SOUTHEASTERN GEOLOGY

PUBLISHED AT DUKE UNIVERSITY DURHAM, NORTH CAROLINA

VOL. 13 NO. 2 SEPTEMBER, 1971

SOUTHEASTERN GEOLOGY

PUBLISHED QUARTERLY

AT

DUKE UNIVERSITY

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1971

STABILITY OF COASTAL PLAIN SURFACES[1/]

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ABSTRACT

Stratigraphic, geomorphic, paleobotanical and climatic evidence indicate that the drainage divides in most of the middle and part of the upper Coastal Plain have not been eroded since the surficial sediments were deposited. These lines of evidence are the general parallelism between the base of the surficial sediment and the divide surface, the common occurrence of an upper fine-grained sediment grading down to a coarser-grained sediment near the base, a regional slope of only two to four feet per mile, and the absence of an integrated drainage net on broad flats. Some of these drainage divides may have been stable for nearly 10 million years. This is a long time for a geomorphic surface to escape erosion, especially one so near a coast that has had several transgressions and regressions of the strand line.

Extrapolation of erosion experiment data from the Piedmont to the Coastal Plain indicates that under a normal vegetative cover there is essentially no erosion in the Coastal Plain on slopes of 2 to 4 feet per mile. But if these surfaces had been bare, rain drop impact and runoff could remove material. Except for an unknown but probably small amount of tree throw and exceptionally hot forest fires, there is

1/ Paper number 3284 of the Journal Series. Joint Contribution from the Soil Conservation Service, USDA, The Department of Soil Science, North Carolina Agricultural Experiment Station, Raleigh, North Carolina and Geology Department, University of North Carolina, Chapel Hill, North Carolina.

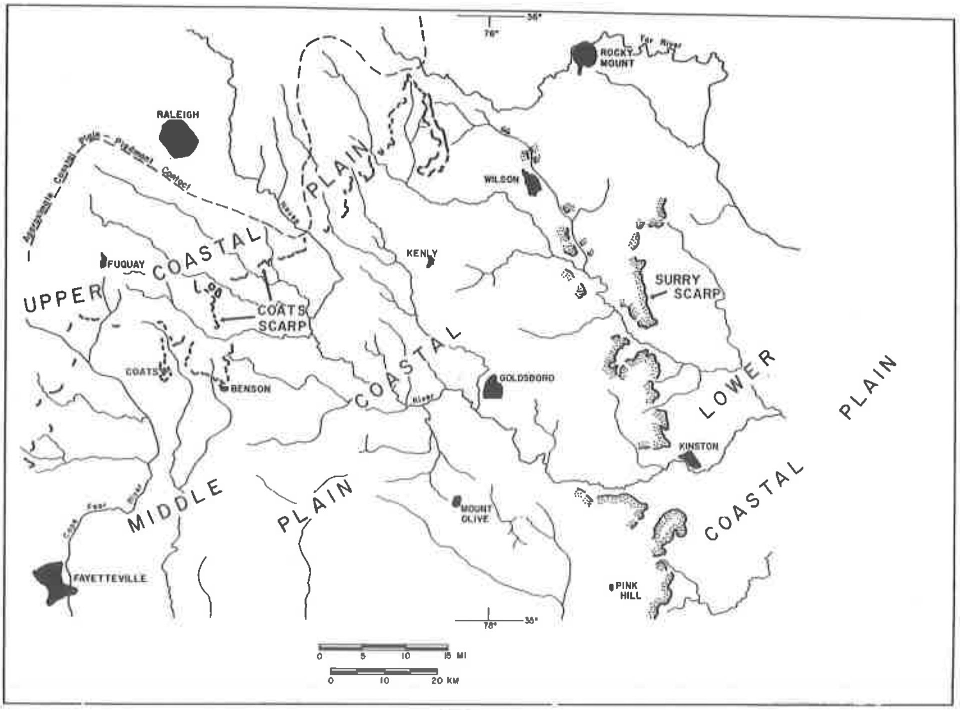

Figure 1. Distribution of the Upper and Middle Coastal Plain in the Neuse Drainage Basin, North Carolina.

no evidence that the soil surface was bare or had a sparse vegetative cover during Tertiary or Pleistocene. The geomorphic, stratigraphic, paleobotanical and climatic evidence all indicate the Coastal Plain divides have not been eroded.

INTRODUCTION

Stratigraphic and geomorphic studies in the Coastal Plain of North Carolina indicate that the middle and upper Coastal Plain surfaces and associated sediments above the Surry Scarp (Figure 1) are post-late Miocene and probably Pliocene in age. In terms of years this means that the sediments were deposited 10 to 3 million years ago. Stream valleys have been cut through these surfaces, but there is no field evidence that suggests the broad, gently undulating to nearly level divides, or table lands, have been eroded. If the field evidence suggests stability of the surface, then it follows that the soils on the surface have been weathering continuously since the surface was formed. Time zero for soil formation in fluvial sediments is the time deposition ceased, or in the case of a marine sediment, when the ocean withdrew.

If the sediment is post-late Miocene or early Pliocene, then the soils may have been exposed to weathering for nearly 10 million years. To many soil scientists and to some geologists, this seems to be an impossible length of time for a soil to have undergone development without at least some erosion. As one reviewer said about a statement that some soils have been weathering for nearly 10 million years, "The removal of even less than one micron a year would lower the surface 3 meters, which is greater than the thickness of the soil profile." Many soil scientists have been trained in the American Midwest where most soils are young. In the Midwest, late Pleistocene or Recent erosion and/or late Pleistocene deposits dominate the landscape, and many of the soils are less than 20,000 years old. Even to a geologist who has worked in the Midwest Pleistocene, it is a shock to come to the Southeast and suddenly have to start thinking of pedogenic processes going on for hundreds of thousands or even millions of years rather than mere thousands of years. It is this factor that makes geologists acquainted with the Midwest Pleistocene skeptical about the age of geomorphic surfaces in the Southeast. The question we raise here is just how stable geomorphic surfaces such as the Sunderland, Coharie, and Brandywine are. Have these surfaces been stable and their form unmodified since the sediment was deposited, or has erosion removed varying amounts of material? We hope to answer these questions partially in the following paragraphs.

EVIDENCE FOR PRE-PLEISTOCENE ORIGIN OF UPPER AND MIDDLE COASTAL PLAIN SURFACES

The dating of nonfossiliferous sandy sediments must be done almost entirely on the physical relations of one sedimentary body to another one of known age. The stratigraphic column for the area of the upper Coastal Plain near the Coats scarp (Daniels et al., 1966b) and the Middle Coastal Plain above the Surry scarp is shown in Figure 2A. The dashed lines in Figure 2 indicate that sediments do not physically overlie the stratigraphically lower unit, but that they are areally separated by scarps (Figure 2B). The Brandywine, Coharie, and Sunderland may all be one formation (Daniels and Gamble, in press), not separated as other authors have suggested. Within the Neuse drainage basin, the youngest fossiliferous formations that the Pinehurst disconformably overlies is the late Miocene Macks Formation. Younger sediments disconformably overlie the Miocene Yorktown Formation (Daniels et al., 1966b; Mundorf, 1946; Pusey, 1960). These surficial formations therefore are post-late Miocene. The fact that the weathering profile crosses the contact between the Macks and the Pinehurst Formations near Benson led us to believe that the Pinehurst and the Macks are closely related. The Pinehurst may be a fluvial deposit related to

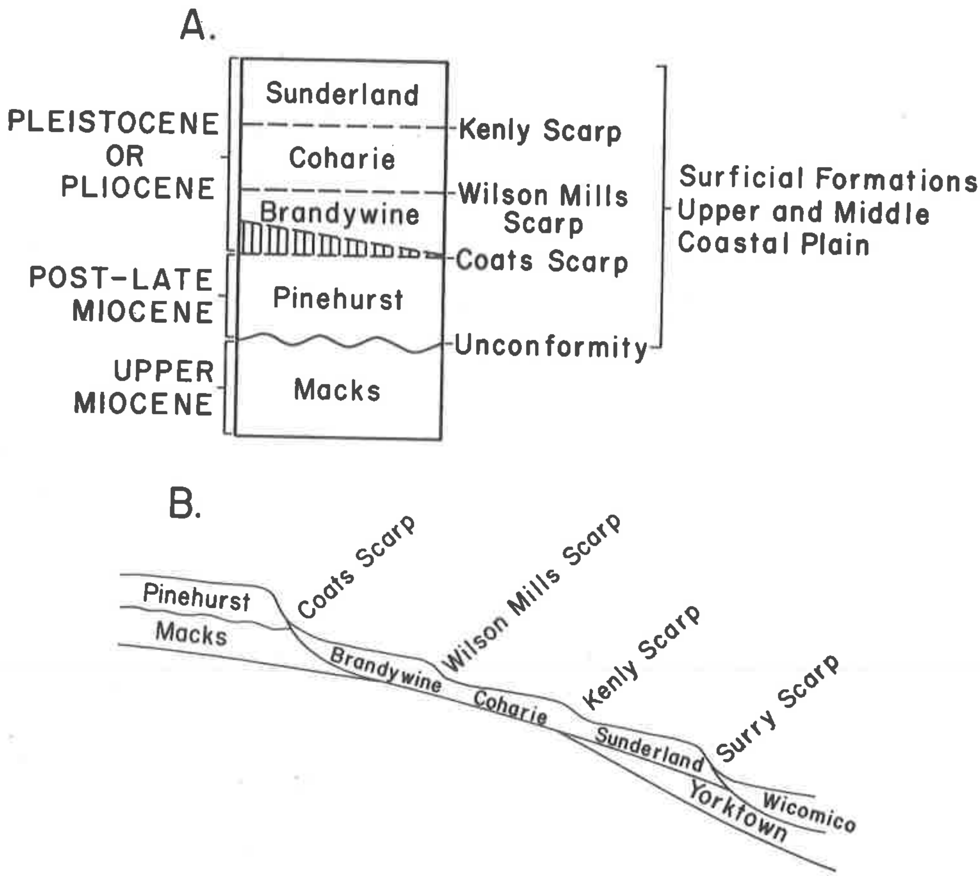

Figure 2. A. Stratigraphic column for upper and middle Coastal Plain along the Neuse-Cape Fear Divide.
B. Generalized section showing relations among stratigraphic units along the Neuse-Cape Fear Divide.

regression of the Macks sea. This would place the Pinehurst within late Miocene or possible early Pliocene.

The Pinehurst Formation is definitely older than the middle Coastal Plain Brandywine, Coharie, and Sunderland surficial sediments because it is truncated by the Coats scarp (Daniels et al., 1966b). With the Macks in the late Miocene, and the Pinehurst and Coats scarp post-late Miocene, it seems reasonable that the middle Coastal Plain probably is no older than early Pliocene. Its minimum age is more speculative.

Surficial sediments of the Sunderland "formation" are truncated by the Surry scarp (Daniels et al., 1966a) and are overlain at the scarp by the Wicomico Formation. The Wicomico is thought to be Pleistocene in South Carolina by Colquhoun et al. (1968). This age is based on a limited number of fossiliferous sections, but it must be considered when placing a lower age limit on the middle Coastal Plain. Because Pleistocene fossils cannot now be subdivided into those typical of upper, middle, and lower Pleistocene, the middle Coastal Plain can only be dated as being pre-Wicomico Pleistocene. Thus, a Pliocene-Pleistocene age must be assigned.

SURFACE CHARACTERISTICS

The upper Coastal Plain near Benson, North Carolina, is dissected and only remnants of the pre-Brandywine depositional and erosional surfaces remain. The Plain View surface, on the Pinehurst Formation (Daniels et al., 1966b), is not continuous along the interstream divides but is cut out by numerous topographic saddles. Large flats are uncommon, and most divides are 1/4 mile wide or less and in cross-section are gently convex or have a slope of less than 50 feet per mile. Maximum relief is about 2 to 3 feet along a 300-foot line (Daniels et al., 1970) with an average slope of 0.7 feet per 100 feet.

The topography along the divide changes across the Coats scarp. In the middle Coastal Plain, broad gently undulating divides replace the discontinuous smooth divide areas of the upper Coastal Plain. Maximum relief averages about 1 foot along a 300-foot line, and the average slope is less than 0.5 feet per 100 feet. Seventy-five percent of the divide are in the middle Coastal Plain has a slope of less than 0.3 feet per 100 feet. Slopes may be 1 to 2 feet per 100 feet from the rim to the floor of depressional or low areas.

Middle Coastal Plain surfaces appear to be nearly level, but they slope gently from the center of the divide toward the modern stream valley. This is a common feature and could suggest some post-depositional erosion of the surface. If postdepositional subaerial erosion was the agent, one would expect a smooth, gently sloping surface without depressions. But undrained depressions are as common near the valley slope as they are in the divide center. This sloping toward the modern stream valley probably is related to the agencies that formed these surfaces.

Several lines of field evidence suggest surface stability in the upper and middle Coastal Plain. These are: the general parallelism along the divide between the base of the sediment and the surface (Daniels, 1966b; Daniels and Gamble, in press); the common occurrence of an upper fine sediment grading downward to a coarser basal sediment (Daniels et al., 1966a, b; 1971; Daniels and Gamble, in press); the absence of an integrated drainage net; a regional slope of only two to four feet per mile; the lack of identifiable erosion debris in the swales on the surface (Daniels et al., 1967); and uniform thickness and sequence of color zonation at the well-drained sites in the upper Coastal Plain (Daniels et al., 1966b).

The field evidence cited above suggests that the gently undulating divide areas of the middle Coastal Plain have not been eroded since the end of deposition. It is true that streams have cut deep valleys into these surfaces in many areas. This erosion, however, has been merely nibbling at the edges of the surfaces which have been stable during the cutting of the valleys. The Plain View is a depositional and erosion surface that has been stable since pre-Brandywine time (Daniels et al., 1966b). This means that the surfaces and soils in the upper Coastal Plain can be nearly 10 million years old. The middle Coastal Plain surfaces and soils can range from 1 or 2 million to possibly 7 million years, assuming that current concepts of the underlying sediments (Figure 2) are correct. Two to ten million years is a long time for a surface to be stable by the concepts developed in glaciated country. If material had been removed uniformly from these surfaces, it is possible that it would not be noted by our field techniques, or that these techniques are too crude to record this type of information. We regard this possibility as unlikely, but some of the evidence that may bear on this is discussed below.

