Stuckey, 1928)
21. "Slate", Haile Mine, S. C., (Nitze and Hanna, 1896).
22. Chloritoid rich part of phyllite, Uchiyama, Masuzawa-mura, Japan, (Seki, 1954).
23. Chloritoid phyllite, Uchiyama, Masuzawa-mura, Japan, (Seki, 1954).
24. Chloritoid phyllite, Shiraito-no-taki, Yahagi-mura, Japan, (Seki, 1954).
25. Sericite-chlorite-chloritoid vein, Okuzu, Otomo-mura, Japan, (Seki, 1954).
26. Muscovite-chlorite-chloritoid schist, Rawlensville Co., Pa., (Hietanen, 1954).

Figure 3. Al_2O_3-(FeO + Fe_2O_3 + MgO + MnO)-K_2O diagram showing relation of analysis of rocks from the slate belt of North Carolina and South Carolina to the field of chloritoid-bearing rocks and selected samples of non-chloritoid-bearing rocks.

Halferdahl (1961) concluded that in three-fourths of the chloritoid-bearing rocks he analyzed alumina is about one to three times as abundant as the mafic constituents, that both are more abundant than the sum of potash plus soda plus calcium or that there is slightly more alumina than mafic constituents even after K_2O, Na_2O, and CaO are calculated out in representative minerals. Further, FeO + MnO is greater than MgO and that FeO + MnO is greater than Fe_2O_3. There is a close correlation between the results of Seki and Halferdahl although the analyses of Halferdahl appear to be a bit more mafic.

Data from the literature on chemically analyzed non-chloritoid-bearing "type" samples from the Carolina slate belt were plotted on a triangular diagram of Al_2O_3-(FeO + Fe_2O_3 + MgO + MnO)-K_2O components (Figure 3). This was done to ascertain the relationships, if any, between chemical composition of the rocks and the concentration field for chloritoid-bearing rocks as outlined by Seki (1954). Of twenty one chemical analyses fifteen plotted within this concentration field and five were relatively close although they did fall outside the boundary. When this remaining group of five are compared with the diagram of Halferdahl and a total (K_2O + Na_2O + CaO) component instead of only K_2O is considered for chloritoid bearing rocks three of the five would be included in the more mafic area of Halferdahl's concentration field.

Although the position of the points for the ideal formulas of chloritoid and chlorite will not shift if the Na, K, and Ca of the Carolina rocks were calculated as appropriate alumino-silicate minerals, the total composition point as plotted for the Carolina rocks would shift toward the mafic constituent corner of the diagram because of the alumina used. This shift would not only closely approach the chloritoid composition position but would also come much closer to the composition point of the Al-rich chlorites found in these rocks. The composition point of the Al-rich chlorite lies above and toward the Al_2O_3 corner of the diagram from the ideal chlorite composition point. This point was expressed by Seki's diagram as lying at the most mafic point of what he considered to be the chloritoid concentration field (Figure 3).

The chemical content and relative ratios for chloritoid-bearing rocks may be partially or almost wholly due to the chloritoid and chlorite content of the rocks. This is true whether the origin of the chloritoid was the result of wholesale hydrothermal introduction or the result of readjustment of the original constituents of the rock through local metasomatic action.

DISCUSSION

It is interesting to note that when the chemical analyses from a number of rock types from the Carolina slate belt are plotted on a triangular diagram Al_2O_3-(FeO + Fe_2O_3 + MgO + MnO)-K_2O, the resulting points fall predominantly within the concentration pattern for chloritoid-bearing rocks as outlined by Seki (1954) and Halferdahl (1961). If, however, the Al_2O_3-MgO-(F_2O_3 + Fe_2O_3) ratios are plotted, then samples 2, 3, 5, and 8 would fall outside the concentration field, but this would still leave over half the total plotted analyses within the chemical limits of chloritoid-bearing rocks. Though the complete mineralogy of the slate belt rocks west of the Deep River-Wadesboro Triassic Basin has not been carried out, it is possible that chloritoid is much more abundant in these rocks than is now suspected. The isolated and differentially selective occurrence of the chloritoids of this study would indicate, however, that more than similar chemical composition of the rocks is involved.

Seki (1954) concluded that the formation of chloritoid was always related with "violent hydrothermal action." He considered this necessary because of the low temperature involved and the resulting slow chemical reactions. This view is quite understandable in the production of chloritoid in rocks where the primary role of the solution is to transport material to the site of deposition. Where the chemical composition of the host rock is already that of a chloritoid-bearing rock, as it is in many of the Carolina slate belt rocks, the role of solutions may not be that of transport, but may be only that of a diffusion medium. This is not to deny the presence or activity of a solution but rather to point out that under these circumstances a high temperature hydrothermal solution with much movement is not necessary.

Chloritoid occurrence of Type 1 discussed above might be considered at this point. In this instance no visible or inferable fracture system is, or has been associated with the area containing chloritoid, yet this "patch" contains the same mineral assemblage as those occurrences associated with visible fracture systems in which fluids could move. This mode of occurrence could not be the result of intense hydrothermal activity. The author considers the formation of chloritoid under these circumstances to be the combined results of appropriate initial local chemical composition combined with other factors such as differential competence of the host rocks and solution acting in a local closed system.

No petrographic evidence exists, in this instance, to indicate that chloritoid replaced chlorite. The variation in size from small at the periphery to large at the center of the "patch" for chloritoid crystals or crystal aggregates would indicate that a transport gradient was active and that the chloritoid constituents were mobile within the confines of the patch whether their origin was previous to, contemporaneous with, or were derived from the associated chlorite. As early as 1881 Gosselet concluded that chloritoid formed next to faults by dynamometamorphism and in rocks of proper composition. He termed the process by which chloritoid formed in the latter as "static metamorphism."

In consideration of chloritoid that formed in the phyllites and schists of this deposit it can be said that crystallization of chloritoid is related directly to planes of cleavage within the host rock. As stated previously, however, these chloritoid porphyroblasts are related to the rock cleavage in position only and not in crystallographic orientation. There is no doubt that this cleavage acted as direct avenues for movement of the chloritoid constituents. With a parent rock material containing all necessary constituents for chloritoid formation the transporting solutions could maintain short gradients to the crystallizing centers with very little movement being necessary.

As a fracture is approached the schistose texture of the host rock is erased, crystals of chloritoid become generally larger and material has been mobile and moved over longer distances. Mobility and extended material movement is substantiated in pure vein fillings where the material must have moved over tens of feet from its origin of position in the wall rocks to its crystallization position in the chloritoid-quartz-hematite vein.

Most notable about the vein filling is its stable content of chloritoid and hematite and the lack of magnetite. Not only does hematite occur near the edges and throughout the veins in euhedral to crushed subhedral grains but it also occurs as crustification at the center of the vein on quartz and chloritoid. The chloritoid occurs adjacent to hematite and with poikilitic inclusions of hematite within the section. Halferdahl (1961) reasoned that because chloritoid is a ferrous aluminum silicate it would not be expected to persist much, if any, beyond the oxygen pressures corresponding to the equilibrium between hematite and magnetite, except for the presence of Fe^{+3} and Mn in the chloritoid structure. This substitution would be limited as shown by Halferdahl (1961). The complete absence of magnetite and presence of the hematite-chloritoid assemblage in the vein would indicate that chloritoid may possibly form and persist at higher pressures of oxygen than is indicated by the hematite equilibrium.

During the growth of the vein movement took place within the vein and parallel to its walls. This movement is evidenced by the crushed grains, the formation of sericite and the resultant schistose texture. It is evident from this that stress existed in the vein during periods of its growth. The undamaged hematite deposition on unstrained quartz and chloritoid crystals near the center of the veins would also indicate that there were periods of time during deposition when only fluid pressures were active. The growth of chloritoid in these veins could, therefore, have occurred under both stress and stress-free conditions.

The chloritoid-bearing veins occur in phyllite and there is little or no evidence of reaction at the vein-wallrock interface. Absence of stress within the vein during its formation would not have permitted simultaneous development of enclosing phyllite. That the vein was not completely stress-free during its whole formational history would show, however, that it was partially involved during the dynamometamorphic development of the surrounding rocks to the chlorite-muscovite subfacies level.

If, however, the time factor is considered in light of the coarse grained schist two factors become quite evident. The schistose textural development, namely shear and cleavage, was instrumental in development of the chloritoid crystals in this rock and rotation of chloritoid crystals took place during their growth. Both of these factors would indicate that growth of the chloritoid crystals took place at the same time as dynamometamorphic forces were active in the host schist.

It would appear, therefore, that the chloritoid of this deposit did form late in the metamorphic history of the host rocks of the area.

CONCLUSIONS

The chloritoid of this deposit crystallized as the triclinic polymorph under both stress and non-stress conditions. It occurs in the chlorite-muscovite subfacies of the green schist facies metamorphosed under both static and dynamic metamorphic conditions. Temperatures below the biotite isograd prevailed.

Many non-chloritoid rocks of the Carolina slate belt have a chemical composition similar to that of chloritoid-bearing rocks. Where this chemical equality exists it is assumed that intense hydrothermal activity is not necessary to form chloritoid but that solutions derived during and through normal dynamometamorphic processes are capable of obtaining and affecting necessary local movement of chloritoid constituents to their sites of crystallization. The chloritoid constiuents are mobile as indicated by laterally secreted vein fillings.

Oxygen pressures above the hematite-magnetite equilibrium point may exist during chloritoid formation. High aluminum and iron values are necessary though the relationship of low magnesium and calcium values is not clear. This may be a simple reflection of low aluminum content in high calcium and magnesium-bearing rocks.

Formation of the chloritoid took place during but probably late in the metamorphic history of the formation of phyllite and schistose rocks of the area.

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THE RIPPLE MARK ANALOG: PRELIMINARY RESULTS

by

William F. Tanner

ABSTRACT

The spool-type analog, which produces shear ridges in loose, moist sand, can be manipulated so that various effects, which are inseparable under water, can be simulated individually. Results of such work show that, with the model, ridge spacing is related directly to grain diameter and efficiency of packing, and inversely to sediment sorting and departure from sphericity. Factors which were not incorporated in the model, but which might be involved, include vertical velocity gradient, presence of clay or organic matter in the sediment to act as "glue", water-wave energy level, bottom slope, water depth, sediment skewness and sediment polymodality.

With the model, ridges are produced as a result of not-quite-perfect transmission of shear-induced stress through those grains which lie in front of the advancing spool. Stress transmission failure is realized at points where individual grains, or clusters of grains, are squeezed up out of the bed. Each small heap triggers a skipping motion of the spool, which then begins all over again to build up a shear on the sand bed. The same mechanism, under water, operates as a result of shear or drag imparted by wave orbital motion or water current action on the sand bottom. Under water, however, eddies develop between the shear ridges, building the latter up to greater heights; this effect is not reproduced by the spool.

THE MODEL

Ripple marks under water are initiated by shearing at the bed surface and are maintained by eddy effects. The shear mechanism can be studied conveniently in the laboratory by means of a ripple mark analog model which does not, however, provide the build-up and maintenance resulting from eddies (Tanner, 1963).