FACTORS AFFECTING EROSION

Because these Coastal Plain surfaces are not perfectly flat, there is always the possibility of some movement of surface material. Many field men believe that as long as there is runoff, in the sense of overland flow, there will be erosion. Some water from the local highs collects in the adjacent lows in areas of medium-textured soils even in heavily forested areas. This is not lateral movement through the mineral soil but largely across the soil surface. The evidence for this is that water in the low area is perched above the water table and the soils in the slightly higher areas show no evidence of free water. But before erosion can take place, the soil material must be detached from the mass. This detachment can be by raindrop splash and by water moving across the surface.

The literature on erosion by raindrop splash is extensive, and there can be no doubt that large amounts of material can be put in motion by this means. Osborn (1954) has estimated that two inches of rain on a bare soil might detach and set into motion more than 110 tons of soil per acre. This material can easily be moved by runoff. Before the stage of "rill" and "gully" erosion is reached, raindrop impact is a more important cause of soil detachment than runoff water (Rose, 1960). Running water gains ability to detach and transport material as turbulence develops within the water (Ekern, 1950). Ekern states that raindrops add a tremendous vertical velocity to the flow and that the combined agents are an effective erosive force. He also found that when raindrop impact is prevented, sediment is deposited because the detaching and transporting capacity of the runoff decreases rapidly. This is illustrated by the dumping of sediment load under automobiles and on hay strips where raindrop impact energy is absorbed by metal or plant parts.

Raindrop splash is most effective on a bare surface. Plant cover intercepts drops and by absorbing the impact energy can effectively reduce soil detachment and movement (Osborn, 1953, 1954). The effectiveness of cover in preventing erosion is related to the amount of cover present. The degree of surface cover and the weight per acre are satisfactory measures of the effectiveness of the cover. The woody stems of trees and shrubs are less effective per unit weight than the finer stems of grasses, but both can reduce splash erosion to zero. For example, 6,000 pounds of oat straw per acre was 100 percent effective in preventing splash erosion (Osborn, 1954).

The unburned forest floor under pines or hardwoods has a 1 1/4- to 3 1/2-inch-thick layer of litter, partially decomposed plant parts, and humus (Heywood and Barnette, 1936). Splash erosion through this floor probably is zero. Even the openings have a dense grass cover where splash erosion should also be ineffective (Heywood and Barnette, 1934). The forest floor also would reduce the erosiveness of any runoff.

Only a small amount of erosion data is available for very low or 0 percent slopes. Forrest and Lutz (1944) found that on bare soil some sandsized material was moved by runoff on tobacco rows with 0 percent slope. In agreement with Hjulstrom's (1935) data they found that the coarser sand fractions were lost in relatively small amounts compared to the 0.02 to 0.10 mm fraction. This erosion resulted in the surface soil becoming more sandy and the remaining sand becoming coarser.

Erosion data from several contrasting areas are summarized in Table 1. Under vegetation on slopes of less than 10 percent, the erosion rates are low, especially in forested areas. But data in Table 1 are not directly applicable to erosion from surfaces with slopes of 0.5 feet per 100 feet or less. Slope is an important factor in determining erosion rates when all other factors are held constant (Duley and Hayes, 1932; Lutz and Hargrove, 1944; Browning et al., 1948; Wischmeier et al., 1958). We have been unable to find erosion data from vegetated plots on 0 percent slopes, but it might be possible to extrapolate from laboratory experiments. Duley and Hayes (1932) experimenting with bare soil in the laboratory found approximately 40 times more soil eroded from 10 percent slopes than 0 percent slopes (4,684 pounds per acre from 10 percent slopes and 117 pounds from 0 percent slopes). If approximately the same ratio could be applied to the North Carolina forested plot (Copley et al., 1944), a site with 0 percent slope would lose a foot of soil in 20 to 40 million years (see Table 1). Such extrapolations are seldom accurate but may be in the right order of magnitude. From erosion plot data of soil scientists and geologic inferences, one can infer stability of the Coastal Plain surfaces so long as they are vegetated. About the only way these surfaces could be eroded by runoff

Table 1. Erosion rates from Iowa, California, and North Carolina.

Station	Clarinda, Iowa[a]		California[b]		North Carolina[c]			
Vegetation	Alfalfa	Bluegrass	Pine	Grass	Grass[d]	Virgin Hardwood	Burned Hardwood	Second growth oak-chestnut
Slope	9%	9%	40%	40%	10%	10%	10%	51% mean (10 to 80%)
Average yearly loss in tons/ acre	0.01	0.03	0.077[e]	0.077[e]	0.01	0.0019	3.08	0.076
Years to erode 1 foot of soil[f]	200,000	130,000	26,000	26,000	200,000	1,000,000+	650	26,000

a Browning et al., 1948
b Kittredge, J., 1954
c Copley, et al., 1944, and Dils, R. E., 1958
d First year record discarded because plot was in wheat
e Includes plant debris
f Computed as 2,000 tons per acre foot

would be to reduce the cover and expose the soil in some manner so splash erosion could become effective. Added support for this conclusion can be found in the work of Schumm (1956). He found that the rate of denudation increased exponentially with increasing relief. If the same principle can be applied to these Coastal Plain surfaces, the denudation rate should be at a minimum because relief on these surfaces seldom exceeds 5 feet and more frequently is 2 or 3 feet.

Tree throw with a frequency as described by Denny and Goodlett (1956) for areas of mature topography in Pennsylvania is a possible mechanism for the exposure and movement of soil. The soil clinging to the roots of a tree can be detached by raindrop impact and made available for transportation by runoff water. But runoff velocity through the forest floor should be very low, and soil movement by this mechanism would be slight and probably very local. The rates of soil movement produced by tree throw probably are much less than Denny found in Pennsylvania because the native Coastal Plain vegetation was hardwood and pine (Braun, 1950). Pine trees contribute little to tree throw because they have large tap roots and break off near the soil line. Four foresters, each with about an average of 25 years experience in the Southeast, believe that tree throw affects only a very small part of 1 percent of present forests in the Coastal Plain (written communication L. S. Metz, Research Soil Scientist, U. S. Forest Service, Research Triangle Park, N. C., 1970). Anyone struggling through a thick Coastal Plain forest is much more impressed by the large number of stump holes he steps into than by the number of tree throws he finds.

Soil movement by insects and burrowing animals has been amply demonstrated by Thorpe (1949) and Lyford (1963). But soil fauna expose very little material to raindrop impact on the forest floor.

Although processes such as tree throw and soil faunal activity may have some influence on erosion of these nearly level surfaces, considerably more quantitative information needs to be collected. Extrapolation of results from the other areas into the Coastal Plain can be dangerous because differences in tree species, soil characteristics, insects and animals involved, and possibly climatic factors may influence the amount of erosion. But catastrophic events such as major fires, extremely severe droughts, or climatic changes may change conditions on the forest floor so erosion can occur.

Under closed stands of pine the unburned forest floor is continuous and has a total weight of 10 to 25 tons per acre (Heywood and Barnett, 1936). A somewhat similar range in total weight of forest floor was found by Alway et al. (1933a, b) in Minnesota under jack pine, Norway pine, white pine, and maple-basswood stands. The 10 to 25 tons of forest floor accumulated in about 10 years in northern Florida and in 60 years or less in northern Minnesota. Even in openings in the longleaf pine a dense mat of wire grass 12 inches thick was found (Heywood and Barnett, 1934). The very heavy forest floor described for unburned forests probably characterized much of the virgin timber in the Southeast before Indians settled the area. This accumulation of fuel also would be conducive to fires during dry seasons.

We have been unable to fine much quantitative data on what happens when a hot fire sweeps through such an area. Fires are extremely variable and even a hot fire that kills trees may burn the forest floor to the mineral soil in places and leave it nearly intact in others (oral communication, C. T. Youngberg, Soils Dept., N. C. State, 1970). Thus under precultural conditions fires may have exposed the soil surface in small areas, but it is unlikely that large areas were uncovered simultaneously. Recovery of vegetation after a fire is rapid and is noticeable in only a few weeks. The bare areas would not remain for long, in most cases less than a year (Taylor and Wendel, 1964; DeCoste et al., 1968). Even in burned areas there is little change in the bulk density, percolation rate, and pore space of soils (Wells and Hatchell, 1968). Thus erosion would be low in bare areas. The sporadic distribution of fires in time and space and their variability in the damage they do make them a relatively ineffective method of producing a bare surface in the Coastal Plain.

The soil under longleaf pine burned annually is partly bare, and erosion by raindrop impact and surface runoff is possible (see Heywood and Barnette, 1934; Wahlenberg et al., 1939). Indians frequently burned forests (Bartram, 1928), and it was a common practice since the east coast was settled (Wahlenberg et al., 1939; Heywood and Barnette, 1934). In the last 10,000 years and especially in the last 300 years, parts of the Coastal Plain forests have been burned by man. How much erosion has occurred and how large an area has been affected cannot be evaluated but, considering the great length of time these surfaces have been in existence, this last stage is very short.

Severe droughts have little influence on the forest floor if they are only 2 or 3 years duration (oral communication, T. E. Maki, Forestry Dept. NCSU, 1970). But how about drastic shifts in climate? The record of climatic change in the Atlantic Coastal Plain since late Miocene is fragmentary but has been extended and summarized by Isphording (1970, p. 342). He concludes that "the climatic conditions in the Gulf Coast during the Pliocene, and possibly extending into a part of the pre-Nebraskan Pleistocene, were not much different from those in the present Gulf Coast." He infers that in New Jersey, "The climate was markedly different from present conditions with humid subtropics as far north as southern New York during the Miocene and early Pliocene." Isphording's contribution was the paleoclimatologic evidence of heavy mineral suites which he found to be consistent with the all too sparse paleobotanical evidence for the Tertiary of southeastern United States.

There is astoundingly little paleoclimatic literature that applies to the Tertiary of North Carolina, although the literature of North America is voluminous. From the small amount of evidence available, it would seem improbable that the climate was ever dry enough during the Tertiary to be classed as semiarid. By inference we can suggest that a continuous vegetative cover was present during Tertiary.

Whitehead (1963, 1964, 1969), Frey, (1953), and others have investigated the later Pleistocene to Recent pollen spectra in several areas of the North Carolina Coastal Plain. More northern types of vegetation are found at various levels, and some pollen diagrams suggest a savannah vegetation at times. Changes in density and type of tree species cannot be denied, but we must question what these changes did to the forest floor. Even what some consider a rapid or sudden change in climate occurs over a period of 1,000 to 1,200 years (Ogden, 1967). Changes in vegetation probably were gradual. Though conditions changed so that species die out during this span of time, their places in the flora are taken by other species better adapted to the new conditions. As Whitehead points out, there are southern type plants mixed with the northern types in many pollen diagrams. Thus it is unlikely that all the vegetation growing before the climatic change would die simultaneously. An orderly change over a period of time is more likely. This implies that at no time was the flora so depauperate as to cause any effective exposure of bare soil.

It is difficult to produce bare surfaces in the Coastal Plain for long periods by catastrophic events such as fire or droughts or by climatic change. The climatic fluctuations that have taken place probably had little influence on erosion rates on these surfaces because whatever variations did occur did not include any known periods of sparse vegetation.

If these surfaces have not been eroded, has anything been added? They could have been built up by addition of eolian materials, especially dusts. Very localized eolian deposits are recognized in the Neuse drainage along stream valleys (Daniels et al., 1969), but there is no evidence for a recognizable blanket deposit on these surfaces. Syers et al. (1969) have given evidence of aerosol deposition on the Coastal Plain surfaces in North Carolina. These aerosols are silts, 10 to 2 micron, derived from the Sahara Desert and transported by trade winds to North Carolina. Syers' work was based on the δ^{18} oxygen values of quartz. The amount of dust deposited this way is unknown. The North Carolina soils used by Syers et al. have a maximum of 17.8 percent silt in the 20 to 2 micron range. Possibly one-half this total is 10 to 2 micron silt. If one-half the silt content of the layers with the lowest δ^{18} oxygen values is assumed to be the original 10-2 micron silt fraction of the sediment, the maximum amount of aerosolic silt in any layer is 6 percent (Figure 3). The δ^{18} oxygen values are comparable to that of metamorphic quartz at about 100 to 150 cm in each profile. Thus, using even the maximum value of 6 percent, this is only a 6 cm layer of 10-2 micron silt.

The amount of atmospheric dust fallout at Barbados is computed by Delany et al. (1967) as 0.6 mm/1,000 years or 0.6 meter every million years; that measured in permanent snow fields by Windom (1969)

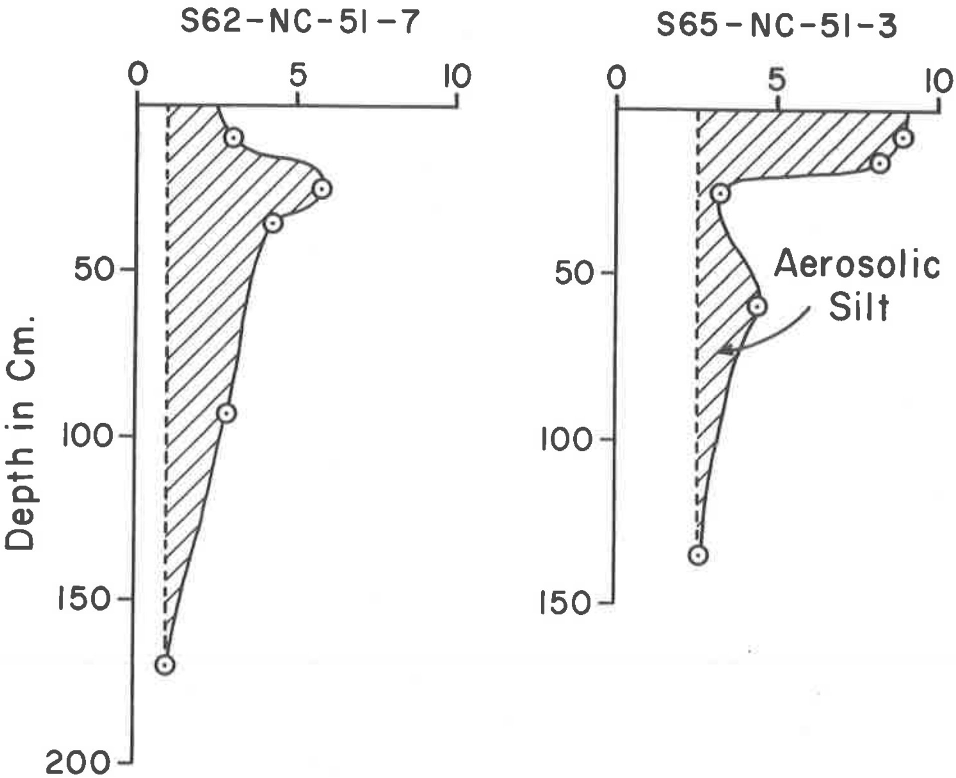

Figure 3. Estimated 2-10 micron silt as percentage of total soil at two profiles sampled in North Carolina by Syers. Hachured areas represent a possible volume of aerosolic silt.

amounts to 0.1 to 1.0 mm every 10^3 years or 1 to 10 meters every million years. These rates are much higher than those computed from the silt content of two North Carolina Coastal Plain soils. The rate computed from the silt content of the soils probably is somewhat low because the aerosols contain large quantities of illite, kaolinite, and feldspar - in some cases over 50 percent (Delany et al. 1967). These minerals can weather easily to clay size or to soluble materials. Thus the 6 cm addition to the N. C. soils may have been 100 percent thicker. But the rates computed by Delany and Windom seem high when applied to North Carolina because there is no evidence for a meter or more of silt on the Coastal Plain surfaces. The surface horizons of most Coastal Plain soils are sandy, not silty.

The solution of calcium carbonate, or any other material, has not played any detectable part in the development of these surfaces in North Carolina. Unpublished data show that the surficial sediments of the Upper and Middle Coastal Plain are composed overwhelmingly of quartz sand and kaolinitic clay. The subsurface formations of the Upper Coastal Plain are now essentially devoid of calcium carbonate. The very rare marine fossils (Macks Fm.) are either silicified or are merely molds (Daniels et al., 1966 b). In the Middle and Lower Coastal Plain some units, especially the Yorktown Fm. (Upper Miocene), contain conspicuous "shell beds." However, the shell beds constitute only a relatively small part of the formation and even a shell bed itself is only very rarely as much as 50 percent calcium carbonate and usually much less.

In the Lower Coastal Plain, the Castle Hayne Formation, which is outside of the area of study, may be dissolved in a local area to give a few scattered sinks. These are usually few and far between and not by any stretch of the imagination could the area be called "karst." It is lacking in even subdued irregularity due to solution.

In short, while a minor amount of lowering of the Coastal Plain due to solution is theoretically feasible, it seems most improbable that the removal of masses of calcium carbonate or any other theoretically soluble material of irregular thickness and variable purity could have resulted in the rather extraordinarily uniform upland surfaces that do now exist.

SUMMARY

Evidence from several sources indicates that erosion of the divides such as those in the Atlantic Coastal Plain can occur if the soil surface is bare or nearly so. There is no evidence that these surfaces have been bare after subaerial weathering started except for very localized areas and for the infinitesimal time periods following catastrophic fires or tree throw. Even on zero percent slopes with a closed canopy and heavy forest floor, it might be possible to erode a very minor amount of material. But extrapolation from erosion data under hardwoods from the western Carolina Piedmont suggests that it may require 20 to 40 million years to remove a foot of soil from the nearly level Coastal Plain surfaces. These surfaces have been in existence for less than half to one-quarter this time and the amount removed, if any, must be small. The addition of material from atmospheric dusts may be as large or larger than the maximum rate of erosion. We believe that there is no theoretical objection whatever to having the present Coastal Plain surfaces essentially unchanged for as much as the 10 million years since upper Miocene time. The real problem is the accurate dating of these surfaces.

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GIBBSITE IN COASTAL PLAIN SOILS,

SOUTHEASTERN UNITED STATES[1/]

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ABSTRACT

Gibbsite, kaolinite, interlayered vermiculite, and hematite occur in soils of the Coastal Plain in southeastern United States, forming the clay fraction of a soil zone 2 to 5 meters thick. These soils, which contain six percent or more gibbsite in the B and C1 soil horizon, occur as erosional remnants on uplands underlain by sand and gravel of the Citronelle Formation and high terrace deposits. They are restricted to a belt that extends from southern Mississippi, through the southwestern corner of Alabama, northwestern Florida, northeastern corner of Alabama, into west-central Georgia. Similar soils that contain very little or no gibbsite occur in the Coastal Plain beyond the gibbsite soils.

Possible source materials for gibbsite are weathered feldspathic rocks of the Piedmont physiographic province and clay in the Coastal Plain sediments. Bauxite deposits of early Eocene age, apparently, are not related to the gibbsite soil because they are separated from the gibbsite soils by unweathered marine sediments. Laterization probably occurred during the Pliocene, because intensely weathered soil overlies Miocene sediments and the lower terraces of Pleistocene age are not laterized.

INTRODUCTION

Gibbsite, kaolinite, interlayered vermiculite, and hematite form the clay fraction of lateritic soils in the Coastal Plain in southeastern United States (Figure 1). They occur as a series of discontinuous blanket deposits within an arc-shaped belt about 550 kilometers long in southern Mississippi, southwestern Alabama, northwestern Florida, southeastern Alabama and western Georgia. The gibbsite soils are very similar to other soils in the Coastal Plain that do not contain

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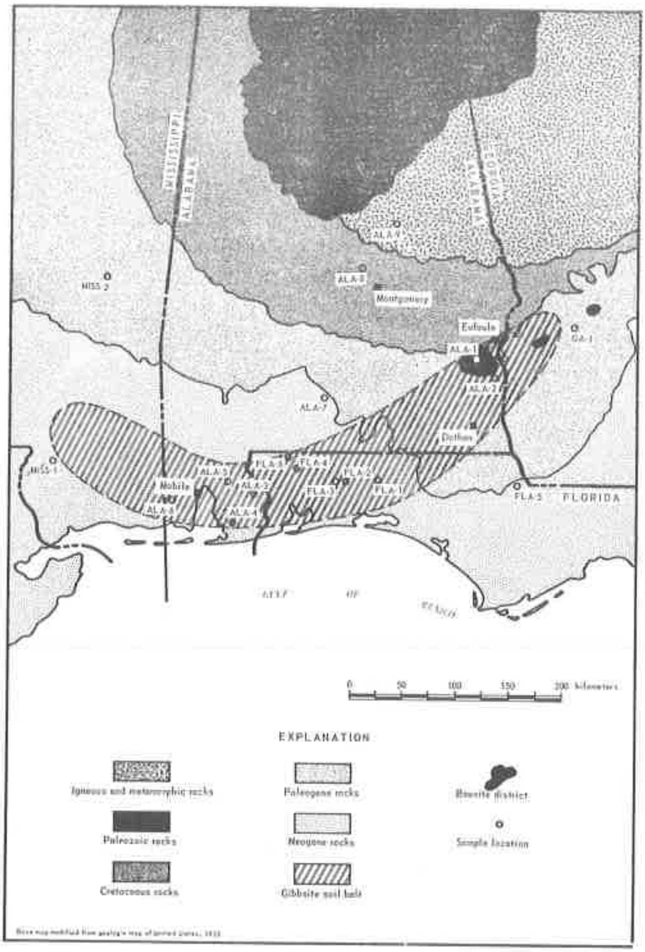

Figure 1. Map of southeastern United States, showing generalized location of gibbsite soil and sample locations.

significant amounts of gibbsite.

The gibbsite soils described here are significant because gibbsite occurs with quartz, although gibbsite and quartz are not mutually stable in a closed system (Kittrick, 1969). Gibbsite soils are defined here as soils containing six percent or more gibbsite in the B and C soil horizons. The clay fraction contains from 15 to about 60 percent gibbsite. These soils are classified within the Paleudults Great Group as defined by (U. S.) Soil Survey Staff (1967, p. 191, 192).

Previous Investigations

Pearson and Ensminger (1948) reported gibbsite as a major constituent in the clay fraction of Norfolk, Ruston, and Orangeburg soil series in the southern part of the Coastal Plain of Alabama. Fiskell (1959) reported gibbsite as the major constituent in the clay fraction of two soil profiles in northwestern Florida. Kaolinite and vermiculite were identified as minor constituents, and feldspar was detected in one sample. He also reported gibbsite and kaolinite were in the clay fraction of similar soils in Georgia and Alabama. Glenn and Nash (1964) described the weathering relationships between gibbsite, kaolinite, and expandable layer silicates in selected soils from the lower Mississippi Coastal Plain. They report a 14.2 A "aluminous chlorite" which, apparently, is identical with interlayered vermiculite reported in other areas. Fiskell and others (1970) gave properties, including mineralogy, of selected Coastal Plain soils from Alabama to Virginia.

Field Work and Methods of Investigation

This report is a result of intermittent geologic and mineral investigations in the Coastal Plain between 1959 and 1968. Field procedures include examining and sampling exposures in road cuts, mines, and gravel pits. The initial reconnaissance, which was made in 1959-1960 to evaluate the gibbsite soils as a source of alumina, included collecting and making partial mineral analyses of approximately 1, 000 samples throughout the southeastern part of the Coastal Plain.

Mineral resources of southern Alabama were investigated from 1963 to 1968 as part of the river basin investigation conducted by the Geological Survey of Alabama. Clay resources were investigated in cooperation with the U. S. Bureau of Mines. The Eufaula bauxite investigation, which was conducted concurrently with the clay studies, included research on gibbsite soils.

During the initial investigation covering large areas of the Coastal Plain, quantitative gibbsite estimates were made using a portable differential thermal analysis apparatus (DTA). The area of deflection caused by gibbsite in the sample, as plotted on a graph, was compared with an area of deflection from standard samples containing a known quantity of gibbsite. Gibbsite was determined chemically on selected samples by boiling the sample in 10 percent sodium solution and precipitating the alumina by 8-hydroxyquinoline.

During the latter part of the investigation, minerals were identified from peak positions on X-ray diffractograms. Preliminary quantitative estimates were made of gibbsite, kaolin, quartz, and illite, based on ratios of principal peak heights as compared to ratio peak heights of a series of standards prepared by mixing various proportions of minerals. Hematite, identified by red color, was computed from chemical analyses. Interlayered vermiculite and illite were estimated somewhat arbitarily "by difference," but consideration was given to peak heights on diffractograms. The total analyses were adjusted to total about 95 percent to allow for 2 to 3 percent anatase and traces of other minerals, keeping the gibbsite-kaolinite ratio as being accurate.

The preliminary analyses were checked against chemical analyses. In the absence of the interlayered vermiculite and illite, the procedure is fairly accurate because quartz, gibbsite, and kaolinite form definite compounds with almost no substitution of other elements. Charts were prepared showing percent gibbsite and kaolinite as related to percent Al_2O_3, SiO_2 and loss on ignition. The percentage of gibbsite present in the sample is obtained from caustic soluble alumina, making adjustments for water of crystallization. Kaolinite is estimated by subtracting gibbsite alumina (caustic soluble alumina) from total alumina and computing kaolinite from the chart showing percent mineral as related to alumina content. Although kaolinite is slightly soluble in hot caustic solutions, the solution of kaolinite was disregarded, because samples of kaolinite yield only 2 to 3 percent "caustic soluble alumina" using this method.

Acknowledgments

The manuscript has benefited by critical review by S. H. Patterson and J. G. Newton, U. S. Geological Survey; T. H. Hughes, University of Alabama; B. F. Hajek, Auburn University; and W. D. Keller, University of Missouri.

GEOLOGIC SETTING

Igneous and metamorphic rocks of the Appalachian Piedmont physiographic province, which have been subjected to intense chemical weathering and erosion, crop out 50 kilometers northeast of the northern terminus of the gibbsite soil zone. The main rock types are quartz mica schist, biotite and hornblende gneiss, amphibolite, phyllite, quartzite, and granitic rocks that range from granite to diorite. Gibbsite and kaolinite occur as soil constituents in mature soil profiles overlying deeply weathered granitic rocks (Clarke, 1963, p. 8-9; Patterson, 1967).

Consolidated, largely unmetamorphosed rocks of Paleozoic age crop out about 200 kilometers north of the gibbsite belt (Figure 1). The main rock types are limestone, dolomite, shale, and sandstone. This region has also been subjected to deep weathering since the close of the Paleozoic Era, but kaolinite and illite are the dominant clay minerals found in the weathered residuum (Clarke, 1964). Gibbsite has not been found as a major soil constituent in the residuum.

The Coastal Plain rock units are composed of predominantly marine sediments that consist of sand, sandy clay, limestone, and calcareous argillaceous sand of variable composition. The age of the outcrops are progressively younger southward, and they range from Cretaceous through Neogene. They dip away from the igneous-metamorphic complex and Paleozoic sedimentary rocks, and they reflect a facies change between Atlantic and Gulf Coastal Plain sedimentation. The region was elevated with respect to sea level during middle Neogene Period and sediments reflect a change from marine to continental deposition.

The topography of the Coastal Plain reflects the underlying geologic formations. Limestone and calcareous shale form valleys, and rock units that are composed predominantly of sand, gravel, and clayey sand remain as uplands, but they are dissected by streams. Uplands, which are a series of plateau, extend from southeastern Mississippi, across southern Alabama and northern Florida, into west-central Georgia. The plane formed by the tops of the plateaus is tilted gently southward toward the Gulf of Mexico. The altitude of this surface ranges from about 175 meters above mean sea level in west-central Georgia to about 18 meters above mean sea level in southwestern Alabama at the southernmost tip of the gibbsite soil belt.

GIBBSITE SOILS

Stratigraphic and Topographic Control

Gibbsite deposits occur in the soil zone of fluvial sediments of Neogene age that have been mapped as high terrace and as the Citronelle Formation. They disconformably overlie older marine sediments. The terrace deposits overlie the Tuscahoma Sand of the Eocene Series in the Eufaula bauxite district in southeastern Alabama, but near the Florida border, the terrace deposits overlie marine sediments of Oliocene Series. In northwest Florida, southwest Alabama, and southeastern Mississippi, the gibbsite soil occurs in fluvial deposits mapped as the Citronelle Formation of Pliocene Series. The Citronelle disconformably overlies marine sediments of the Miocene Series (Springfield and LaMoreaux, 1957; Marsh, 1966; Vernon, 1942).

The gibbsite soils are confined to uplands, where they occur as remnants of an old soil horizon that has escaped erosion. The correlation of gibbsite soils with the flat to slightly undulating surface of erosion dissected plateaus was confirmed by extensive sampling and observations in numerous road cut exposures. They do not occur in valleys or on the younger Pleistocene terraces.

Mineralogy

Gibbsite in soils cannot be recognized in the field unless part of the gibbsite occurs as concretions. The concretions found in Alabama and Florida soils are inconspicuous, ranging from 1 to 10 millimeters in diameter. They are brownish yellow to nearly white and they are soft enough to be cut with a knife. Gibbsite is easily recognized on diffractograms of samples where it is a major constituent by the 4.85, 4.37, and 4.31 A peaks. However, where gibbsite is a very minor constituent the 4.85 A peak is used. Kaolinite has been identified in all gibbsite soils tested.