The model consists of a one-piece wooden spool, having two wheels of different diameter (Figure 1). Each wheel has a flat edge on which it rolls. The model would follow a curved path if rolled on a flat, smooth surface. The larger wheel, however, is guided along a straight track. As a result of this guidance (inasmuch as the entire spool is a single rigid piece), the smaller wheel has a tendency to either slide over the surface or to drag any loose material along with it. How this drag, or shear, operates can be seen by rolling the smaller wheel along a thin piece of wood which rests on ball bearings. The thin wood

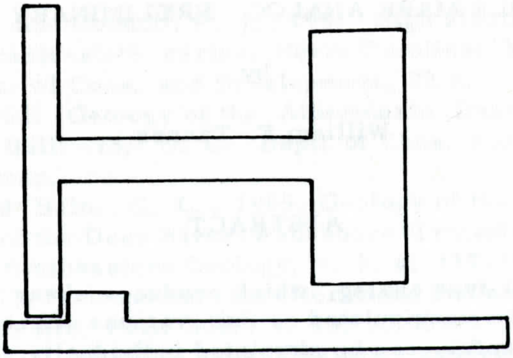

Figure 1. Diagram of spool, showing large-diameter wheel restricted by flange to straight line travel, and small-diameter wheel resting on moist sand bed.

moves a distance equal to the difference between the two wheel diameters per revolution of the wheel. Its displacement, in the direction of wheel travel, is a shear motion. Without the ball bearings, the same piece of wood is dragged along under the smaller wheel, but less easily, less smoothly, and less far. The ease with which it is dragged depends in good part on the ease with which it slips along the surface on which it rests; that is, the motion of the thin wooden plate is a function of the drag exerted from above and of the resistance from below.

Operation of the model so that the smaller wheel rolls across a single layer of exposed ball bearings produces several interesting effects. If the ball bearings are closely packed (hexagonal pattern), a model is obtained which is more interesting than the thin piece of wood. The latter is rigid, and does not deform as a result of shear applied from above: it either moves (slippage below the thin wood), or it does not move (slippage above the wood). A layer of ball bearings is not rigid, and therefore only a few particles are directly affected, at any given moment, by the small wheel. The tendency for a single bearing is to roll under the wheel. In close packing, however, single bearings are not free to roll, inasmuch as other particles, not subject to shear at the same time, are in the way. As a result of this situation, a stress is propagated through the bearings in front of the advancing wheel. Transmission of this stress may be obvious up to distances as great as 50 to 100 diameters in front of the small wheel.

When crudely-spherical particles are used (i. e., BB's or lead shot), the transmitted stress results in an arching of a small group of spheres, typically 5 to 20 diameters ahead of the wheel. This arching occurs where several grains are more efficiently packed or slightly larger than their neighbors. The site of arching can be predetermined by placing a few slightly larger particles in an otherwise uniform layer.

If the arched group is close to the advancing wheel the small arch will probably collapse downward as the wheel passes over it. If arching begins at a great distance in front of the small wheel the arch may collapse upward as the wheel approaches, forcing one or more particles above the general level of the layer surface. These results are obtained with almost perfect sorting ($\sigma = 1.00$), almost perfect sphericity, and almost perfect packing. In actual sand-sized sediments, these three conditions are not met.

A mixture of two sizes (BB's plus glass beads) produces the result that the smaller pieces are elevated, whereas the larger pieces tend to stay put. This is a general statement, which describes all experiments to date, regardless of whether the sample was skewed toward the finer or coarser sizes; it is also intuitive. Spherical particles are also more likely to be squeezed out (i. e. , up) than are non-spherical ones.

OBSERVATIONS

These experiments suggest that diameter, sorting, shape, and packing are important in controlling the response of the bed to passage of the smaller wheel. In general, the following observations have been made:

(1) Spacing between extruded clusters ("piles") is inversely related to the numerical value for sorting: with good sorting (small standard deviation) the spacing tends to be great and the ridges relatively low.

(2) Departure from sphericity tends to reduce the spacing.

(3) More efficient packing tends to increase the spacing.

(4) Spacing may be affected by particle size skewness.

(5) If all of the above effects can be minimized, spacing is a function of grain diameter. When sand is used, instead of ball bearings or beads or lead shot, details are difficult to observe, but approximately the same results are obtained. There is, however, a shear threshold which must be overcome, and this parameter also influences the spacing between piles or ridges. The shear threshold is perhaps a measure of the slippage, or inefficiency, in the system.

A damp, brightly-patterned cloth, in place of the bed of ball bearings, shows how displacement in the bed accumulates to a maximum, at which moment a ridge is built and the smaller wheel skips momentarily. After this adjustment, the build-up of shear starts over again. Fibers in the cloth change position many diameters in front of the advancing wheel, buckling up to form ridges which are then pressed into sharp creases. The cloth behind the wheel is stretched.

Loose moist sand responds in a similar manner. Individual grains are dragged a slight bit, leading to a clustering and squeezing-out (to form piles or ridges) well in advance of the wheel. The regular

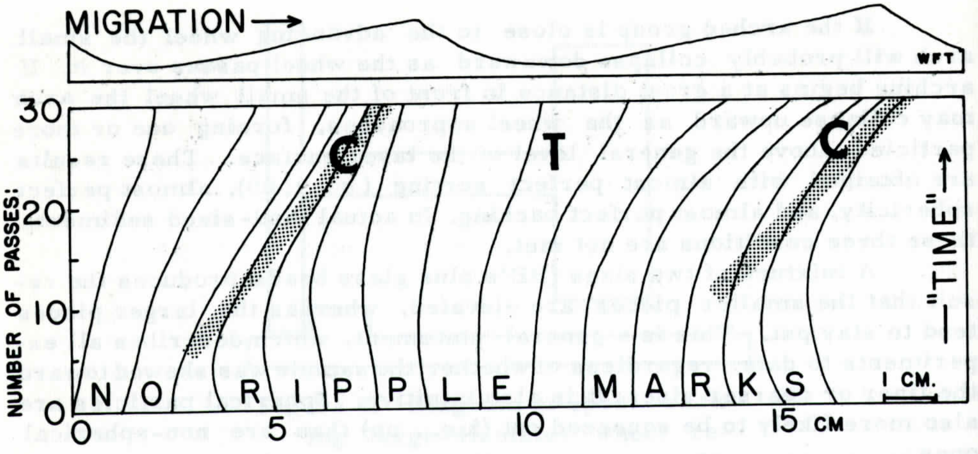

Figure 2. Chart of key grain location (spacing along horizontal axis) versus time (number of passes along vertical axis). Key grains, chosen for distinctive color and seeded in the sand bed at 1.0 cm intervals, were deformed toward the right (spool movement toward the right). Ripple mark crests (stippled strips) appeared after a few spool passes and also migrated toward the right. C = crest, T = trough. At the top is a vertical profile of the ripple marks after 30 passes.

spacing between such ridges--in uniform materials--is evidence that a systematic deformation, such as a drag-slip-drag, is at work. The fact that such ridges appear after two to four passes of the wheel is evidence that the motion of grains within the bed is relatively great when subjected to shear. A sand bed seeded with brightly colored grains can be used to study individual particle motion. If the marker grains are carefully mapped, their displacements can be followed rather accurately (Figure 2).

RESULTS

Results of studies with sand beds can be summarized as follows: (1) There appears to be a definite relationship between ridge spacing (wave length, λ , in cm) and sand grain diameter (millimeters). This can be expressed as:

$$\lambda = ag^{0.133\,\Delta}$$

for certain sieved sands and certain spools having axles differing by Δ cm from an optimum length of 18.6 cm. The difference in wheel diameters for these spools varied from a minimum of 1 mm to a maximum of 7 mm (guide wheel diameter = 7.6 cm). The constant a is probably close to a = 1.

(2) The difference in wheel diameters does not affect ridge spacing but appears to control the amplitude of the ridges. A difference in diameter of 1 mm produced very faint ridges; larger differences resulted in higher ridges. A spool having two identical wheels made no ridges at all. The height of a single ridge depends on the amount of material squeezed out of the bed during a unit time (that amount of time needed for the smaller wheel to travel between two points: a. the point from which the ridge-site is first affected, and b. the point at which the wheel passes over the ridge and hence out of range). During this time (or in this distance) the amount of grain displacement must be related to the diameter differential, with a greater differential resulting in more shear and hence more displacement of individual grains.

(3) The shear ridge crest migrates at roughly twice the average speed of individual grains.

(4) Grains on the steepest slopes move most slowly; those near the shear ridge crest but on the gentlest slopes move most rapidly (Figure 2).

(5) Wheel translation speeds of 35 cm/sec, and slower, produced no shear ridges.

(6) Wheel translation speeds of 140 cm/sec, and faster, produced no shear ridges.

(7) In general, spacing of ridges is not a function of translation speed.

(8) The ripple index (spacing divided by height) of shear ridges made by the model is typically 15:1 to 20:1.

In shallow water, shear ridges lead to the formation of eddies on the sand bottom, and these in turn build the ridges up into ordinary sub-aqueous ripple marks. When this build-up is complete, the shear mechanism which first formed the ridges has been transferred to the top surface of the eddy layer and therefore no longer shapes the sand surface. Under surface waves, the passage of three to ten orbits (i. e., wave crests) is sufficient to form shear ridges; thereafter, reversing bottom eddies shape the ripple marks. Hence the drag-slip-drag mechanism is not important for most of the history of any single ripple field.

CONCLUSIONS

In view of the fact that shear initiates, but eddies maintain, the sub-aqueous ripple mark field, not all of the work done with the analog is pertinent to the prototype. The following provisional list of conclusions, possibly suitable for transfer to the prototype, is suggested:

(a) Ripple mark spacing is directly related to grain diameter (based on items 5 and 6, in the first list above). All other things being equal, coarse sands should exhibit wider ripple mark spacing than do fine sands. Both grain diameter and spacing are measurable quantities; hence departure from the relationship stated here should be taken as numerical evidence for some effect other than particle size.

(b) Spacing is inversely related to the numerical value for sorting (standard deviation); that is, wide spacing reflects good sorting (small values for).

(c) Spacing is directly related to the efficiency of packing, and therefore inversely related to porosity; that is, wide spacing indicates efficient packing and low porosity (relatively).

(d) The heights of ripple marks built under water reflect the action of bottom eddies, rather than the initiating mechanism, and therefore probably cannot be used for studying the latter. Under air, however, true eddy-built ripple marks are not formed, and the heights of the shear ridges are related directly to the numerical value for sorting; that is, high ridges indicate poor sorting. High ridges are commonly interpreted in terms of low values for the ripple index (i.e., low for wind-blown sands, and hence in the neighborhood of 15 or more). Again, both ridge height and sorting are measurable quantities, and hence any departures obtained should be taken as significant bits of information. Wind-blown sands have ripple mark indices up to around 70, in both modern sands and lithified sandstones (e.g., more than 100 observations from Mesozoic sandstones of the Colorado Plateau area, U.S.A.; Tanner, 1966); this is generally a measure of excellent sorting.

These suggestions relate spacing and ridge height to grain size, sorting, and packing; previous statements indicate a possible relationship to sphericity and grain size skewness. However, these sediment characteristics are not the only factors which influence spacing. Spacing is also a function of effective water depth (Tanner, 1959). It is probably also related to the vertical velocity gradient (this can be expressed by altering the coefficients a and b in the equation given above). Therefore departures from the relationships stated here may be significant in terms of environmental differences, such as effective water depth. The data are not sufficiently numerous, however, to permit a specific statement to be made at this time.

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THE DIABASE OF THE BUTNER - CREEDMOOR AREA, GRANVILLE COUNTY, NORTH CAROLINA

by

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ABSTRACT

Reconnaissance mapping of residual soils and boulders of diabase reveal the presence of some large diabase units in the Butner - Creedmoor area, Granville County, North Carolina. The diabase material is found to occur in two distinctly different structural forms:

1. As narrow, more or less vertical intrusive structures called dikes.
2. As bodies of diabase material of great areal extent.

Geologic interpretation of the field data suggests a flow-like structure for much of the diabase, rather than sill-like bodies. The extrusive character of the diabase seems to be indicated in a number of exposures. Although no detailed stratigraphic studies were performed, it is felt that the intrusion and extrusion of diabase occurred during the later Upper Triassic.

INTRODUCTION

During a geologic investigation for the North Carolina State Highway Commission, extensive occurrences of diabase were encountered in the southwest corner of Granville County, North Carolina. The rather poor engineering properties of the residual diabase soils necessitated a more detailed study of this unit along the alignment of the proposed highway project. The field work was conducted at intervals in 1965/66.

Geologically, this area is located in the northern part of the Durham Triassic Basin (Figure 1).

LOCATION AND GEOGRAPHY

The area covered by reconnaissance mapping is described by the Granville County Line to the south and west; to the north by the NE-SW trending contact between Triassic sediments and the rocks of the Piedmont Province; the eastern boundary of the mapped area follows a N-S line, one to two miles east of NC 50 and US 15.

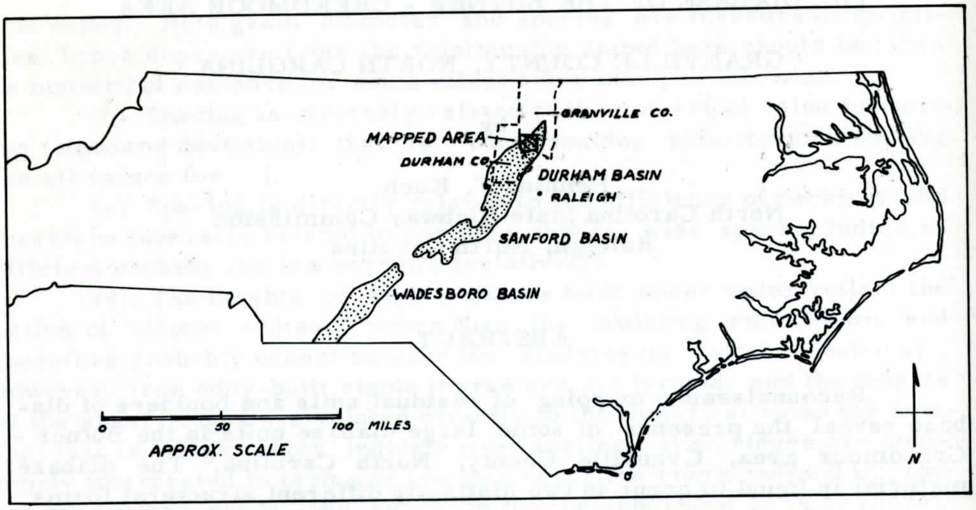

Figure 1. Index map of North Carolina showing Triassic basins and the mapped area.

The area drains towards the south into Neuse River through a number of streams. The major tributaries are Knap of Reeds Creek, Ledge Creek, and Robertson Creek. The divide between the Neuse River and the Tar River drainage is located in the NE corner of the investigated area.

Both the Neuse River and Knap of Reeds Creek have well developed flood plains. All other streams have cut narrow ravines with moderately steep slopes. Both the flood plains and the bottoms of ravines contain alluvial material of varying composition (Figure 2).

Within the mapped area elevations range between 260 feet and 500 feet above MSL. In general, sediments of Triassic age occur at elevations between 270 feet and 400 feet, while adjoining rocks of the Piedmont Province range between elevations of 300 feet and 500 feet.

It is notable that most of the land typified by soils derived from diabase is wooded and shows a typically characteristic plant ecology. These areas generally have a more or less level topography and are poorly drained. On the other hand, the relief energy of the surrounding areas is more intense. Little farming is carried out on acreage covered by the residual soils of diabase.

Generally, water wells are not found to exist in areas containing this soil type. This observation is supported by the fact that the Records of Wells in Granville County does not list any wells within the area underlain by diabase or covered by its residual soils (personal communication from H. E. LeGrand, United States Geological Survey).

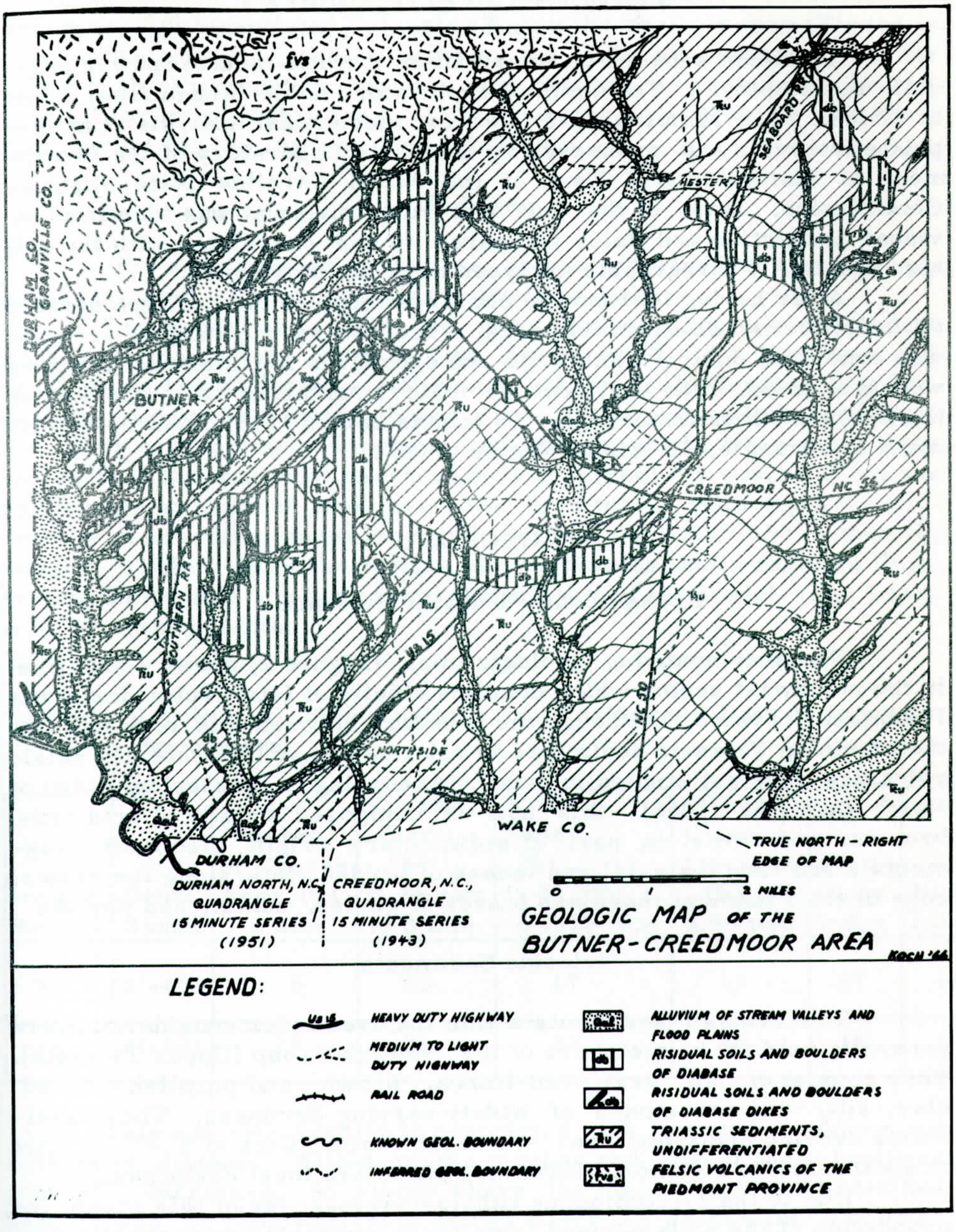

Figure 2. Geologic map of the Butner - Creedmoor area.

METHODS AND PROCEDURES

To obtain subsurface information, a truck-mounted power auger drill was employed. The depth of exploratory holes ranged from 3 feet to 47 feet. Soil samples were taken for mechanical analysis at appropriate intervals and processed by the Soils Laboratory of the Department of Materials and Test of the North Carolina State Highway Commission. Most of the soil and rock material was identified by visual inspection and later compared to and correlated with the soil test results. Petrographic analyses were not performed.

After having learned the characteristics of the soil type derived typically from diabase in this area, a reconnaissance mapping program was conducted during which soils and exposures in numerous locations were examined. The pertinent geologic boundaries were marked on the Durham North Quadrangle, 15 Minute Series (1951), and the Creedmoor Quadrangle, 15 Minute Series (1943).

GEOLOGY

Rocks of the Piedmont Province

The accompanying geologic map (Figure 2) shows a unit (fvs) in the northwest sector which consists chiefly of felsic volcanics of the Piedmont Province. Although the boundary between this rock unit and the adjoining Triassic sediments is relatively easy to locate, the felsic volcanics were not studied in detail. According to the Geologic Map of North Carolina (1958), this unit is composed chiefly of acid tuffs, breccias and flows, in part of sedimentary origin, also mafic fragmentals and flow material and lenses of bedded slate; along the eastern edge of the Piedmont there are lenses of gneiss, schist, and phyllite.

Triassic Sediments

The Triassic sediments within the area under consideration are generally held to be members of the Newark Group (Upper Triassic). They consist of light gray, red-brown, brown, and purplish colored clay, silt, and sandstones of widely varying hardness. These sediments overlap and interfinger with each other. Dips of 5° to 20° east to southeast are common. However, there are local deviations.

Due to the discontinuous lithological make-up of this group, the properties of the soils derived from these sediments vary greatly over a short distance. Despite a thorough and extensive drilling program, no reliable indications were found to justify any long range stratigraphic correlation in this area. It is believed that the Triassic sediments of the Butner-Creedmoor area constitute interfingering lithofacies rather than distinct formations. A similar conclusion was drawn by

R. L. P. Custer (1966). Therefore, this complex of sediments is shown as undifferentiated Triassic (Ʀu) on the geologic map (Figure 2).

The western contact between the Triassic sediments and the rocks of the Piedmont Province is very distinct and can easily be recognized for most of this area. J. W. Harrington (1948) has described this contact for the Durham Triassic Basin.

Diabase

The area mapped as diabase, is characterized by a brown clay soil associated with boulders and cobbles of diabase.

The typical soil which has been derived from diabase parent rock was mapped as Iredell Clay Loam by R. B. Hardison and D. D. Long (1912) during a soil survey of Granville County. The Iredell Clay Loam is a brown, heavy, clay soil of generally high plasticity. It is nearly impervious and exceedingly sticky when wet. Near the surface this material carries about 25 percent (by weight) of coarse aggregates and coarse sand (particles 0.25 mm). With depth the percentage of coarse granular material increases to around 50 percent (by weight). Table 1 gives the results of a mechanical soils analysis illustrating the typical change of characteristics of the Iredell Clay Loam with increasing depth:

Table 1. Mechanical Soil Analysis of Weathered Diabase from the Study Area.

Sample Number	Depth Range	Number of Coarse Aggr. (%) > 2 mm	Coarse Sand (%) 0.25 mm to 2 mm	Fine Sand (%) 0.05 mm to 0.25 mm	Silt and Clay fraction (%) 0.05 mm	Plasticity Index	Liquid Limit
52	0'-4'	5	20	17	63	47	66
53	4'-10'	20	36	20	44	14	29

This soil type usually is associated with rounded boulders and cobbles of diabase. With depth the number and size of these boulders generally increases. In places where the topsoil or overburden has been eroded or removed these stones cover the ground like a cobblestone pavement. Locally these stones have been used for the construction of yard walls and building foundations.