Interlayered vermiculite or sometimes referred to as interlayered 2:1 expanding layer silicate is a constituent in soils of the Coastal Plain (Fiskell, 1959; 1970). It is associated with kaolinite, and sometimes it occurs as a minor constituent with both kaolinite and gibbsite. It is not common in bauxite deposits (Bardossy, 1966, p. 238). During this investigation, this mineral was recognized on diffractograms, but no structural or chemical work was performed.

Soil Profiles

The gibbsite under discussion is a product of soil development, and therefore soil nomenclature is used (U. S. Soil Conservation, 1960; 1967), with some modifications for simplicity. The term lateritic soil is retained to indicate soils that have reached an advanced state of weathering and contain more than 12 percent sesquioxides of iron and/or aluminum in the clay fraction. The soil may contain iron or aluminum oxide concretions. Gibbsite soils have been mapped in the Greenville, Red Bay, Orangeburg, Dothan, Tifton, Bowle, Malbis and Marlboro soil series, but soils outside the gibbsite belt that have been mapped in these soil series do not contain significant quantities of gibbsite (Fiskell, 1959; Fiskell and others, 1970), (Figure 2), (Table 1).

Although gibbsite soils of the Coastal Plain apparently were developed under forest cover, the organic horizons, 01 and 02, have been largely destroyed during historic times by plowing and by fire. The organic horizons have played an important part in the lateritic and gibbsitic soil formation, because the leaf and humus mat protected the soils from mechanical erosion.

Two subdivisions of the A horizon are recognized. The A1 horizon is a mineral horizon with an accumulation of humified organic matter. It is medium gray in color. Generally this zone is very thin, about 10 centimeters thick unless disturbed by plowing. The thickness of this zone reflects intense oxidizating conditions where organic matter is rapidly destroyed. The A2 horizon is a leached commonly sandy zone that lies between the A1 horizon and the B horizon (argillic). Thickness of this horizon is variable, generally between 20 centimeters to one meter, but in northwestern Florida, the thickness increases southward from less than 20 centimeters in the northern boundary of the gibbsite zone to more than 3 meters at the southern boundary. The A horizon contains very little gibbsite.

Both the B and C horizons are similar in appearance, but the B

Table 1. Approximate mineral and chemical analyses of typical samples of gibbsitic soil.

Sample No	Soil profile classification	Sample depth (meters)	Clay and silt fraction (percent)	Approximate mineral analyses of clay-silt fraction (percent)					Chemical analyses of clay-silt fraction (percent)			
				Quartz	Gibbsite	Kaolinite	Hematite	Vermiculite and illite	Al_2O_3	SiO_2	Fe_2O_3	Loss on Ignition
Ala-1	B, C1	.2 - 2.7	32	10	30	40	10	2 ±	35	31	15	16
Ala-2A	B	.9 - 1.5	37	10	20	40	15	10	31	36	15	15
Ala-2B	C1	1.5 - 2.7	29	20	15	35	16	10	28	40	16	14
Ala-3A	B	.3 - 1.2	52	10	45	25	10	5	41	25	11	20
Ala-3B	C1, C2	1.2 - 5.5	37	10	45	25	12	5	40	25	12	20
Ala-4	B, C1	.5 - 4.7	23	25	40	15	10	5	33	39	9.5	15.5
Fla-1	B, C1	1.2 - 5.6	20	10	30	40	12	5	40	25	12	20
Fla-2	B, C1	2.3 - 4.8	15	5	40	35	15	-	42	20	15	20
Fla-3	B, C1	.3 - 5.0	18	5	40	30	15	5	43	20	15	20
Fla-4	B, C1	.3 - 2.4	44	10	25	40	14	10	37	29	14	15

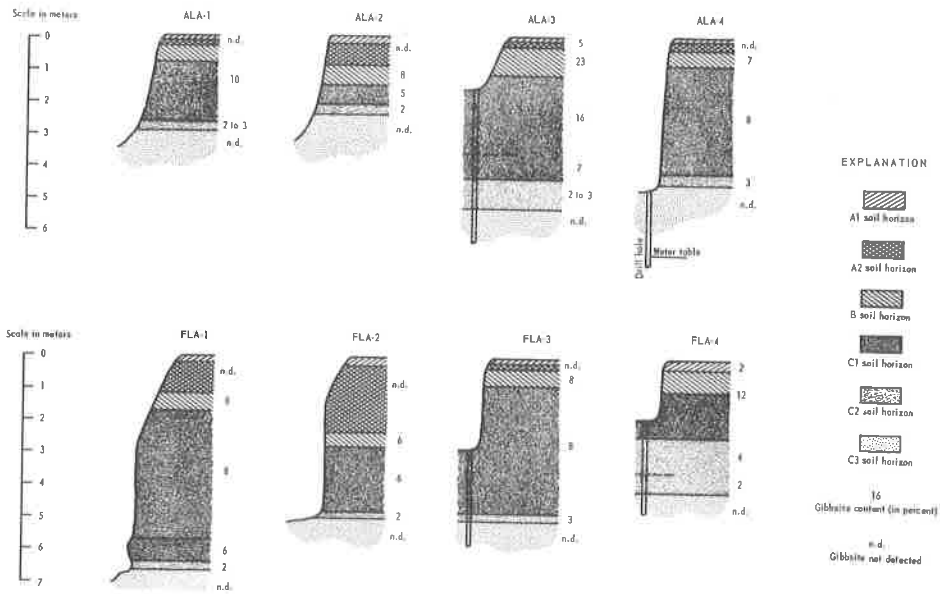

Figure 2. Typical gibbsite soil profiles.

horizon, as designated here, is more pronounced in illuvial concentration of silicate clay. Gibbsite occurs in both the B and C1 horizons. The B horizon is a moderate red color (5R 4/6) or a moderate reddish brown (10R 4/6). It is generally between 50 and 80 centimeters thick. The sand content is high, ranging from 50 to 80 percent.

The C1 horizon is similar to the B horizon, except that it has a more sandy texture. The thickness is variable and generally ranges from 1.5 meters to 5 meters. The C2 horizon is a mineral zone containing more sand, less clay, and also less gibbsite. The color is usually a shade lighter than the C1 horizon because of the greater sand content. The C2 horizon thickness ranges from about 20 centimeters to 2 meters, but it is generally less than 1 meter thick. Unconsolidated sand or sand and gravel underlie the soil zone.

In some areas such as northwest Florida near the Alabama border, gibbsite soils have been subjected to recent erosion, possibly because of early cultivation methods. The original A horizon has been completely removed, and a new Ap horizon is being developed from the old B horizon. This area is characterized by red surface soil with a high clay content that contains gibbsite.

Zoning of Gibbsite and Kaolinite

The gibbsite content of the clay fraction of the B and C1 soil horizons is highest in the central part of the gibbsite zone. The fringe areas of the gibbsite belt contain less gibbsite and more kaolinite. Interlayered vermiculite and illite (including intensely weathered muscovite) may be detected, but they occur as minor constituents. The limits of gibbsite bearing soil are gradational.

Lateritic and near-lateritic soils outside the gibbsite soil belt generally contain a minor amount of gibbsite, but in many soils the gibbsite content is below detection limits. Kaolinite is usually the dominant mineral in the clay fraction, but interlayered vermiculite and/or illite are usually present (Fiskell and others, 1959). The lateritic soils outside the gibbsite belt are similar to the gibbsite soils in physical appearance and occurrence, being restricted to upland terrace soils. The main difference is the mineral content of the clay fraction. The horizontal zoning of gibbsite and kaolinite is illustrated in Table 2.

ORIGIN OF GIBBSITE

Five requirements needed for the formation of bauxite are: (1) aluminous source material; (2) good drainage; (3) humid tropical or subtropical climate; (4) time; and (5) protection from erosion (Clarke, 1966, p. 904-905). The gibbsite soils are discussed with respect to these five requirements.

Aluminous Source Material

Gibbsite can be formed from almost any aluminous material as an end product of laterization after an infinite time under ideal conditions (Harder, 1952, p. 35-36), but with limited time and marginal conditions for laterization, aluminous source material is a controlling factor. Three possible sources of alumina are considered. They are: (1) early Eocene bauxite deposits such as those at Eufaula, Alabama (Warren and Clark, 1965), Andersonville and Warm Springs, Georgia; (2) clay in Coastal Plain sediments that were eroded, redeposited with sand, and laterized; and (3) weathered feldspathic debris eroded from the igneous-metamorphic complex of the Piedmont and Blue Ridge physiographic provinces.

No evidence was found during the investigation of the Eufaula bauxite district to indicate that eroded bauxite deposits could be a source of gibbsite in soil. Bauxite fragments or pebbles were not found in the soil or terrace gravel associated with lateritic soil. The gibbsite soils are separated from the bauxite horizon by unweathered marine sediments. This is illustrated schematically in Figure 3. The northern end of the Eufaula bauxite district has been deeply eroded and many bauxite deposits have been partially removed by erosion, but they are below the gibbsite soil zone.

Weathered feldspathic fragments and mica crystals in the fluvial sediments containing gibbsite soil indicate the crystalline igneous-

Table 2. Mineral estimates of lateritic soil samples collected from B and C soil horizons, reported on quartz-free basis (in percent).

Sample number	Gibbsite	Kaolinite	Hematite	Interlayered vermiculite and/or illite	Remarks
Mis-1	10	50	15	20	West of gibbsite belt in transition zone
Ala-6	20	55	15	5	Western end of gibbsite belt.
Fla-6	20	60	15	--	Northern edge of gibbsite belt.
Fla-5	5	60	10	20	East of gibbsite belt.
Mis-2	--	50	15	30	High terrace north of gibbsite belt.
Ala-7	5	55	15	20	High terrace, about 40 kilometers north of gibbsite belt.
Ala-8	5	50	15	25	High terrace 140 kilometers from gibbsite belt.
Ga-1	--	50	15	30	High terrace, northeast of gibbsite belt.

metamorphic complex is the probable ultimate source of the sediments. Plagioclase feldspar weathers directly to gibbsite in the humid tropical climate and orthoclase feldspar weathers to kaolinite and on continued desilication may be changed to gibbsite (Harrison, 1934). Sufficient feldspathic debris has been eroded and redeposited with the fluvial deposits to account for all high alumina minerals in the gibbsite soil.

The contact between the fluvial deposits containing gibbsite soil and the underlying sediments is an erosional disconformity that indicates channel cutting and filling. Clays and sand eroded along the channel would be redeposited with other fluvial sediments. Although the igneous-metamorphic complex of the Piedmont uplands is the most probable ultimate source of the fluvial deposits containing the gibbsite soil, the streams during the Pliocene time crossed sedimentary outcrops of Cretaceous and Paleogene age. Clay from Coastal Plain sediments must be considered a partial source for the gibbsite and associated high-alumina minerals, but the importance of reworked clay is hard to evaluate.

The table below gives mineral estimates on a clay bed from the top of Miocene age sediments in southwestern Alabama, sample Ala-5, and from weathered granite gneiss, Ala-9, from the Piedmont physiographic province. The table is in percent and was computed on a

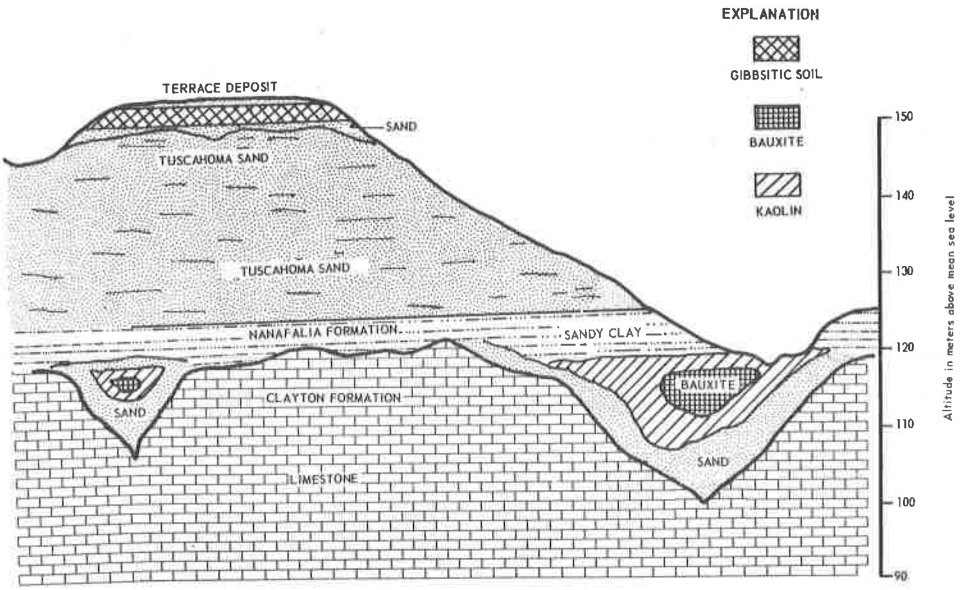

Figure 3. Sketch showing geologic and topographic relationship between gibbsite soil and bauxite deposits in Eufaula bauxite district, Alabama.

quartz-free basis.

Sample Number	Ala-5	Ala-9
Gibbsite	--	35
Kaolinite	30	40
Hematite	--	10
Interlayered vermiculite and/or illite	45	10
Montmorillonite	20	--

Time and Climate

The time in which laterization could have occurred extended from late Miocene to Pleistocene and may have included early interglacial intervals. The fluvial deposits containing the lateritic soil disconformity overlie formations ranging in age from middle Cretaceous through middle Miocene, but the marine and river terraces formed during the latter part of the Pleistocene do not show evidence of intense laterization. This places definite time limits on the period of possible laterization.

Fossil evidence indicating climatic conditions is rare in lateritic exposures because most organic remains are destroyed by the intense oxidization and leaching that produce laterization. However, recently discovered remains of crocodiles and camels in the Pliocene sediments near Mobile, Alabama, indicate a warm climate (Lamb, 1968).

Drainage and Protection from Erosion

The lateritic soil deposits have not been destroyed by mechanical erosion because they are in flat areas with excellent subsurface drainage. The soils are porous and there is very little surface runoff. Although vegetation has inhibited soil erosion, the deposits have been dissected by streams and completely removed from some areas, but mechanical erosion has been restricted largely to stream valleys.

SUMMARY AND CONCLUSIONS

The gibbsite soils were formed by lateritic weathering in clastic continental sediments that have been mapped as high terrace and Citronelle Formation. Weathered rocks containing plagioclase from the igneous-metamorphic complex north of the gibbsite soil zone were probably the main source of alumina for the gibbsite. Weathering occurred both in the crystalline rocks prior to erosion and in the clastic soils after fluvial deposition.

Laterization occurred mainly during the Pliocene because sediments containing the gibbsite soil unconformably overlie rock units of Miocene age, and the low terraces of Pleistocene age are not laterized. The lateritic soils are restricted to the flat uplands underlain by porous sands where there is excellent drainage. With forest cover and flat, well drained soils there was little or no mechanical erosion and insoluble products of weathering were not removed.

During laterization, silica and cations such as calcium, sodium, and magnesium were leached by rain water moving downward through the soil. Aluminum, iron, and silica were leached from the A soil horizon and precipitated in the B and C soil horizons as gibbsite, hematite, and kaolinite. Interlayered vermiculite, judging from field relationship, may be a weathering stage between illite or montmorillonite and kaolinite.

Quartz and gibbsite are not stable in the same environment, but they occur together in exceptionally well drained deposits. Apparently, the gibbsite soils have been above the water table since formation. Kaolinite with no gibbsite occurs in the sandy soils that have been subjected to flooding or water saturation for any extended period.

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ORGANIC CARBON IN SEDIMENTS OF THE NORTH CAROLINA CONTINENTAL RISE

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ABSTRACT

The organic carbon content of 208 samples from 18 piston cores on the North Carolina continental rise was investigated. Length of time of exposure of the sediment to oxidizing conditions appears to be the single most important factor causing variation of organic carbon content within single cores and between different portions of the continental rise. Exposure time appears to be a function of sedimentation rate and the importance of the contour current contribution.

INTRODUCTION

The purpose of this investigation is to characterize the organic carbon content of continental rise sediments and to determine the factors responsible for variation in abundance of organic carbon with core depth. The North Carolina continental rise is ideal for this study because detailed bathymetry is available (Newton and Pilkey, 1969) and much is already known about local sediments and sedimentary processes (Heezen, et. al., 1966; Rona, 1969; Golden, 1970; and Field and Pilkey, in press).

A large amount of data on the organic carbon content of marine sediments has been gathered since the publication of Trask's (1939)

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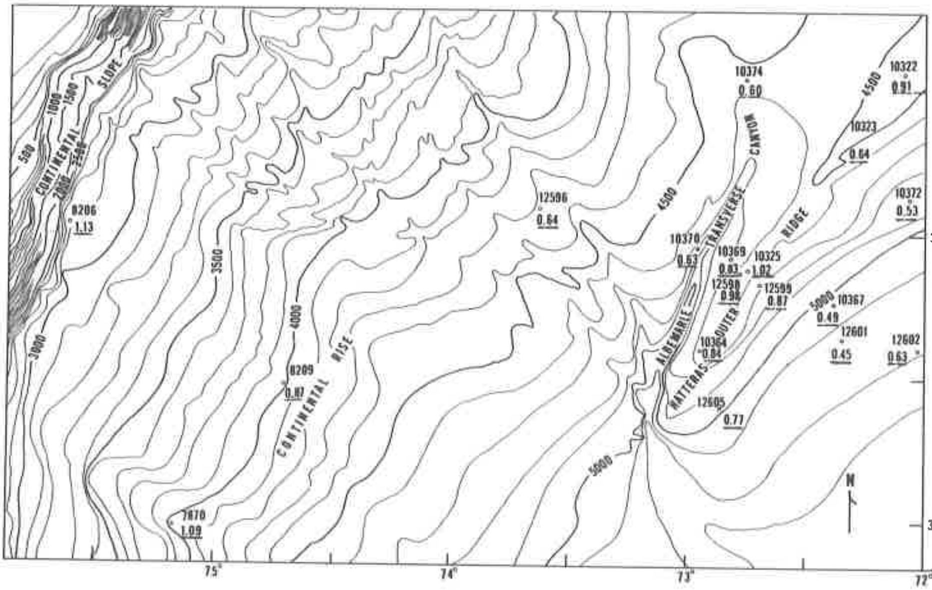

Figure 1. Map of the North Carolina continental rise study area showing sample points. Two numbers are shown next to each coring location. The upper of 4 or 5 digits is the R/V EASTWARD station number. The lower number which is underlined is the average percent organic in each core. Depth contours are in meters.

paper on the subject. Most of the available information is concerned with continental shelf or nearshore sediments (e. g., Bordovsky, 1965; Bryne and Emery, 1960; Gorsline, 1963; Kofoed and Gorsline, 1963; Keller and Richards, 1967). Observations of the distribution of organic carbon in continental slope and deeper environments include those of Gross (1967), Arrhenius (1963), Hobson and Mensels (1969), and Nino, Emery, and Kim, (1969).

A wide variety of techniques have been used to analyze organic carbon content and as a result values obtained by different investigators are often not comparable in detail. It is apparent, however, that the highest concentrations of organic carbon are associated with the finest grain size sediments, probably as a function of grain surface area available for absorption and because the organic material is itself, typically fine grained.

In the present study, organic carbon values were determined by the back-titration method of Allison (1935). No correction factors (as discussed by Nino, et al., 1969) have been applied to these data. It is assumed that errors inherent in the method apply to the whole sample suite and do not influence the relative concentration trends significantly.

Figure 1 is a map of the study area showing sample locations.

All samples are Ewing Piston cores obtained from Duke University's R/V EASTWARD. The 18 cores range in length from 6 to 12 meters. A total of 208 organic carbon analyses were performed at varying subsample intervals in the piston cores.

North Carolina continental rise sediments are dominantly silts and clays. Sand layers are thin and infrequent except in the Hatteras canyon system and on the landward side and crest of the Hatteras outer ridge. Here, turbidity current deposits form a significant portion of the sedimentary column.

Organic carbon analyses were performed mainly on the lutite sections of the cores. What proportion of this material has been derived from turbidity currents, contour currents or hemipelagic sedimentation is not known. Abundant mottling, however, indicates that turbidity currents are not likely as important overall as the latter two sedimentation mechanisms.

Acknowledgments

This study was supported by a grant from the National Science Foundation. It is part of a continuing investigation of the Hatteras outer ridge. The writers gratefully acknowledge the biological oceanographic program of the Duke University Marine Lab for the use of NSF supported R/V EASTWARD. K. S. Rodolfo and Graham Giese read the manuscript and offered many helpful suggestions.

DISCUSSION

Areal Distribution of Organic Carbon

The histogram of Figure 2 shows the frequency distribution of the organic carbon content of North Carolina continental rise sediments. Absolute values of organic carbon in the cores range from greater than 3 percent to below detection (3 percent is the maximum organic carbon percentage determinable by the method used). The modal frequency class is 0.8 percent to 1.00 percent organic carbon.

Average organic carbon contents of each core given in Figure 1 reveal several regional trends. Cores 8206 and 7870 on the upper rise contain the highest average values of 1.13 to 1.09 percent. Most of the other cores contain considerably less organic carbon with the exception of the cores clustered on or near the crest of the Hatteras outer ridge. Those cores from locations deeper than 4700 meters on the seaward flank of the Hatteras outer ridge contain significantly smaller average amounts of organic carbon than cores from any other rise area.

Sediments on the ridge and its flanks are probably deposited by a combination of pelagic mechanisms plus bottom currents and turbidity currents (Heezen _et al._, 1966; Rona, 1969; Field and Pilkey, in press).

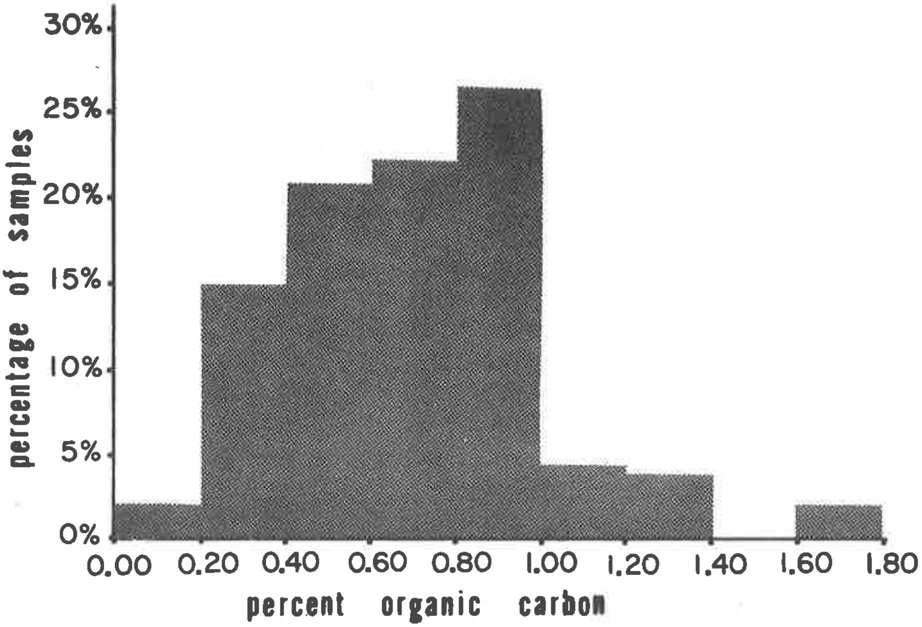

Figure 2. Histogram showing the frequency of distribution of the percentage of organic carbon in 208 continental rise samples. Not shown are samples containing greater than 2 percent carbon.

On the landward side and crest of the ridge, graded sand layers, some with coarse shallow water material indicate that turbidity current deposition is important. On the other hand, the ridge blocks many (but not all) of the turbidity currents coming down the Albemarle Transverse Canyon as evidenced by the reduced numbers of sand layers on the seaward ridge flank. The net result is an overall lower rate of sedimentation on the seaward flank of the Hatteras outer ridge relative to the landward side.

It is proposed that the difference in average organic carbon content of cores on either side of the ridge may be due in part to this difference in sedimentation rate. That is, on the seaward side of the ridge more average time has been available for oxidation of material prior to burial; hence, the lower amounts of organic carbon.

Another possibility, that is difficult to confirm with available data, is that contour current derived sediment is relatively more important on the seaward side of the ridge than on the landward side. Such sediment may have traveled great distances (Heezen, et al., 1966) allowing considerable time for oxidation of organics. A possible indication that contour currents are most important on the seaward side of the ridge is the abundance of diatoms in poor condition owing to solution

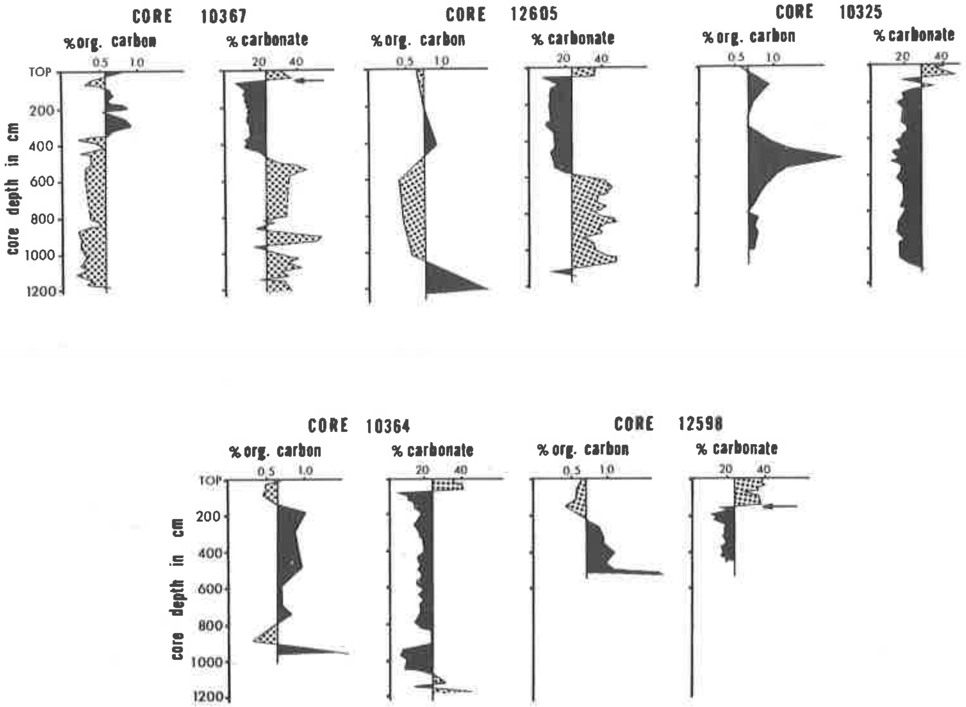

Figure 3. Plot of percent $CaCO_3$ and percent organic carbon with core depth for five Hatteras outer ridge cores.

effects (Golden, 1970). Because siliceous diatom frustules are relatively insoluble in sea water, their poor physical condition may be a reflection of long exposure time over long distances of transport. The fact that seaward sediments generally contain much larger percentages of partially dissolved diatoms than found on the landward side favors the idea of a relatively large bottom current sediment contribution here.

Variation of Organic Carbon with Core Depth

Petrographic (smearslide) observations of several hundred subsamples indicate that coccoliths are the principal form of $CaCO_3$ in the lutite sequences of these cores. Within the top 200 cm of each core, a sharp increase in carbonate percentage was observed. Constituent studies of these samples show that the increase corresponds to a sudden increase in coccoliths and a decrease in fine non-carbonate grains. This relatively abrupt change of sediment type, probably accomplished over a period of about 1,000 years is attributed to changing sea level. Radiocarbon dates in cores 10367 and 12598 of 8535 $\pm$ 180 years B. P. and 8500 $\pm$ 190 years B. P., respectively, (arrows in Figure 3) were obtained for this sediment boundary. Based on Currays (1960) sea level curves this is a reasonable date for submergence of the adjacent outer shelf by rising sea level. When the sea level rose over the edge of the present continental shelf, the rate of direct supply of terrigenous sediment to the Hatteras Canyon system in particular and the continental rise in general should have been strongly reduced. That is, no longer were the rivers emptying directly onto the continental slope but instead it is probable that estuarine systems developed trapping much of the river contribution. Meanwhile, the pelagic contribution to the rise, including coccoliths, should have become relatively more important. Thickness of the last glacial and the post-glacial sedimentation intervals indicate (as would be expected from the aforementioned model of sea level influence on continental rise sedimentation) that sedimentation rates were higher during the last glacial (low $CaCO_3$ sequences) than in post glacial time. Thus $CaCO_3$ abundance in Carolina rise sediments is inversely related to the rate of sedimentation.