The diabase material occurs in this area in two distinctly different structural forms:

1. As dikes.
2. As horizontal bodies of varying areal extent.

1. The dikes are more or less vertical tabular intrusive structures composed of rounded and sub-rounded boulders and cobbles of diabase, which are enclosed in a usually dark brown clay soil. The dikes of this area rarely form any continuous geomorphic features. They vary in width from a few feet to several hundred feet. Some of them follow a straight line of strike, while others seem to be slightly curved. Some dikes consist mostly of residual soil, usually the smaller ones, while the wider dikes show a greater abundance of boulders and lesser amount of soil matrix. But this statement is quite relative. The stone-matrix proportion can change abruptly within one and the same dike over a short distance. The major directions of strike of these dikes is found to be either NNW-SSE or NE-SW.

2. The second mode of occurrence of the diabase material is as bodies of great areal extent, covering several square miles in the Butner area (Figure 2). The majority of these sheet-like masses of diabase are confined to an elevation range of 320 feet to 390 feet. Detailed topographic profiles across the diabase area southeast of Butner indicate an average surface slope of less than 1/2 degree toward the south. During the field work, several unsuccessful attempts were made to penetrate the diabase with the drill machine. Thus, no data are presently available as to the dip of this body.

As was learned from the study of several exposures contact metamorphic effects apparently have discolored and baked the underlying sedimentary material (Triassic). The overlying sediments, however, do not appear to be altered by excessive heat. The contact between the diabase material and the younger Triassic sediments exposed in some localities has the characteristics of a hiatus. The thickness of this sheet-like, highly disintegrated diabase body, is estimated to be on the order of 60 feet. The field data, obtained during a reconnaissance survey, suggest that many of the diabase bodies in the Butner-Creedmoore area represent remnants of flow-like structures.

CONCLUSIONS

Reconnaisaance mapping has revealed the presence of large diabase units in the Butner - Creedmoor area in Granville County, North Carolina. In many instances the boundaries of this material coincide with boundaries of the Iredell Clay Loam of the Agricultural Soil Map of Granville County.

The diabase material is found to occur in two distinct structural forms:

1. Narrow, vertical tabular intrusive structures, called dikes.
2. Sheet-like bodies of diabase material of great areal extent and having a thickness of usually less than 80 feet.

Geologic interpretation of the field data suggests a flow-like structure for some of the larger diabase bodies. A number of dikes might have served as feeders.

Macroscopic observations of the grain size and texture of the diabase seem to indicate a general decrease of crystal size from the center to the edges of diabase bodies. In areas believed to be feeder zones, crystals are appreciably larger (0.1 inch in diameter) than those observed toward the edges of a diabase mass, where the individual crystals often become macroscopicly undistinguishable.

The difference between the regional dip of the Triassic sediments (5° - 20°E) and the regional overall surface slope of the larger diabase bodies (less than 1/2°S) suggests the possible presence of an uncomformity.

The contact between the diabase material and the overlying sediments seems to represent an erosional surface (hiatus).

Reinemund (1955) and Harrington (1948) assign dike, sill, and sill-like structures to the diabase occurrences of the Sanford and Durham Triassic Basins respectively. This is certainly true for many of the diabase bodies. However, some of the sill and sill-like masses, found along the western edge of these basins may well be recognized as flows, if re-investigated under this aspect.

Further extensive and systematic field work will be required in order to clarify the structure and age relationship of this easy to map unit of diabase.

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A PLIOCENE TENNESSEE RIVER HYPOTHESIS FOR MISSISSIPPI

by

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ABSTRACT

It is hypothesized that the discontinuous gravel defended ridges which extend southwestward from Scott County, Mississippi, through Brookhaven are relict deposits from a Pliocene river course that may have been a southwestward extension of the Appalachian Tennessee River trend. Anomalous southeastward drainages in southern Mississippi are explained by this hypothesis.

* * *

The hypothesis that the Tennessee River flowed southwestward across Mississippi from the northeast corner of the state during Pliocene time would solve several of the State's stratigraphic and geomorphic problems. The course of such a Pliocene river, as we might visualize it, would be like that of the "Appalachian" river proposed by Grim (1936, p. 215) for Eocene time. The proposed river would have conducted a powerful stream of Appalachian waters through Tishomingo, Clay, Winston, Scott, Lincoln, and Franklin Counties in Mississippi toward a junction with the Mississippi embayment somewhere in Louisiana.

The large picture of the contemporary or Recent drainage pattern in Mississippi can be readily explained in light of what is known of the areal geology. With some striking exceptions, the major stream control today has been the dip and strike of the formations over which the streams travel, the obsequent slopes and local structures. We note, however, that certain streams violate the regional rule and are not related to dip and strike of formations: the Bogue Chitto, Bouie Creek, Big Black Creek, Okaloma Creek and much of the Leaf River in southern Mississippi flow extended courses at a pronounced angle to both dip and strike directions of the "Miocene" belt (Figure 1). This southeastward drainage system is explained by ridges of northeast trending, somewhat discontinuous (by virtue of dissection) fluvial sediment, which forms the high level surface at Brookhaven, Mississippi. The thick superficial deposits that underlie the ridges, and which appear to have defended them from complete removal by later erosion, have been mapped as Citronelle Formation (Pliocene-Pleistocene) (Mississippi Geological Society, 1945). It is a point in our hypothesis that very coarse gravels in the fluvial deposits under the ridges were once

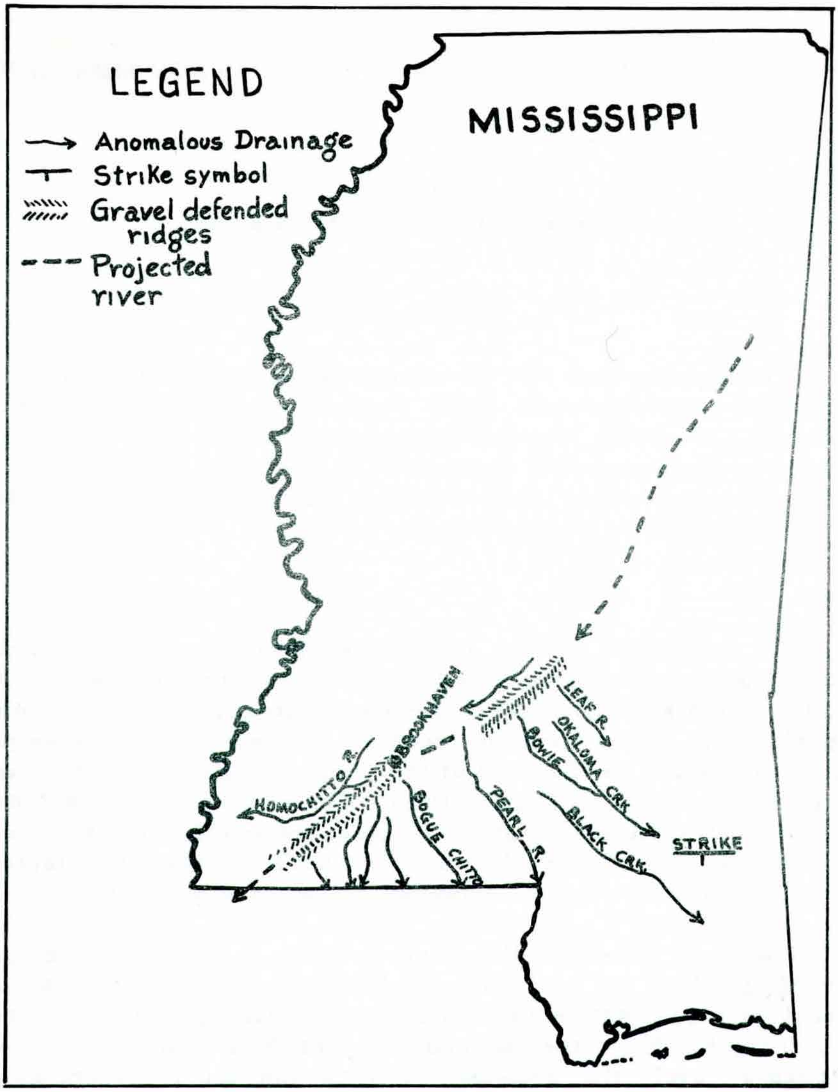

Figure 1.

laid down by a very large river flowing southwestward, and that they were once a part of that river's bed.

The Citronelle question is now a classic controversy, and it can be linked to the equally classic "orange sand" problem of Harper (1857) and Hilgard (1860). It is not possible to review here the many unresolved questions those problems have raised, but we should remark that a great deal of the still undissected interfluvial high level surfaces in southern Mississippi are directly underlain by an approximately ten foot layer of blocky, somewhat cross-bedded, orange-red, clayey sand, which is a distinctly mappable (but as yet unmapped) unit over a large area. This unit is well exposed near Lucedale, Mississippi, where it supports an important farm surface of mature soils. This surface was no doubt synchronous with the Williana surface of Fisk (1952). The sand sheet in places overlies thin gravels which appear to be its own basal channels, and it also unconformably overlies Citronelle sands and gravels and the "Hattiesburg" or "Pascagoula" clay beds. The contact with Citronelle is in many places sharply angular. The lateritic appearance of the orange-red sand sheet strongly suggests a pluvial climate, which is an environment consistent with the closing phase of Pliocene in Nebraska, when algal limestones were deposited (Schultz and Stout, 1941, p. 16).

Matson and Berry (1916, p. 173-174) noted that the percentage of sand and gravel in the Citronelle diminishes from north to south. If the Citronelle sheets were spread southward by the reworking of the main river bed deposits of the ridges, we would indeed expect to find the coarsest materials of the Citronelle close to the source ridges of sands and gravel toward the north and to find that the Citronelle becomes younger southward as it has been reworked by streams flowing in a generally southward direction. The similarity of the gravels would lead one to correlate them as a continuous sheet, as they have been mapped by the Mississippi Geological Society (1945). The Citronelle of Alabama-Florida and the Citronelle of Texas-Louisiana could not have been derived exclusively from the ancestral river segment we are here considering, and we would have to find other pre-Pleistocene drainages for them. The mineralogically dissimilar character of the three Citronelle areas, as recognized by Matson and Berry, does point to a tripartite source for the gravels.

If we move northeast along the Brookhaven ridge (Figure 1), over the relatively narrow belt of thick (more than 100 feet) coarse sands and gravels we enter Scott County and an area of mapped Citronelle outliers. These remnants of very coarse gravel contain petrified logs and indicate a major stream with considerably more flow than streams there today. Further along the trend, in Winston County, boulders up to 905 pounds (Mellen, 1939, p. 41) have been found in the terrain of so-called Holly Springs sands. The post-Tallahata deposits of Choctaw County also consist of gravel and boulder beds. Although the Choctaw deposits have been referred to the Eocene, suggesting Ralph Grim's Eocene "Appalachian" river, a later age is certainly not out of the question since an Eocene age has not been, and indeed cannot be, proven. We might also wish to question the Upper Cretaceous age for some of the Tuscaloosa gravel and cobble beds in Tishomingo County. With more detailed study they too may prove to be a part of our proposed ancestral Pliocene river. Certainly the alignment of the gravel defended ridges in southern Mississippi suggests a water direction that would justify the as yet undatable and somewhat anomalous gravel deposits of Tishomingo, Choctaw, and Winston Counties.