Figure 3 is a plot of abundance of $CaCO_3$ and organic carbon in 5 cores, all from the Hatteras outer ridge area. Abundances are plotted with respect to an arbitrary line chosen to best display correlations between the two parameters. It is clear from Figure 3 that in at least 4 of the 5 cores shown, there is a strong inverse relationship between percentages of $CaCO_3$ and organic carbon. That is, the lutite sequences of relatively high $CaCO_3$ correspond closely to sequences of relatively low organic carbon content. In other words, low organic carbon content sequences are characterized by low sedimentation rates (and thus also relatively small amounts of terrigenous clays with organicabsorption sites).

No consistent decrease in organic carbon content with core depth was observed and it is assumed that diagenetic loss is less significant than abundance variations related to sedimentation rate.

CONCLUSION

The length of time of exposure of continental rise bottom sediment to pre-burial oxidizing conditions appears to control variations in organic carbon content on the rise. Exposure length is a function of sedimentation rate and the importance of the contour current contribution.

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POSSIBLE PETROGENIC RELATIONS OF THORIUM, URANIUM, AND Ce/ (Nd+Y) IN DETRITAL MONAZITE FROM SURRY AND STOKES COUNTIES, NORTH CAROLINA*

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ABSTRACT

One new analysis of detrital monazite from Stewarts Creek in Surry County, N. C., and two new analyses of detrital monazite from Big Creek in Surry and Stokes Counties, N. C., disclosed 4.9 - 6.1 percent ThO_2 and 0.71-1.9 percent U_3O_8. The two samples of monazite from Big Creek are among the most uranium-rich monazites on record from the United States. The atomic ratios Ce/(Nd+Y) range from 1.77 to 2.12 and are among the lowest reported from the southeastern States, indicating that the rare earths in these monazites tend to be unfractionated.

The Stewarts Creek monazite is not as rich in uranium as are the Big Creek monazites, which strongly resemble monazite from the Cherryville Quartz Monzonite near Shelby in Cleveland County, N. C. Considerations of the compositions of the detrital monazites and the reported paragenetic relations of monazite in the postmetamorphic plutons in Surry and Stokes Counties raise the question of whether these plutons might be the root zones of Paleozoic felsic volcanic centers. Answers to this question might be found in the lithology of rocks immediately underlying Triassic sediments exposed in the basins to the east

*Publication authorized by the Director, U. S. Geological Survey.

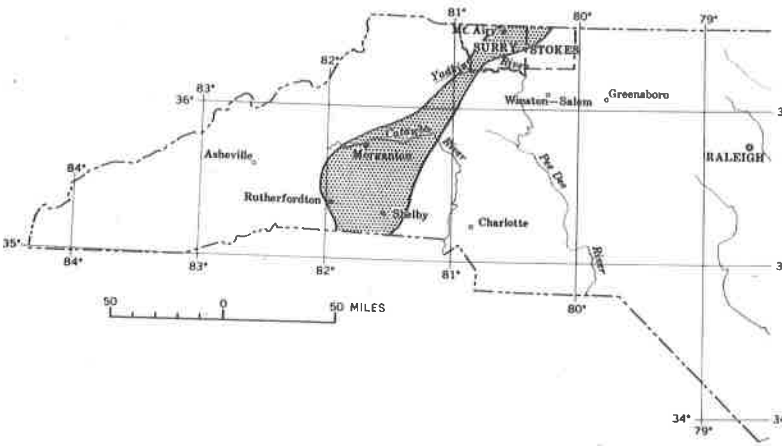

Figure 1. Index map showing location of Surry and Stokes Counties, North Carolina, and the western monazite belt (patterned) in the Piedmont of North Carolina. Outline of monazite belt is adapted from Mertie (1953, pl. 1).

of Mount Airy, in the composition of rock fragments in the Triassic conglomerate, and in the shape of the plutons as defined by structural features and gravity profiles.

INTRODUCTION

Previous Work

Surry and Stokes Counties, N. C., lie athwart the western monazite belt in the Piedmont of the southeastern States (Mertie, 1953, pl. 1) where that belt passes northeastward into Virginia (Figure 1). Monazite in granite near Mount Airy, Surry County, was first discovered by J. B. Mertie, Jr., in 1945 (Mertie, 1953, p. 24-25). He made further discoveries of monazite in granite and granite gneiss in the same area during 1952. Granitic rocks in the Sandy Ridge pegmatite district, Stokes County, were reported by Griffitts and others (1953, p. 143-146) to contain sporadic monazite, but specific localities were not identified. In a detailed description of the rock quarried at Mount Airy, Dietrich (1961, p. 7-8, 10, 12, 33-35) discussed accessory monazite as a product of early crystallization. No reports of the occurrence of monazite in the schists and quartzites of Surry and Stokes Counties are known to us.

The distribution of detrital monazite in Surry and Stokes Counties was studied in 1952 by A. M. White and G. A. Miller of the U. S. Geological Survey, who panned concentrates from 40-pound samples of

Table 1. Rock Units in the Source Areas of Monazite-Bearing and Monazite-Free Concentrates from Surry and Stokes Counties, N. C.

[Rock units adapted from Butler and Dunn, 1968, fig. 8]

Rock unit dominant in distributive province of concentrate[1]	Number of concentrates from streams draining rock unit		Percentage of monazite-bearing concentrates from unit	Percentage of monazite in monazite-bearing concentrate	
	Monazite-free	Monazite-bearing		Range	Mean
mag	3	1	25	-	4
mag+hms	1	4	80	2-12	7
fg+bg	1	3	75	tr-2	tr
hms	4	8	75	tr-4	1
gs	9	2	18	tr-2	1
gs+hms	2	-	0	-	-
gs+bgn	7	-	0	-	-
gs+fmg	1	-	0	-	-
fmg	3	-	0	-	-
fmg+hms+gs	2	2	50	tr-2	1
fmg+gs+bg	-	2	100	tr-2	1
bgn	19	5	21	-	tr
bg	5	6	55	tr-1	tr
bg+gs	2	3	60	tr-8	3
unmapped	6	2	25	-	tr

[1]Explanation of symbols:

mag = Granodiortie at Mount Airy and other postmetamorphic plutons.
fg = Premetamorphic foliated granitic gneiss.
hms = Hornblende schist, mica schist and gneiss, and quartz-feldspar gneiss in approximately equal amounts.
gs = Garnetiferous, graphitic muscovite-chlorite schist and phyllite, locally containing quartz-rich phyllite.
fmg = Feldspathic gneiss, mica-garnet schist with thin amphibolite layers; laterally equivalent to gs.
bgn = Biotite gneiss, commonly garnetiferous; may be equivalent to bg.
bg = Biotite and hornblende gneiss and schist; augen gneiss.
tr = Trace, less than 1 percent.

stream sediments at 103 localities (Overstreet, 1967, p. 208-209). Thirty-eight concentrates were monazite bearing, but only 17 contained 1 percent or more of monazite (Figure 2). At the time of that study maps of the regional geology of the counties were lacking, but recently this deficiency has been partly remedied (Espenshade, 1967; Butler and Dunn, 1968, fig. 8). Some geologic controls on the distribution of detrital monazite in Surry and Stokes Counties can be evaluated from the map of Butler and Dunn.

The amount of thorium and uranium in monazite from Surry and Stokes Counties has not previously been reported.

Present Investigation

The present investigation is part of a continuing study of selected chemical characteristics of monazite and other resistate detrital minerals from the southeastern United States. The purpose of this study is to explore the uses of selected detrital minerals and their chemical characteristics as clues to geologic features in areas of deeply weathered crystalline rocks. In this paper new chemical data for three monazites from Surry and Stokes Counties, N. C., are compared with similar data for other monazites from the southeast, and petrogenic interpretations of the granodiorite at Mount Airy are inferred. We hope that these speculative interpretations will generate further interest in the origin of this rock. Some possible further studies are identified.

Sources of the Monazite

Concentrates from the streams draining the geologic units identified by Butler and Dunn (1968, fig. 8) as hms, bg, and mag are more likely to be monazite-bearing than are concentrates from other source areas, and they tend to contain more monazite than do concentrates from other units (Table 1).

Many of the distributive provinces represented by the concentrates contain several kinds of rock; therefore, various combinations of the rock units are considered to be dominant in some of the provinces, as shown in Table 1. The most regularly monazite-bearing distributive province, and the one from which concentrates have the greatest percentage of monazite, is a combination of the unit hms with the unit mag (Table 1).

Concentrates from streams that drain areas underlain by combinations of the units bg and gs are among the richer, averaging 3 percent monazite (Table 1). Curiously, streams that drain the combined units are a better source for monazite than streams that drain the component units individually, but more monazite seems to come from the bg unit.

The unit bgn possibly is the equivalent of unit bg according to

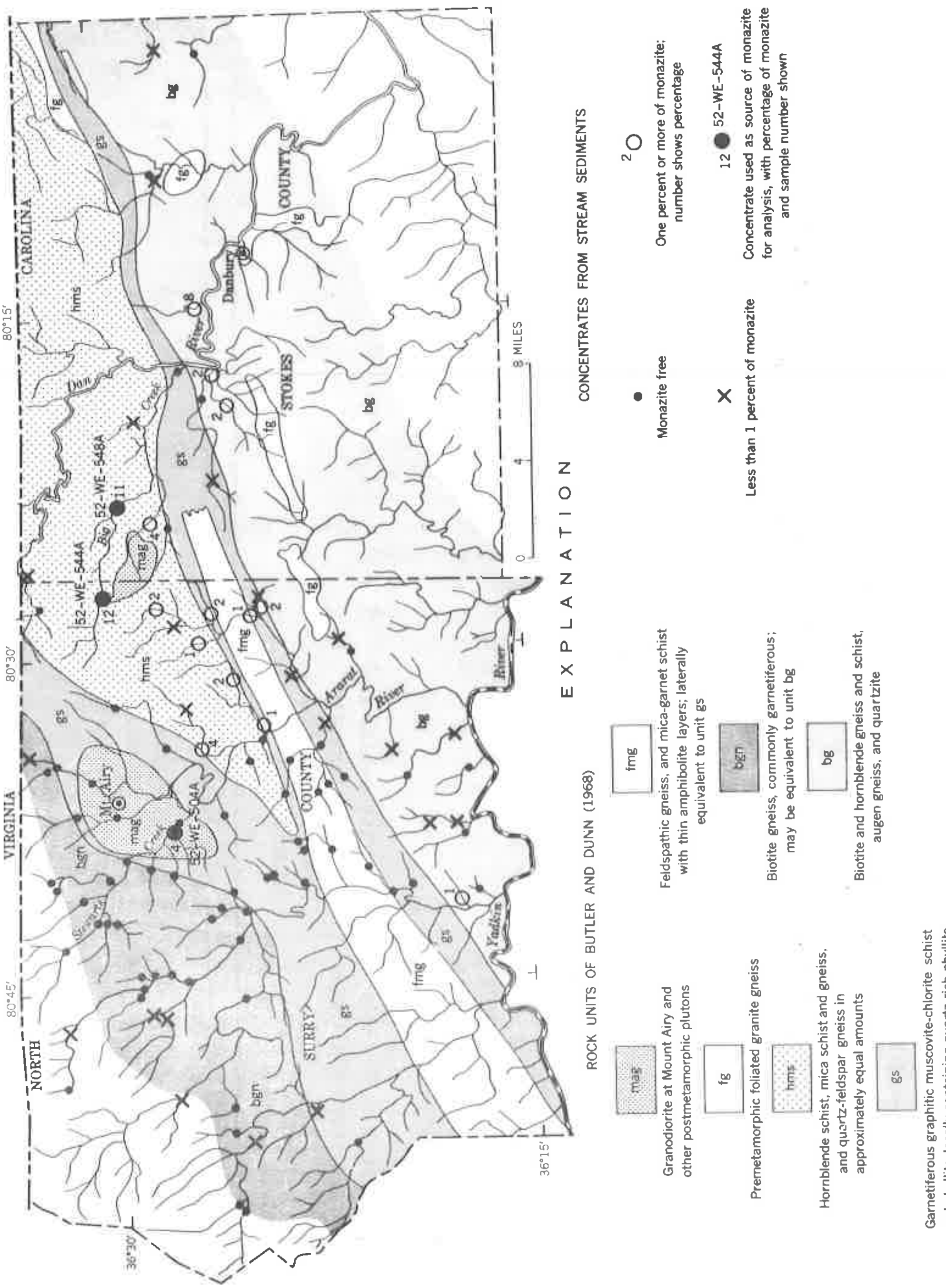

Figure 2. Locations of concentrates with detrital monazite in Surry and Stokes Counties, North Carolina, compared to the general distribution of rock units outlined by Butler and Dunn (1968, fig. 8).

Table 2. Locality, Source, and Composition of Samples of Detrital Monazite from Surry and Stokes Counties, N. C.

[Analyses of ThO_2 and U_3O_8 by J. J. Warr, Jr., U. S. Geological Survey, October 8, 1964; Ce/(Nd+Y) by Sol Berman, U. S. Geological Survey, December 30, 1968.]

Sample No.	Locality	Source	ThO_2 (percent)	U_3O_8 (percent)	U_3O_8/ThO_2	Ce/(Nd+Y)
		Surry County				
52-WE-504A	Southwest tributary to Stewarts Creek 2.5 miles upstream from confluence of Stewarts Creek with the Ararat River	Granite dominant.	6.1	0.71	0.11	2.12
52-WE-544A	Big Creek 0.75 mile upstream from the County line.	Biotite schist and granite; granite dominant; staurolite in schist.	5.9	1.7	.29	1.88
		Stokes County				
52-WE-548A	Big Creek 3. miles downstream from County line.	Biotite schist and granite; granite dominant; staurolite in schist.	4.9	1.9	.38	1.77

Butler and Dunn (1968, fig. 8). Where these two units are monazite bearing they tend to contribute similar percentages of monazite to stream concentrates, but the unit bg is about twice as frequently monazite bearing as the unit bgn. The greater frequency of monazite-bearing samples from streams that drain areas underlain by bg suggests that bg is of higher metamorphic rank than bgn (Overstreet, 1967, p. 17).

The three samples of detrital monazite selected for analysis are from streams that drain units mag and hms + mag. At the three sample localities (Table 2) the dominant source rock is granite. For sample 52-WE-504A the granite is interpreted to be unit mag of Butler and Dunn (1968, fig. 8), and the monazite in the sample is inferred to have come largely, if not entirely, from this unit. This is the rock called granodiorite by Butler and Dunn, leucogranodiorite by Dietrich (1961), and granite by Mertie (1953, p. 24-25). Sample 52-WE-544A is from a drainage basin underlain by a mixture of units mag and hms. Because granite is dominant at this sample locality, the monazite may be mainly derived from unit mag instead of unit hms and also may be of igneous origin. The influence of unit mag on the distribution of monazite in the basin of Big Creek appears to be shown by the higher percentages for monazite in concentrates from streams in the immediate area of unit mag than in concentrates from streams elsewhere in unit hms.

ANALYSES OF MONAZITE

The monazite was prepared and analyzed for thorium and uranium by the same procedures described in an earlier report (Overstreet and others, 1969, p. 65-66). Thorium was determined spectrophotometrically with thorom (Fletcher and others, 1957). Checks of the thorium determinations were made by another spectrophotometric method using arsenazo III as the reagent (May and Jenkins, 1965). Uranium was determined fluorometrically after isolation of the uranium and fusing with a flux containing NaF (Grimaldi and others, 1954). The percentages of cerium, yttrium, and neodymium were determined spectrochemically by Sol Berman, U. S. Geological Survey, following procedures described by Murata and others (1953, p. 292).

The results of the analyses are given in Table 2. The two samples from Big Creek are among the most uranium-rich monazites on record from the United States, and all three monazites disclose ratios for Ce/ (Nd+Y) that are among the lowest on record for the southeastern States.

GEOLOGIC RELATIONS OF THE COMPOSITION OF THE MONAZITE

The available data on the abundances of thorium and uranium in

Table 3. Percentages of ThO_2 and U_3O_8, Ratios of U_3O_8/ThO_2, Ce/(Nd+Y), and Estimated Σ La+Ce+Pr of Monazites from Granitic Rocks in the Southeastern United States [Leaders (---) = no data]

Locality by State and county or counties	Source of monazite (Saprolite unless otherwise stated; Nos. in parentheses are analyses plotted on fig. 3)	Number of analyses	ThO_2 Percentage Range	ThO_2 Mean	U_3O_8 Percentage Range	U_3O_8 Mean	U_3O_8/ThO_2 Range	U_3O_8/ThO_2 Mean	Ce/(Nd+Y) Range	Ce/(Nd+Y) Mean	Estimated Σ La+Ce+Pr from the mean of Ce/(Nd+Y) (after the method of Murata and others 1957, fig. 5)	Reference(s)
					POSTKINEMATIC GRANITIC ROCKS Piedmont province							
North Carolina												
Surry----------	Granodiorite, detrital, Stewarts, Ck. (1)----------	1	-----	6.1	-----	0.71	-----	0.11	-----	2.12	74	This report.
Surry & Stokes--	Detrital, Big Ck. (2-3)----------	2	4.9 - 5.9	5.4	1.7 -1.9	1.8	0.29 -0.38	.33	1.77-1.88	1.82	69	Do.
Cleveland-------	Cherryville Quartz Monzonite (4)-------	3[a]	5.6 - 6.9	6.4	-----	2.34	-----	.42	2.19-2.49	2.34	76	Overstreet (1967, tab. 68).
	Cherryville Quartz Monzonite, detrital (5)----------	1	-----	5.9	-----	1.4	-----	.23	-----	----	--	J.J.Warr, Jr. (written commun. 1964).
					SYNKINEMATIC GRANITIC ROCKS Blue Ridge province							
North Carolina												
Macon & Jackson-	Whiteside Granite (6-9)----------	4	4.3 - 5.7	5.0	0.13 -0.35	0.24	0.02 -0.07	0.04	2.77-3.40	2.99	81	Overstreet (1967, tab. 60)
					Piedmont province							
North Carolina												
Rutherford------	Toluca Quartz Monzonite (10)---------	2[a]	4.3 - 8.8	6.5	-----	0.76	-----	0.17	2.65-2.83	2.74	79	Overstreet (1967, tab. 61); Murata & others (1957, tab.3)
Cleveland-------	Toluca Quartz Monzonite (11-12)------	23[b]	4.3 - 7.3	6.1	0.35 -1.10	.74	0.05 -0.25	.14	2.12-2.80	2.49	78	Overstreet (1967, tab. 66); Murata & others (1957, tab.3)
	Toluca Quartz Monzonite, detrital (13-21)----------	9	5.08- 7.84	6.9	.28 - .41	.35	.039- .079	.055	-----	----	--	Overstreet & others (1963a, tab. 5). Mertie (1953, p.12).
South Carolina												
Cherokee--------	Toluca Quartz Monzonite, detrital (22-23)----------	2	6.21- 6.45	6.33	.22 - .23	.22	.035- .036	.036	-----	----	--	Do.
Georgia												
Oconee---------	Detrital (24-25)----	2	6.0 - 6.1	6.0	.43 - .44	.44	.07 - .07	.07	-----	----	--	Overstreet & others (1969, tab. 2).

GRANITE OF UNCERTAIN RELATIONS
Piedmont province

Virginia												
Chesterfield----	Granite-------------	1	-----	6.8	-----	----	-----	----	-----	2.20	75	Murata & others (1957, p. 149).
Georgia												
Spalding--------	Granite at Zetella (26)----------------	1	-----	4.42	-----	0.26	-----	0.059	-----	----	--	Mertie (1953, p.12)
	Detrital (27-31)----	5	5.4 - 5.7	5.5	0.13 -0.24	.17	0.02 -0.04	.03	-----	----	--	Overstreet & others (1969, tab. 2).
Troup-----------	Detrital (32)-------	1	-----	4.1	-----	.33	-----	.08	-----	----	--	Do.
PEGMATITE AND VEIN QUARTZ Blue Ridge province												
North Carolina												
Mitchell--------	Pegmatite, Deer Pk. No. 5 mine (33)-----	1c/	5.48-5.54	5.51	-----	0.02	-----	0.0036	-----	----	--	Bliss (1944, p. 327, 329).
Yancey----------	Pegmatite, No. 20 mine----------------	1	-----	8.18	-----	----	-----	----	-----	1.34	61	Murata & others (1957, p. 149).
Madison---------	Pegmatite, Mars Hill (34-35)-------------	d/ 7	5.06- 7.0	6.2	0.019-0.042	0.029	0.003-0.006	.004	-----	----	--	Marble (1936, p.456-457); Lane (1937, p. 58); Pratt (1916, p. 48); Schaller (1933).
Piedmont province												
North Carolina												
Cleveland and Rutherford----	Pegmatite of Toluca Quartz Monzonite----	46	3.8 -11.2	6.1	-----	----	-----	----	2.27-2.90	2.57	78	Overstreet (1967, tab. 61, 63); Murata & others (1957, tab. 3).
Cleveland-------	Quartz vein possibly related to the Cherryville Quartz Monzorite-----------	1	-----	6.1	-----	----	-----	----	-----	2.27	75	Overstreet (1967, tab. 66, p. 205).
Virginia												
Amelia----------	Pegmatite near Amelia Court House (36)----	1e/	7.21- 7.89	7.55	-----	0.2	-----	0.026	-----	1.73	68	Lane (1934, p.28); Murata & others (1953, p.294); Overstreet (1967, p. 268).

a/One determination of U_3O_8.

b/Three determinations of U_3O_8. Ratio from tables 4 and 5, Overstreet and others (1963a).

c/Two determinations of ThO_2 and one of U_3O_8. Ratio U_3O_8/ThO_2 from mean of ThO_2.

d/Seven determinations of ThO_2 and four of U_3O_8.

e/Seven determinations of ThO_2 and two of U_3O_8 have been reported (Overstreet, 1967, p. 267-268). For this table the latest analysis showing both ThO_2 and U_3O_8 (Lane, 1934, p. 8) is used for ratio of U_3O_8/ThO_2.

monazites from granitic rocks in the southeastern United States are summarized in Table 3, together with the ratios of U_3O_8/ThO_2 and Ce/(Nd+Y), and the estimated Σ La+Ce+Pr, for comparison with the composition of monazites from Surry and Stokes Counties. Most of the analyses given in Table 3 are of monazite from saprolite or unweathered rock where the geologic source is known. Analyses do not exist for nondetrital monazite from South Carolina, and few are available from Georgia; therefore, analyses of detrital monazite from petrologically uniform provenance are introduced to extend the geographic coverage to those states. Other analyses of detrital monazite are also listed for North Carolina.

Thorium and Uranium

The three samples of monazite from Surry and Stokes Counties, N. C., have an average tenor of 5.6 percent ThO_2, nearly identical to the average of 5.67 percent ThO_2 reported by Mertie (1953, p. 12) for 54 samples of detrital monazite from the western Piedmont of the southeastern States. The range of ThO_2 for the Stewarts Creek and Big Creek monazites is well within the ranges previously reported for other monazites from the southeast (Table 3). No real possibility exists for discriminating the monazites from many different rocks in the southeastern States on a basis of ThO_2 alone. However, a marked difference exists between the percentages of U_3O_8 in the Big Creek monazite and all other monazites in Table 3 except those from the Cherryville Quartz Monzonite. When the percentages of ThO_2 and U_3O_8 are plotted (Figure 3), three fields are disclosed that include the four categories shown on Table 4: (1) monazites from the postkinematic rocks; (2) monazites from the synkinematic rocks and granites of uncertain relations; and (3) monazites from the pegmatites.

Abundance of ThO_2 and U_3O_8 in the Big Creek monazites fall between those of the two monazites from the Cherryville Quartz Monzonite. The four points define an elongate uranium-rich field in Figure 3 characterized by percentages of ThO_2 similar to Mertie's regional average percentage of ThO_2 (Mertie, 1953, p. 12). This field is identified with four of the five monazites from postkinematic granitic rocks shown in Table 3. However, the Stewarts Creek monazite falls well outside this field in an area adjoining the uranium-rich parts of the synkinematic field occupied mainly by monazites from the Toluca Quartz Monzonite. Evidently, we have too few analyses of monazites from the postkinematic granitic rocks to define this field, but the association of the Big Creek monazites, in terms of ThO_2 and U_3O_8, with monazites from the Cherryville Quartz Monzonite is distinctive.

Monazites from the Toluca Quartz Monzonite and from granitic rocks in Oconee County, Ga., tend to occupy the thorium-rich part of the field for monazites from synkinematic granites. Monazites from the Whiteside Granite and the granites of uncertain relations typically

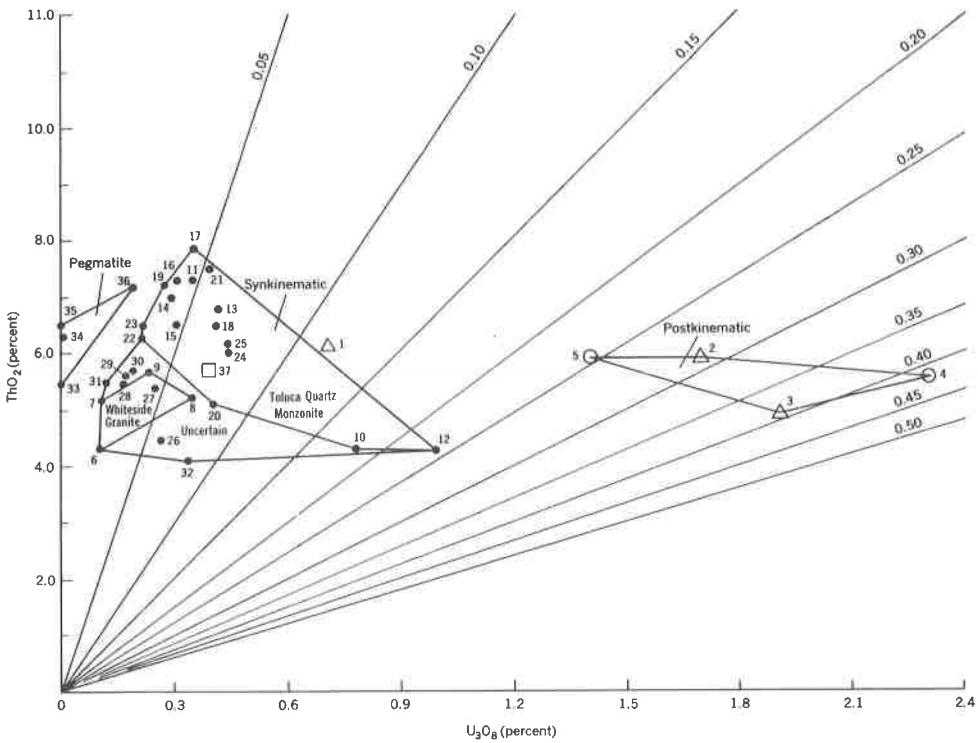

Figure 3. Thorium and uranium in monazites from the southeastern United States. Dots, 21 monazites from Table 3 with percentages of both ThO_2 and U_3O_8; triangle 1, Stewarts Creek monazite; triangles 2 and 3, Big Creek monazite; circles, monazite from Cherryville Quartz Monzonite; square, Mertie's (1953) average of 54 detrital monazites. U_3O_8/ThO_2 shown by radial lines.

are in the low-thorium parts of the field. These relations are thought to be dependent on regional differences of pressure and temperature at consolidation of the granitic rocks, as reflected by the metamorphic facies of the wallrocks (Overstreet, 1967, table 2). For the Toluca and the granitic rocks in Oconee County, Ga., emplacement and consolidation took place under conditions of the sillimanite-almandine subfacies of regional metamorphism, but the Whiteside Granite and the other Georgia granites shown in Figure 3 formed at the staurolite-kyanite subfacies and lower facies of regional metamorphism (Overstreet and others, 1969).