There is admittedly much work to be done to prove the validity of the hypothesis we have here suggested. If the hypothesis were to be proven, it has the merit that it would neatly tie together in a uniform explanation the following unresolved anomalies:

1. The unusual thickness and coarseness of the fluvial deposits cropping out along a persistant northeast trend of ridges in southwest Mississippi.
2. The high rolling hills near Brookhaven which rise above the Coastal Plain as an apparent extension of the Appalachian topographic trend in an area where one would not expect to find so much topographic relief.
3. The poorly understood surface deposits of boulders and logs in Winston and Choctaw Counties.
4. The extensive sheet of lateritic orange-red sand with its mature soils (Hilgard's orange sand?).
5. The Pleistocene age of Citronelle along the coast, whereas Citronelle farther north underlies material on which was developed an early Pleistocene or Williana surface.
6. The Pliocene hiatus, embarassing for those stratigraphers who would make all of the Citronelle gravels of Mississippi-Louisiana be Pleistocene and Recent.

Any single hypothesis which will resolve so many diverse problems should be considered in pursuing further field work.

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ON SUSPENDED SEDIMENT SAMPLING BY FILTRATION(1)

by

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ABSTRACT

Filtration is often used to study the suspended sediment of natural waters. The majority of studies have used cellulose membrane filters. The recently developed metal membrane filters have a number of advantages over cellulose membrane filters, and the author recommends them for determinations of the concentrations of suspended sediment. For a complete study of suspended sediment two samples are required, one large sample preferably collected on a metal filter for mass determination, and a smaller sample collected on a cellulose filter for optical examination and size analysis.

* * * * * * * * * * *

The cellulose membrane filters used in the majority of suspended sediment studies are available from several suppliers in a wide range of pore sizes with closely controlled mean pore diameters. Among their important advantages are low unit cost, high flow rate, and transparency.(2) Their ash weight is less than 0.0001% of the original tare weight when combusted at 820 C. They have several disadvantages. The contain glycerol, a soluble material, amounting to as much as 6 percent of the total filter weight, they are very hygroscopic, and they are highly susceptible to electrostatic charges.

A clear discussion of a procedure for using cellulose membrane filters recommended by the author as the best available is to be found in Winneberger, Austin, and Klett (1963). To secure accuracy with

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(1) Contribution No. 101 from the Chesapeake Bay Institute and The Department of Oceanography, The Johns Hopkins University. This work was supported by the Department of Chesapeake Bay Affairs, State of Maryland, and the Bureau of Commercial Fisheries, U. S. Department of the Interior.

(2) While normally opaque, they can be made transparent by the application of a liquid with a refractive index near that of cellulose.

any procedure using cellulose filters the soluble material must be removed from the membranes, and the membranes dessicated before weighing. Simultaneous dessication of many filters in a large dessicator requires approximately 20-30 minutes between weighings--the time necessary for re-equilibration of the water content of the filters and the dessicant. In the method of Winneberger, et. al. this waiting time is eliminated by using individual dessicators. Handling dessicated cellulose filters is very difficult because soaked membranes curl badly. The recommended method relieves this handling difficulty to some extent. Finally, it is imperative that cellulose membrane filters be weighed under an alpha-emitting source(3). This technique eliminates electrostatic effects, a frequent source of spurious results.

The metal membrane filters(4) currently available, are made of pure silver, and come in most of the same pore sizes as do cellulose filters. They are superior to cellulose filters in several ways. They do not contain any soluble material, and therefore require no presoaking. They are not hygroscopic, so they can be weighed without preliminary dessication. They are less susceptible to electrostatic charges than are cellulose filters, but they must still be weighed under an alpha-emitting source. Handling difficulties resulting from curling are entirely eliminated--an important consideration if field work aboard a rolling and pitching vessel is to remain tolerable. Metal membrane filters would appear to be at a disadvantage since they cost nearly three times as much as cellulose membrane filters. However, such is not the case. They can be back-washed and reused if they have not been heated above 370 C. Their disadvantages are high unit cost, and obstinate opacity.

Since neither the cellulose nor the metal membrane filter is ideal in all respects, it was found expedient to adopt both kinds of filters and to devise a dual procedure for studies of suspended sediment. Metal membrane filters of 0.8 micron APD are used to collect material for mass determinations, and cellulose membrane filters of 0.22 micron APD are used to collect smaller, auxiliary samples for microscopic identification and sizing. If both mass and size are to be found, two samples, one large and one small, are required regardless of the filters and procedures used. Samples of very small(5) volumes of

(3) The Staticmaster Ionizing Unit, Model No. 2U500 mounted on a Flexible Arm Staticmaster Positioner BFL available from Nuclear Products Co., 10173 E. Rush St., El Monte, California, has been found to be most useful.

(4) Available from Selas Flotronics, Box 300, Spring House, Pa., 19477.

(5) The terms small and large must be defined in terms of the conditions existing in the region under study. The required volumes depend upon the concentration of the suspended matter and the mixing in the area.

water show great variability in the concentration of suspended sediment. To build a stable statistical measure of sediment mass in a large volume of water would require many small samples, many analyses, and much calculation. It is more efficient to filter large volumes of water letting the apparatus do the averaging. The resulting large samples of sediment have the added virtue of making it easier to attain precision in the mass determinations. However, such large samples of sediment do not lend themselves to microscopic identification and size analysis. The material is piled too thickly on the filter obscuring many of the particles and causing agglomeration. For size analysis the ideal sample is a single-particle layer with no particle touching another.

Since as many as 200 samples have been successfully handled in a single day aboard ship in moderately heavy weather, a few comments on practical matters may be of interest. Although it is not necessary to use individual dessicators with metal membranes, when large number of metal filters are being processed at one time the individual dessicators are extremely useful. They greatly facilitate identification of individual samples and help reduce mistakes in logging results. In addition, dessication of the samples of suspended matter is necessary before weighing since the suspended matter is generally hygroscopic. It is easier to put the filters in the dessicators immediately after the initial weighing. This permits identification to be placed on the dessicators rather than on the filters.

The individual dessicators are similar to those described by Winneberger, Austin, and Klett (1963). They are easily made from 4-ounce, wide-mouth, squat-form jars. The jars are filled with about one-half inch of dessicant into which is stuck a short, 30 mm, length of 40 mm glass tubing. The weighed filter rests on top of the glass tubing.

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The author is indebted to Blair Kinsman and James Carpenter for their helpful suggestions for improving the manuscript.

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water shows great variability in the concentration of suspended sediment. To build a reliable statistical measure of sediment mass in a large volume of water would require many small samples, many analyses, and much calculation. It is more efficient to filter large volumes of water, letting the water pass through the membrane. The resulting large amounts of sediment have the added virtue of making it easier to get precision in the mass determination. However, such large amounts of sediment do not lend themselves to microscopic identification and size analysis. The material is piled up thickly, a thick layer obscuring many of the particles and masking and obscuring. For size analysis the ideal sample is a single-particle layer with no particle touching another.

As many as 24 suspended-sediment samples, have been successfully handled in a single day aboard ship in moderately heavy weather. A few comments on procedural matters may be of interest. Although it is not necessary to use individual glass filters with metal membranes, when large numbers of small filters are being processed at one time the individual desiccators are extremely useful. They greatly facilitate identification of individual samples and help reduce mistakes in logging results. In addition, desiccation of the samples of suspended matter is necessary before weighing, since the suspended matter is generally hygroscopic. It is easier to put the filters in the desiccators immediately after the initial weighing. This permits identification to be placed on the desiccators rather than on the filters.

The individual desiccators are similar to those described by Winneberger, Austin, and Klett (1963). They are easily made from 4-ounce, wide-mouth, square-form jars. The jars are filled with about one-half inch of desiccant, into which is stuck a short, 30 mm. length of 40-mm. glass tubing. The weighed filter rests on top of the glass tubing.

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The author is indebted to Mary Kilgrow and Joanne Carpenter for their helpful suggestions for improving the manuscript.

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GEOMORPHOLOGY OF RIVER VALLEYS IN THE SOUTHEASTERN ATLANTIC COASTAL PLAIN: A DISCUSSION[1]

by

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INTRODUCTION

Colquhoun (1966), in an excellent recent article on river valleys in the southeastern Coastal Plain, stated "Similar landforms are found to be underlain by similar soils and lithologies in the surficial Coastal Plain sequences. Thus, where environments are known through isolation of biotic assemblages, extension of these environments can be accomplished through the aid of soils maps." Colquhoun then summarizes in a table the relations between a few of the major landforms in the Coastal Plain and the interpreted environments and common soil groups that underlie them. We know that good soil maps can be an aid in mapping geologic environments. But we do want to point out some of the dangers in using soils in this kind of correlation. In many places there is little cause and effect relation between depositional environment and soils; in others, there may be considerable cause and effect relation. We believe that we can explain this variation in cause and effect relations by outlining how soils are mapped, the criteria used, and some of the current concepts of soil classification.

METHODS AND CONCEPTS OF SOIL SURVEY

Before 1933, most soil maps were made on plane table surveys on a scale of 1 inch equals 1 mile, and were necessarily general in nature. Since about 1934, most soil maps in the southeastern states have been made on aerial photographs at a scale of 4 inches equals 1 mile. Aerial photographs have many advantages, but probably the main one is greater mapping accuracy produced by interpretation of the photographic pattern. On the 1 inch equals 1 mile base maps,

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there are many inclusions (soils other than the one mapped, that are either minor in area or very similar) within a delineation. Inclusions have not been eliminated by use of aerial photographs and larger scales, but the amount has been reduced.

Soil mapping and geologic mapping are similar, but most soil maps are much more detailed, and there is much less extrapolating between observations. Most soil surveys published since 1938 have been mapped at a rate of less than 1 square mile per work day. Soil Scientists commonly make 50 to 75 borings a day in mapping, but borings do not usually extend below 42 inches.

Detailed soil maps show soil series name, type and phase; for example, Norfolk loamy sand, gently sloping, eroded phase. Norfolk is the series name and loamy sand the type, based on texture of the plow layer. The phase is based on soil properties of importance to agricultural use; in this case, the slope gradient and erosion.

Until about 1960, color and texture of the B horizon (texture based on the USDA soil-texture triangle, Soil Survey Staff, 1951) were the main criteria used in differentiating soils formed in the sediments of the southeastern Coastal Plain. The color of the B horizon was used as an indication of internal soil drainage. This meant that there could be soils in seven series with about the same B horizon texture, the major difference being in drainage, as shown by the colors. Soils saturated with water much of the time and with black or very dark gray surface horizons high in organic matter were called very poorly drained. Those with pale surface horizons and with gray colors dominating in the B horizon were called poorly drained; if gray mottles in a yellow or brown matrix were within 12 to 24 inches of the surface, the soil was somewhat poorly drained; if yellow or brown colors dominated but some gray mottles were between 24 and 30 inches, the soils were moderately well drained. Well drained soils have no gray mottles within 30 inches of the surface. But there were three separations within the well drained soils, based on color of the B horizon. Soils were placed in different series, depending upon whether they had yellowish brown, yellowish red, or red B horizons. This sequence of soils, that differed largely because of drainage, was called a catena (Aldrich, 1956) or a toposequence. Where B horizons were sandy clay loams, this catena from the red to the gray member was made up of Orangeburg, Ruston, Norfolk, Goldsboro, Lynchburg, Rains and Portsmouth series. Soils of these series were mapped from the upper to the lower Coastal Plain and some of them from Maryland to Louisiana. Other catenas, differing mainly in the texture of the B horizon, were mapped over similar ranges of physiographic conditions.

All soil surveys made before 1938 showed considerable geologic bias. This was because American soil scientists in the early decades of this century were largely trained as geologists. Naturally they felt that geologic processes were of first importance in soil classification. In the late 1920's and the 1930's, the emphasis shifted somewhat to characteristics of the soils, but topographic position, source of parent materials and kinds of materials beneath the soil continued to influence the classification (Ableiter, 1949, p. 321; Ligon and Lyford, 1953, p. 403-404). For example, very similar soils were placed in different series, such as Kalmia and Norfolk, because one formed on a river terrace and the other on adjacent upland. Grady soils were mapped only in depressions in the upland, often called "Grady ponds," and were differentiated from other wet soils largely by landscape position.

Soil series are now much more restricted than formerly. Thickness of the A2 horizon, presence of fragipan horizons (Soil Survey Staff, 1960) (dense, brittle subsurface horizons), and presence of plinthite (iron nodules that become irreversably hard on repeated wetting and drying) are now used as series criteria. Thickness ranges of the solum and of the B (argillic) horizon now have narrower ranges than formerly. Soils formerly mapped in the Norfolk series would now be mapped in 5 or 6 different series, such as Lakeland, Wagram, Goldsboro, Fuquay, Dothan, etc. This doesn't mean that the old mapping was bad; it means that a larger range of soil properties was once allowed within a soil series. Most published surveys were mapped using the more general concept of soil series.

A revised scheme of soil classification, introduced by American soil scientists at the Seventh International Congress of Soil Science in 1960 (Soil Survey Staff, 1960) and adopted with further revisions for use by the United States Department of Agriculture in 1965, is based primarily on morphology and composition of the soil. In this revision, classes in all categories are set apart entirely by soil characteristics that can be observed and measured. The geologic processes resulting in parent materials of soils are of no significance in soil classification, except as their influence has carried over into present soil characteristics. For these reasons, soils in any one series can occur on several different landforms and geomorphic surfaces and on sediments deposited under several different environments.

Users of soil maps for broad interpretations should keep in mind the following principles: (A) Soil classification at any time reflects the current understanding of soil genesis (Simonson, 1962) because that understanding governs the selection of properties as class criteria. In 1927 it was logical to separate all soils into two classes, one in which calcium carbonate had accumulated and the other in which aluminum and iron had accumulated. It was thought that the former developed only in regions of low rainfall and the latter in regions of moderate to high rainfall, and that the classes were mutually exclusive. It has since been found that they are not exclusive, and that processes of accumulation and removal are going on all the time in soils at different rates and to different degrees. Hence, these classes do not have the significance they were thought to have. It is apparent that no scheme of soil classification can be better than the soil science of its day. As new knowledge is added, classification must take it into account. (B) Classification evolves while mapping is done. Each survey area completed adds to the information on which the legend for the next can be based. Maps of adjoining counties completed and correlated at different times may vary greatly, not only in amount of detail but in definitions of mapping units and the bases of which units are separated.

RELATIONS BETWEEN SOILS AND DEPOSITIONAL ENVIRONMENT

A few examples can point up the difficulty in interpreting the relations between soils and the depositional environment of their sediments.

The area near Newton Grove, North Carolina, is typical of hundreds of square miles of the Coharie geomorphic surface in central North Carolina. The surface is gently undulating and has no established drainage pattern. The underlying sediments are mixed sand and clay with about 10 to 15 percent silt. The sediments have little structure, and discontinuous pebble lenses are common. Gravel is most abundant at the base. It would be very difficult to differentiate these sediments from those underlying the Coharie terrace of Colquhoun (1965). Sediments and the overlying surface in the Newton Grove area probably are of fluvial origin. Down the divide at Mount Olive, the character of the sediments and surface changes. Mount Olive is in a small, basin-like, low area on the Coharie surface. The topography has only very gentle undulations. The underlying sediments are mixtures of fine sand, 15 to 35 percent silt, and 20 to 40 percent clay. The bedding is dominantly horizontal and clayey and silty lenses are numerous. Some of the sediments have a pH of less than 3.0 after repeated wetting and drying. These samples with a pH of less than 3.0 have more than 0.5 percent sulfur. Sediments at Newton Grove have pH's above 4.0 and essentially no sulfur. The Mount Olive area probably was part of a brackish water swamp or estuarine environment. The Newton Grove area probably corresponds very closely to Colquhoun's Continental Emerged environment, and the Mount Olive area is closest to his Continental Submerged marsh environment (Colquhoun, 1966).

The distribution of soil series at Mount Olive and Newton Grove is shown in Figure 1. Soils in the same series or groups of series occur at both places. The only mappable difference in morphology of the soils[2] is that at Mount Olive all the soils have fine sandy loam or

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[2] Soils were mapped on 8-inch-per-mile photographs at a rate of 100 to 150 acres per day. These areas were mapped as part of a joint research project of the Soil Conservation Service, USDA, and North Carolina State University at Raleigh studying the relations between geomorphology and soils.

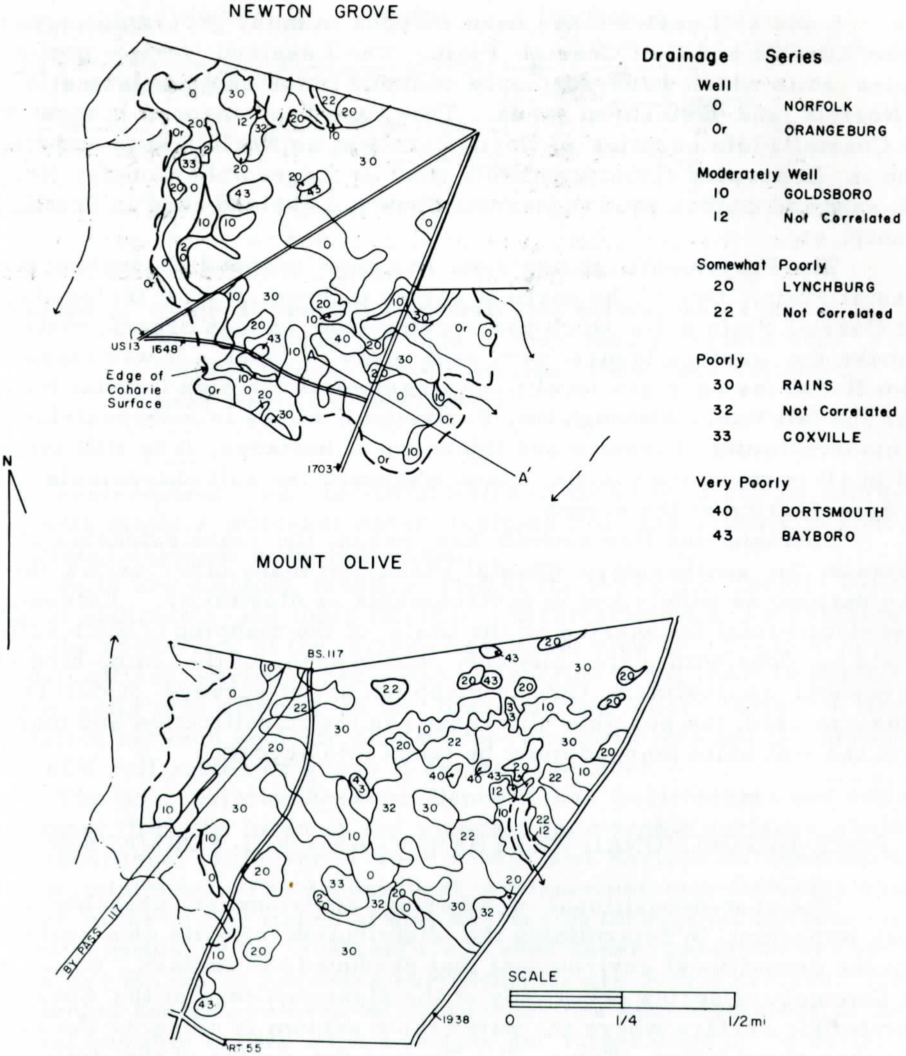

Figure 1. Distribution of soil series on the Coharie geomorphic surface at Newton Grove and Mount Olive, North Carolina.

loamy fine sand surface horizons; whereas at Newton Grove, the surface textures are medium sandy loam or loamy medium sand. In both areas, the better drained soils occur near the dissected edge of the surface and on the irregular shaped "high" areas away from the edge. These "high" areas are only 0.5 to 3 feet higher than the lowest area on the local landscape. The poorly drained soils are in the "low" areas of the surface or on the smooth flat divides.

Some soil series have been mapped in many geographic areas of the Atlantic and Gulf Coastal Plain. The Lakeland series now includes soils which were formerly mapped in the southeastern states as Norfolk and Galestown sands. They have been mapped in most of the Coastal Plain counties of North Carolina; as Norfolk sand and fine sand in Brunswick County, Norfolk sand in Edgecombe County, Norfolk sand and coarse sand in Harnett County, Lakeland sand in Scotland County, etc.

The Portsmouth series also has been mapped throughout the Coastal Plain. One of the earliest series mapped, it first included all wet Coastal Plain soils which had accumulated enough organic matter to make the surface layers very dark gray or black. It was mapped from flat areas near sea level to depressions in the high Coastal Plain near the fall line. Although the Portsmouth series is now restricted by modern limits of texture and thickness of horizons, it is still mapped in all parts of the Coastal Plain wherever the soil characteristics fit the definition of the series.

Lakeland and Portsmouth are two of the more extensive soil series in the southeastern Coastal Plain, but many other series have been mapped as widely and in environments as dissimilar. Extension of environmental boundaries on the basis of the mapping of such soils should be done with care, and only within areas of the same kind of survey and approximate time of mapping; or, if surveys of different dates are used, the possible differences in the classification and make-up of the soil units mapped must be taken into account.

POST-DEPOSITIONAL WEATHERING AND SOIL FORMATION

The post-depositional weathering environment probably is more important in determining the distribution of soils on a surface than the depositional environment that produced the surface. Soils in the Orangeburg series occur only at the dissected edge of the Coharie geomorphic surface where the water table seldom is closer to the surface than about 5 feet (Figure 2). Toward the center of the drainage divide, the Norfolk soils may have a water table within 2 feet of the surface during the wettest part of the year. The Coharie surface in the Newton Grove area probably is better drained now than at any other time in its history. This is because the streams have cut back toward the divides and locally have cut below the base of the surficial sediments. It is likely that earlier the site of the Orangeburg soils had a water table as high or higher than the present site of the Norfolk soils. The yellowish red B horizons of the Orangeburg probably is a rather late development in the history of the Coharie surface.

The distribution of the Orangeburg and Norfolk soils on the landscape is not the result of different depositional environments because the sediments under the Coharie at Newton Grove were deposited

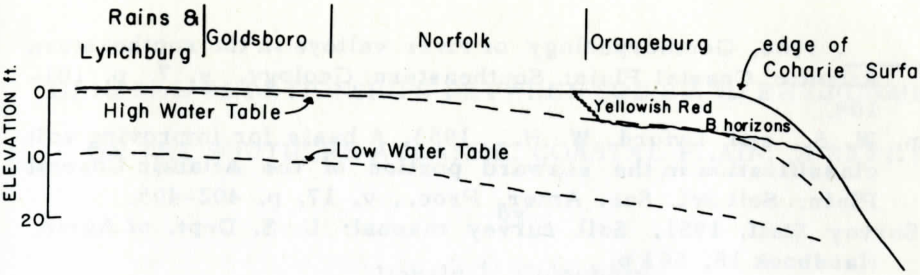

Figure 2. General relations between soil series, water levels, and the dissected edge of the Coharie surface near Newton Grove, North Carolina. Data were collected near line A-A' in Figure 1.

in one environment. But the distribution of these soils can be correlated with slightly different water regimes that are related to post-depositional changes in the landscape.

Soils are a highly modified rind at the top of the Coastal Plain surficial sediments and their internal characteristics are a combination of the properties of the original sediment plus the effects of their post-depositional environment. The interaction is the most important. Because soils are mapped on the basis of their characteristics, any correlation between geologic depositional environments and the distribution of a soil series or group of series is tenuous.

The best relation between depositional environment and soils would most likely be found on the youngest geomorphic surfaces where post-depositional differences in drainage and horizon differentiation are at a minimum. The depositional environment may influence soil drainage in the initial stages of soil weathering through its influence on the configuration of the surface; i.e., sand dunes generally are well drained and marsh or estuarine environments are poorly drained. The environment and source area influences texture and mineralogy of the sediment and this in turn affects texture and mineralogy of the soils.

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GENERAL RELATIONSHIPS BETWEEN SOILS AND ENVIRONMENTS ON THE SOUTHERN ATLANTIC COASTAL PLAIN: A REPLY

by

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INTRODUCTION

The preceding discussion explains to my satisfaction many of the dangers involved in extension of lithology and geomorphology based on modern soil maps alone, without necessary subsurface investigations in surrounding areas, provides a good history of the evolution of soil mapping in this country to date, which should be borne in mind when working on the Coastal Plain or in the Piedmont, and illustrates well an example of non-correlation between topography and soils.

Similar geomorphic surfaces are underlain by generally similar soil types and lithologies on the Coastal Plain. The correlation is most apparent in the Lower Coastal Plain and is often obscure in the Middle and Upper Coastal Plain (Figure 1). Good general correlation between soil types and landforms exists commonly in many areas on the lower terraces; the Wicomico, Penholoway, Talbot, Pamlico, Princess Anne, and Silver Bluff. For example, in these areas ancient bar-like sand bodies, which surficially show linear ridges and swales as well as suppressed dune-like topography, are expressed physiographically on topographic maps as well as by outline of differing soils on county soil maps (see Figures 60 and 61 and discussion, Soil Survey Staff, 1966, p. 74-76). Thus, the sand body observed landward of the Bethera Scarp or that landward of the Cainhoy Scarp in Berkeley County, South Carolina (Figures 2 and 3), or that landward of the Waccamaw River near Conway, South Carolina, in Horry County, or many others present on the lower Coastal Plain are all expressed on county soil maps (Figure 4). Moreover, and possibly by coincidence, in the case of the Cainhoy Scarp feature, flats developed behind the Cainhoy Scarp, which are shown on the Berkeley County maps to be underlain by a sandy soil and to change landward to loamy or clayey soils (Figure 4) are found to be expressed overlying sandy and silty-clayey lithofacies in the subsurface at depth (Figure 3). Similar observation can be noted in other areas as well and with other landforms and sedimentary bodies in such common occurrences that I must conclude that a general relationship often exists between the types of soils shown on the county soil maps and the nature of the landform itself. This relationship is expressed by general similarity of the boundaries of groups of soil types and the

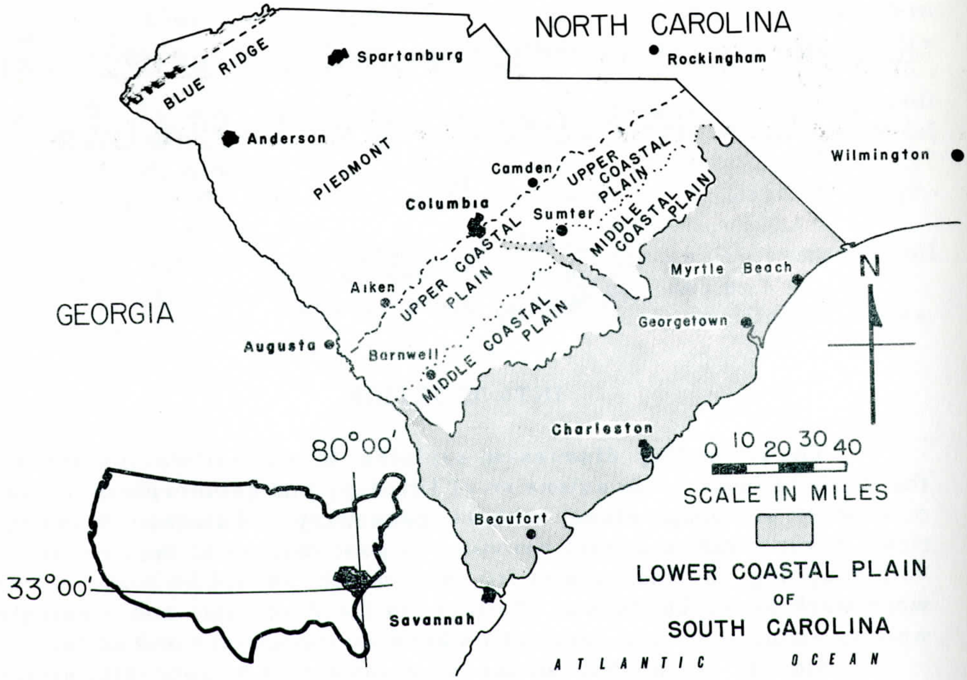

Figure 1. Major physiographic divisions of the South Carolina Coastal Plain.

topographic boundaries of the landform (Figures 2 and 4) as well as general similarity of the soil types within and between particular landform types (Figure 4). This does not imply that a landform may be identified on the basis of soils unique to it alone; for, as is abundantly clear, many of the soil series and types noted in my table of "some common soil types---" (Colquhoun, 1966, p. 108) occur over the entire Coastal Plain.

FACTORS OF SOIL FORMATION

Soils are generally accepted to be a product of the properties of the original sediment as well as post depositional environment. Relief, vegetation, parent material, climate, and age of the landforms are factors of soil formation. Both parent material and relief are related in many areas of the Lower Coastal Plain (Figure 1) where landforms resemble terrestrial and marine erosional, transportational, and depositional bodies formed during fluctuations of sea level.

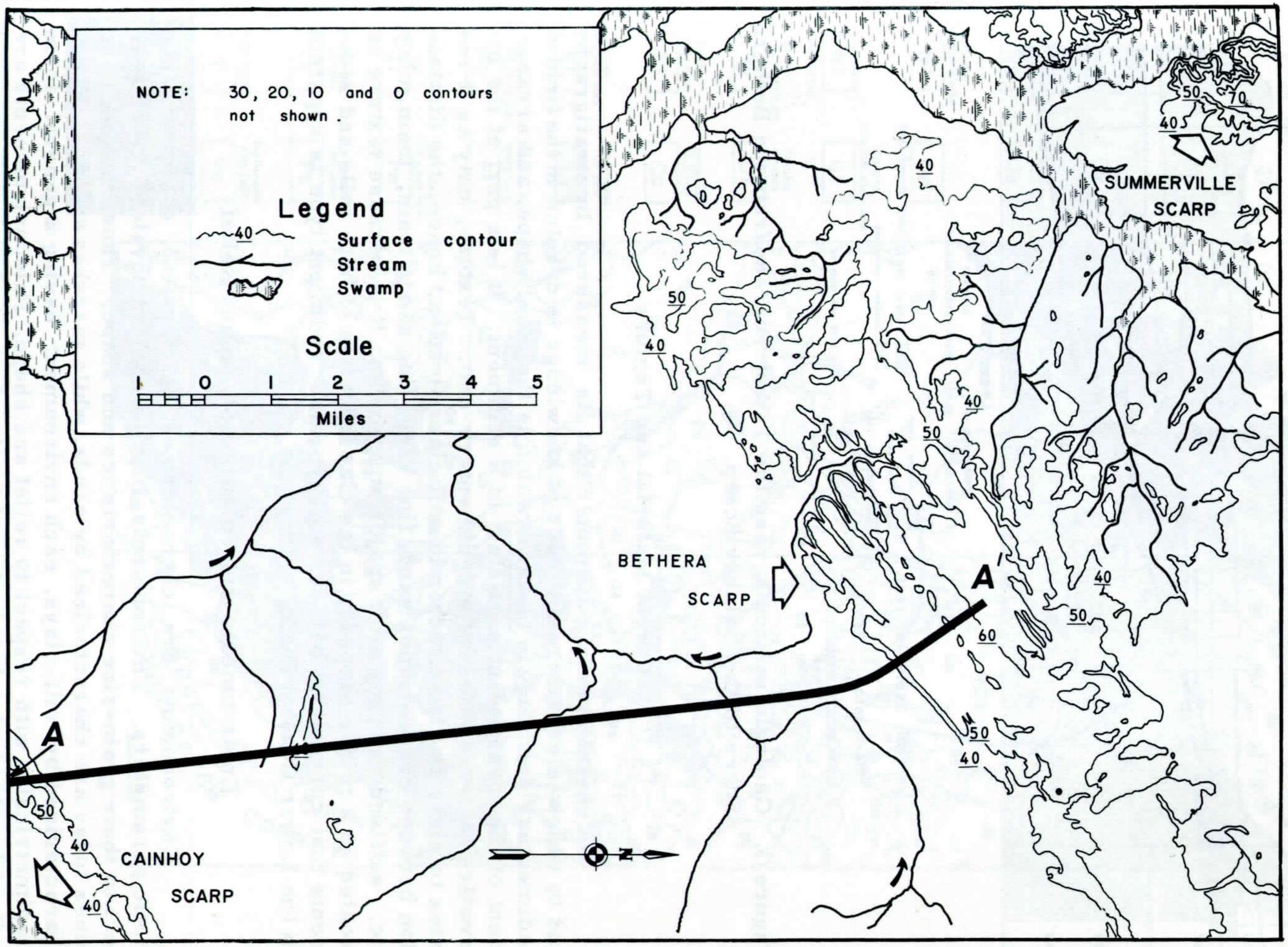

Figure 2. Major topographic features of the Talbot Terrace. Thirty, 20, 10, and 0 contours not shown.

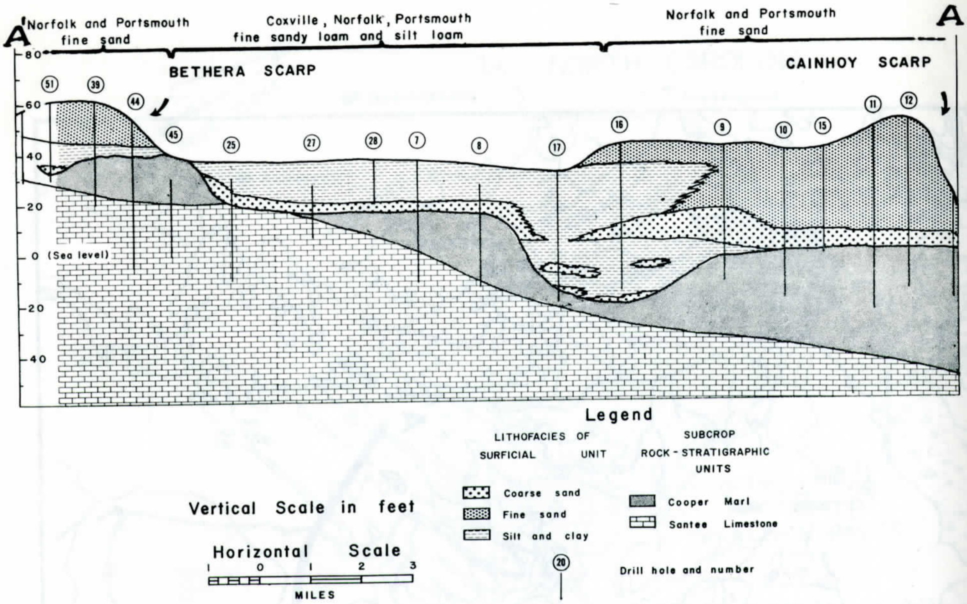

Figure 3. Geologic section across the Talbot Terrace from the Bethera Scarp to the Cainhoy Scarp.

Parent Material and Texture

An environmental parameter that is measured in stratigraphy and to which a considerable body of knowledge is devoted in the field of sedimentary petrology is texture, which is the size, shape, and arrangement of the constituent particles of a sediment. It is a part of the internal characteristics of a sedimentary unit. Texture, only as it relates to size, is also a factor in soil classification, hence, the distinction between coarse sand, sand, fine sand, fine sandy loam, loam, clay, etc., suffixed as types after major soil series. It is because texture is treated as a major property in the classification of both soils and sediments that correlation between soil type and sediment type is apparent on the Lower Coastal Plain.

Environments and Texture--Shape and Relief

Environments are characterized by physical, chemical, and biotic parameters. Environments may be studied physically with respect to their grain-size characteristics and shape. Thus, beaches and dunes today are characterized by sands, while marshes and swamps are characterized by silt-clays, each environment having a general form planimetrically with respect to relief and shape. Paleonvironments are defined by physical, chemical, and biotic characteristics as expressed within sediments. Thus, physically ancient beaches and dunes are

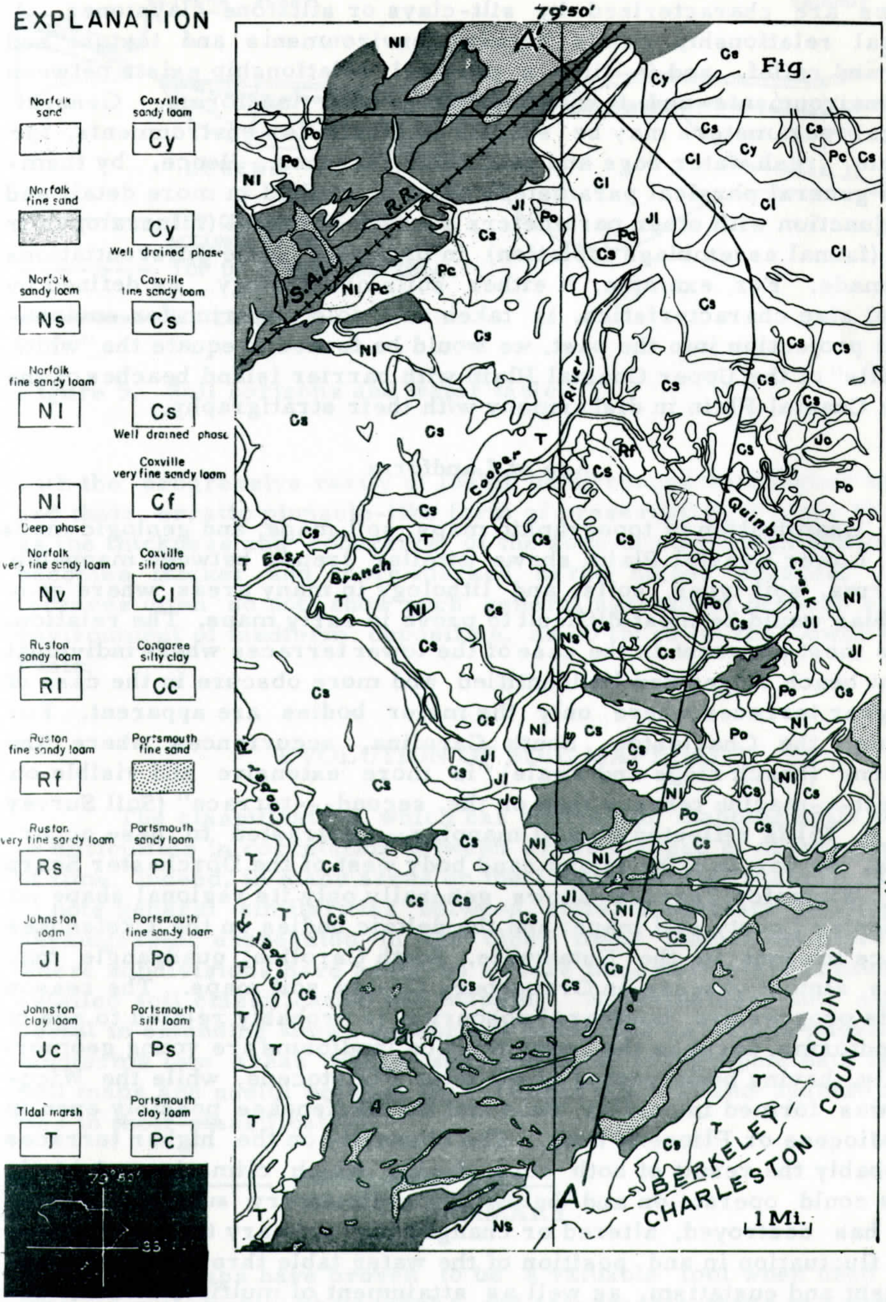

Figure 4. Soil map of a portion of Berkeley County, South Carolina; Latimer, W. J., et al., 1918.

characterized by sands or sandstones, while ancient marshes and swamps are characterized by silt-clays or siltstone-claystones. A physical relationship exists between environments and texture and shape and relief, and in time a physical relationship exists between paleoenvironments and textural lithology and landforms. General physical parameters may be similar in dissimilar environments, for example, fresh water bogs and salt water marshes. Hence, by themselves general physical parameters must be studied in more detail and in conjunction with other parameters such as chemical (mineralogy) or biotic (faunal assemblage isolation) in order for some differentiations to be made. For example, if either soils or lithology, as defined by general size characteristics, is taken as a sole criterion for environmental projection into the past, we would be forced to equate the "white sandhills" of the Upper Coastal Plain with barrier island beaches of the Lower Coastal Plain in disharmony with their stratigraphy.

Age of Landform

Examination of topographic maps, soil maps, and geologic maps of the Lower Coastal Plain shows similar trends between mappable landforms, soil types, soils, and lithology in many areas where geologic bias would be most difficult to prove in early maps. The relationship is most apparent in the case of the lower terraces where individual ancient beach ridges can be identified and more obscure in the case of the higher terraces where only the major bodies are apparent. For example, the Charleston, South Carolina, occurrences where "the landform" (beach ridge and swale) "is more extensive and visible on the first---marine terrace than on the second---terrace" (Soil Survey Staff, p. 75) is reflected by soil mapping. Contrasted to these occurrences, the bar-barrier island sand body west of the Dorchester Scarp on the Wicomico Terrace shows generally only its regional shape on Dorchester County soil maps; and the deltaic bodies on the Okefenokee Terrace present in the Eutawville, South Carolina, quadrangle map area is almost obscure on Orangeburg County soil maps. The reason for this progressive landward obscurity is probably related to age of the landforms, because the lower terrace landforms are young geomorphically, having been formed in the later Pleistocene, while the Wicomico was formed in earlier time and the Okefenokee possibly even in Late Miocene or Pliocene time. The obscurity on the higher terraces is probably the result of both time during which climatic and biotic forces could operate on and below the sedimentary surface, erosion which has destroyed, altered or changed the primary topography long range fluctuation in and position of the water table through stream encisement and eustatism, as well as attainment of multiple alluvial depositional stages in the past in many higher terrace areas. These factors as well in part probably account for the general presence of carbonate material within the soils of the lowest terraces contrasted

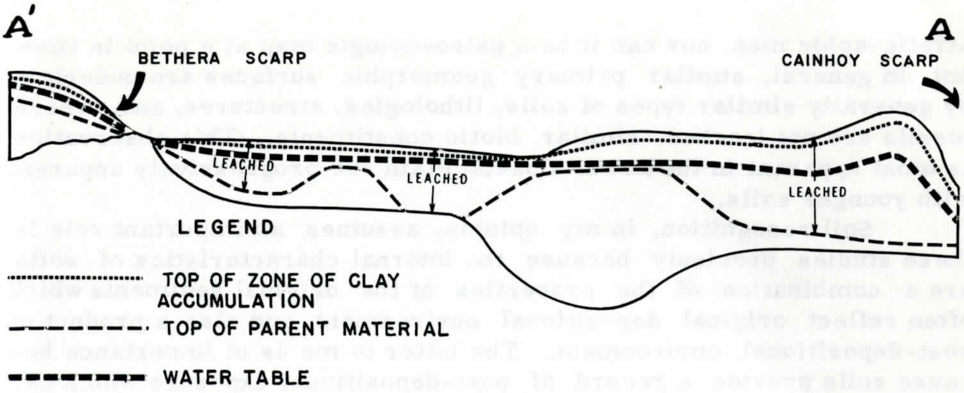

Figure 5. Soil horizons and water table.

with the progressive rarity of fossil occurrences with higher terraces and their erratic pinnacle-like form of preservation (Figure 5) as well as the thickness and coherence of the zone of clay accumulation which becomes thicker and more marked in the higher terraces. Higher terraces often do not show such general agreement between soils and environment of landform deposition, as do those of the Lower Coastal Plain.

EVOLUTION OF SOIL MAPPING

The classification which has evolved in mapping soils can lead to difficulties in correlation between areas mapped at different periods of time. Indeed, in many areas on the Lower Coastal Plain, it is easier to note general similarities between soils, subsurface lithology, and physiographic expression in the older maps than in the more recent where subdivisions have assumed a more important place in much more detailed soil classification and mapping. Nevertheless, soil maps are useful in extending known subsurface conditions as well as physiographic landforms into areas where subsurface or topographic data is weak. Soil maps are useful because they reflect textural and spacial properties in their classification.

SUMMARY

Soil maps have proven to be a valuable tool when used in conjunction with aerial photographs and topographic maps in extension of known subsurface and surface data. By themselves they have inherent difficulties for this use as has been stated, and they must be evaluated constantly when making correlations. A soil map can never be a rock-stratigraphic map, nor can it be a paleoecologic map at a point in time; but, in general, similar primary geomorphic surfaces are underlain by generally similar types of soils, lithologies, structures, and, where fossils are not leached, similar biotic constituents. This observation is most apparent in the Lower Coastal Plain and progressively apparent with younger soils.

Soil recognition, in my opinion, assumes an important role in these studies precisely because the internal characteristics of soils are a combination of the properties of the original sediments which often reflect original depositional environment and also a product of post-depositional environment. The latter to me is of importance because soils provide a record of post-depositional exposure which can conceivably be missing within a study area. For example, in Pleistocene glacio-eustatism, sea level fall is accompanied by encisement of rivers and erosion. Record of that fall is difficult to determine since sedimentary deposits related to it may be concentrated largely on the continental shelf. However, weathering in divide areas continues during the fall and during the subsequent rise which buries the former land surface. If interpretation is possible, these buried divide areas provide important evidence for marine regression and subsequent transgression, to the extent that they remain preserved. Thus at the base of surficial sediments underlying the Talbot Terrace northwest of the Bethera Scarp in Berkeley County, South Carolina, the former divide areas formed on the older Wicomico, Penholoway, and Cooper Marl sediments commonly express a zone of clay enrichment and leaching very similar to weathering profiles within the surficial sediments today.

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