The average percentage of ThO_2 in the monazites from the southeastern States tends to decline with increasing CaO content of the whole rock (Figure 4), but the cause of this relation in the southeastern monazites is not clear. Averages of the values used in Figure 4 show:

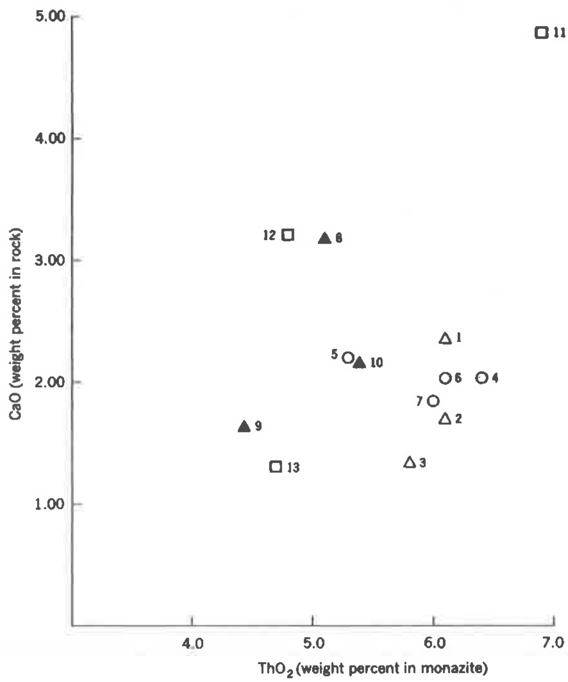

Figure 4. Relation between whole-rock contents of CaO and percentage of ThO_2 in monazite. Open triangles represent postkinematic granitic rocks: 1, Granodiorite, Mt. Airy, N. C., CaO from average of four values given by Dietrich, 1961, p. 28; 2, Do., CaO by U. S. G. S. from sample collected by J. B. Mertie, Jr., 1953; 3, Cherryville Quartz Monzonite, CaO by U. S. G. S. Circles represent synkinematic granitic rocks in sillimanite-almandine subfacies wallrocks: 4, Toluca Quartz Monzonite, average of 14 ThO_2 in Figure 3 and three determinations CaO; 5, Do., averages for two samples from type locality, Cleveland County, N. C.; 6, Do., average of 21 determinations of ThO_2 from Overstreet (1967, table 65), and three CaO; 7, Oconee County, Ga., CaO from Watson (1902). Solid triangles represent synkinematic granitic rocks in staurolite-kyanite subfacies wallrocks: 8, Whiteside Granite, CaO by U. S. G. S.; 9, Zetella, Spalding County, Ga., CaO by U. S. G. S. on sample collected by J. B. Mertie, Jr., 1953; 10, Spalding County, Ga., CaO average of three from Watson (1902). Squares represent metasedimentary rocks of sillimanite-almandine subfacies: 11, Biotite gneiss, Shelby quadrangle, N. C., from Overstreet and others (1963a, table 4; 1963b, table 8); 12, Biotite schist, do.; 13, Sillimanite schist, do.

	Percent	
	CaO	ThO_2
Synkinematic granites in sillimanite-almandine subfacies wallrocks	1.97	6.2
Postkinematic granites in staurolite-kyanite subfacies wallrocks	2.09	6.0
Synkinematic granites in staurolite-kyanite subfacies wallrocks	2.27	5.2

Lee and Bastron (1967, p. 355) observed a similar trend in monazites from a compositionally varied granitic stock in Nevada.

The pegmatitic monazites form a separate field (Figure 3) characterized by low percentages of U_3O_8 associated with high percentages of ThO_2. Analyses showing CaO in the host rocks are not available, but they doubtless would be among the lowest if they were diagrammed on Figure 4, and the CaO in the quartz vein for which a thorium-rich monazite is listed in Table 3, would certainly be the lowest in the group. Thus, for the pegmatitic monazites probably high values of ThO_2 are associated with low values for CaO.

The ThO_2 in monazite of metamorphic origin is plotted against whole-rock CaO in Figure 4 for comparison with monazites from granitic rocks. No effect of whole-rock CaO can be seen on the distribution of ThO_2 in these metamorphic monazites.

Uranium/Thorium Ratios

The U_3O_8/ThO_2 ratios (Table 2 and Figure 3) for detrital monazite from Surry and Stokes Counties are greater than the mean ratio reported by Mertie (1953, p. 12) for 54 specimens of detrital monazite from the southern States. Regional differences in the uranium/thorium ratios of monazite from granitic rocks in the southeastern States are such that these ratios decrease by almost half an order of magnitude between monazites from postkinematic granitic rocks and monazites from synkinematic granitic rocks, and decrease by nearly another order of magnitude to the monazites from pegmatites of the Blue Ridge. However, monazite from pegmatite in the Virginia Piedmont does not fit this pattern (Table 3). These differences tend to support the inference developed in a study of U_3O_8/ThO_2 ratios of Georgia monazites that a relation may exist between the rock in which monazite was formed and the uranium/thorium ratio of the monazite (Overstreet and others, 1969, p. 72). However, the data are too sparse and too few monazites of known geologic provenance have yet been analyzed for U_3O_8 for the petrogenic relations to be known.

Ce/ (Nd+Y) and Estimated Σ La+Ce+Pr

The ratio Ce/ (Nd+Y) of the monazites from Surry and Stokes Counties, N. C. (Table 2) was determined for use as an additional means of regional comparison, and the ratios were used as a ready way to obtain the quantity sigma (Σ). Sigma is the sum of the atomic percentages of the three most basic rare earths, La+Ce+Pr. It was introduced by Murata and others (1953, p. 296-299; 1957, p. 150) as a measure of the stage of fractionation of the rare-earth elements in cerium-earth minerals.

Their concept (Murata and others, 1953, 1957) assumes simple fractional precipitation in which less basic rare earths are precipitated preferentially in appropriate mineral phases, and the magmatic liquid becomes increasingly richer in the more basic rare earths. High values for Σ indicate highly fractionated monazites. The assumed unidirectional course of fractional precipitation could be reversed at any stage if minerals of the more basic rare-earth elements were to commence to crystallize. Inasmuch as a substantial part of the rare earths in a rock is in apatite, sphene, epidote, and other rock-forming minerals, the course of fractional precipitation of the rare earths in a magma also involves their entry into such minerals. The values of sigma reported by Murata and others (1957, table 3) for monazites from many sources range from 58.1 to 87. Those values above 80 they regarded as highly fractionated and values below 70 as relatively unfractionated. By comparison with their general values, the sigma of monazites from Big Creek (Table 3) is relatively unfractionated, and sigma of monazite from Stewarts Creek is intermediate.

Sigma of the monazite from Stewarts Creek (Table 3) more nearly resembles that of monazite from granite in Chesterfield County, Va., and of monazite from the Cherryville Quartz Monzonite and possibly genetically related vein quartz in Cleveland County, N. C., than that of monazite from other granitic rocks in the southeast. Sigma of the monazites from Big Creek most closely resembles that of monazite from pegmatites near Amelia Court House, Va. However, the ratio U_3O_8/ThO_2 of the Amelia monazites is more than an order of magnitude less than the ratio of the Big Creek monazites. When both sigma and U_3O_8/ThO_2 are compared, the closest counterparts to the Big Creek and Stewarts Creek monazites are those from the Cherryville Quartz Monzonite.

Using sigma as a measure of fractionation leads to an ordering of the monazites in Table 3 such that those classed as least fractionated (ones with lowest sigma) are from the Spruce Pine pegmatites, and those classed as the most fractionated are from the Whiteside Granite:

Source of Monazite	Sigma
Pegmatite, Spruce Pine	61
Pegmatite, Amelia Court House	68
Postkinematic pluton(?), Big Creek	69
Postkinematic granodiorite, Stewarts Creek	74
Granite of uncertain origin, Chesterfield County	75
Quartz vein possibly related to Cherryville Quartz Monzonite	75
Postkinematic Cherryville Quartz Monzonite	76
Synkinematic Toluca Quartz Monzonite	78
Pegmatite from Toluca Quartz Monzonite	78
Whiteside Granite	81

The only genetic groupings of monazites in the list are those from the Toluca Quartz Monzonite and its associated pegmatite, and those from the Cherryville Quartz Monzonite and the possibly related quartz vein. In the first of these pairs sigma is the same for the quartz monzonite and the pegmatite. In the second pair the trend of fractionation is in the reverse direction to that postulated by the general assumption of a simple fractional precipitation. Inasmuch as sigma for monazites from genetically related rocks fails to reflect the postulated trend in fractionation, it is unlikely that the regional range in values of sigma for these monazites can yet be used as a measure of the stage of fractionation of the monazites in the different rock units despite the apparent relation exhibited in the ranking listed above.

The order above is also matched in the examples of sigma given by Murata and others (1957, table 3): values for sigma from 58 to 69 are associated with seven of the nine analyses of monazite from pegmatite; values from 74 to 80 are associated with monazites from granite and quartz monzonite. This order, and that in the examples given by Murata and associates, follows the order predictable from observations that some pegmatitic monazites are more enriched in the heavier rare earths (have smaller sigma) than are monazites from the granitic source rocks (Vainshtein and others, 1956; Vlasov, 1966a, p. 244). Evidently some factor other than simple unidirectional fractional precipitation of the rare-earth elements in monazite is indicated. In that process the pegmatitic monazites might be expected to show the greatest compositional effect of continued fractional precipitation--that is, show the greatest sigma--if the pegmatites are correctly identified as late-stage products of magmatic crystallization. For the pegmatites related to the Toluca Quartz Monzonite, there are many examples of the pegmatite filling fractures in the parent quartz monzonite. Monazites from the parent quartz monzonite and later pegmatite have been shown by Murata and others (1957, p. 154-155) to have compositions in one locality that differ in the sense to be expected from the postulated process of fractional precipitation, and elsewhere to differ in the reverse sense. Average sigmas for monazites from each type of rock are, however, identical (Table 3).

The order of the sigmas listed above suggests that pegmatitic monazites tend to differ from granitic monazites generally in the reverse sense to that postulated by simple preferential fractional precipitation of the least basic rare earths. Murata and others (1959, p. 12) have shown that the order of precipitation of the rare earths may depend on their state of solution. If the rare-earth elements are present as simple cations, then the less basic rare earths form less soluble hydroxides and phosphates, and the simple fractional precipitation assumed above proceeds. If these elements are present instead in some coordination complex as anions, then the order of precipitation and the sense of sigma are reversed.

Important as the state of solution of the rare-earth elements may be in the crystallizing rock materials, the trend of precipitation may be dominated by the mineral phases that are taking up the rare earths. Data on this are very sparse for rocks from the southeastern States. To our knowledge the only study is by K. J. Murata (written commun., 1954) who investigated the distribution of rare-earth elements in biotite, garnet, and monazite in the Toluca Quartz Monzonite. Both the biotite and garnet were found to contain lanthanum and yttrium, and both minerals contained more yttrium in relation to lanthanum than did the monazite from the same rock. Selective deposition of the yttrium-earth elements in the silicate minerals as crystallization proceeded would result in the enrichment of the residual fluid in the cerium earths as the rocks differentiated. This process would lead to the same direction of increase in sigma as the concept of simple fractional precipitation of cations of the less basic rare earths.

The kind and degree of partition of the rare earths in such calcium-bearing minerals as apatite, epidote, and the plagioclase feldspars have yet to be studied for the southeastern granitic rocks. Until these data are available it is not known if apatite, epidote, and the feldspars contain the basic rare earths in greater or less proportion to the less basic rare earths than the proportions existing in analyzed monazites. Thus, the possible effect of the precipitation of the rare earths in these minerals on the trend of the sigma for monazites is not known.

The concept of fractionation has been approached above from the assumption of magmatic crystallization. Yet in table 3, the sigmas of monazites from the most differentiated rocks--the pegmatites--are the smallest and the sigma of a granitic rock--the Whiteside Granite--is the largest. If the Whiteside Granite is not a magmatic rock but is of metamorphic origin, and if in metamorphic differentiation the most basic rare earths tend to migrate into the mobilized part of the metamorphic rock, then monazite that formed in mobilized rock would be enriched in lathanum, cerium, and praseodymium, the elements in greatest relative abundance in the Whiteside monazites.

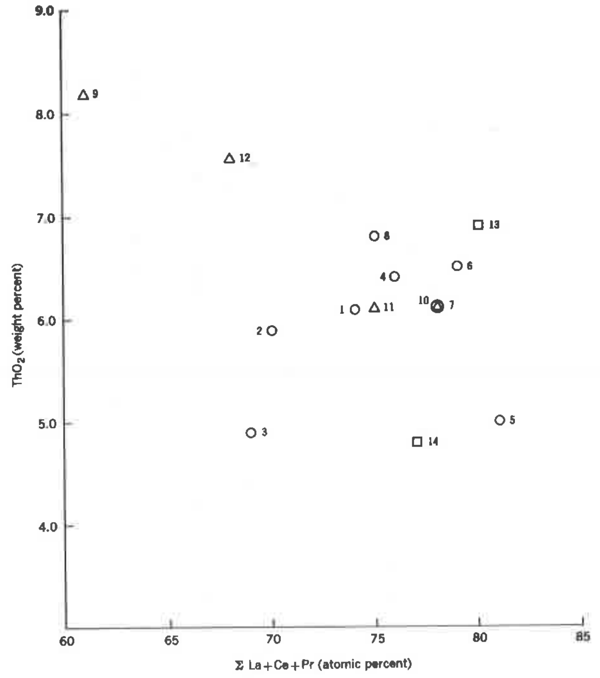

Figure 5. Relation between sigma and ThO_2. Locations of sources of monazite shown in Table 3 and/or Fig. 4. Granitic rocks (circles): 1, Stewarts Creek, N. C.; 2, Big Creek, N. C.; 3, Big Creek, N. C.; 4, Cherryville Quartz Monzonite, N. C.; 5, Whiteside Granite, N. C.; 6, Toluca Quartz Monzonite, N. C.; 7, Toluca Quartz Monzonite, N. C.; 8, Chesterfield County, Va. Pegmatite (triangles): 9, Yancey County, N. C.; 10, Pegmatite of Toluca Quartz Monzonite, N. C.; 11, Amelia County, Va.; 12, Quartz vein, Cleveland County, N. C. Metamorphic rocks (squares): 13, biotite gneiss, N. C.; 14, Biotite schist and sillimanite schist, N. C.

Relations Between Sigma, ThO_2, and U_3O_8

The granitic and pegmatitic monazites listed in Table 3 show a general decrease in ThO_2 with increasing sigma (Figure 5), a feature related by Lee and Bastron (1967, fig. 10, p. 354-355) to increasing content of CaO in the rock in which the monazite formed. This trend is shown more clearly by Figure 5 than by the direct plot of ThO_2 in monazite against CaO in the host rock (Figure 4). In Figure 5 both sets of analyses are on the same grains, whereas the data for Figure 4 are from two materials, several pairs of which consist of monazite from one locality and source rock from another.

Monazites from metamorphic rocks (Figure 5) do not follow the general relation of decreasing ThO_2 with increasing sigma, nor do the sigmas of the monazites from biotite schist and sillimanite schist reflect the considerable differences in percentages of CaO in the two schists (Figure 4). Thus, CaO in the metamorphic host rocks does not appear to be a factor in determining the percentages of thorium and the rare earths in monazites of metamorphic origin.

The three uranium-rich monazites shown in Table 3 tend to have lower values for sigma than do the monazites with ordinary abundances of U_3O_8 (Figure 6). The most uranium-rich monazite in the group is from the granitic rock leanest in CaO, the Cherryville Quartz Monzonite (Figure 4). In the ordinary monazites U_3O_8 also tends to decrease with increasing sigma. Among the metamorphic monazites, the pair with equal sigmas is displaced in percentage of U_3O_8 in a sense that reflects the lesser CaO in the sillimanite schist. However, sigma does not show the difference in CaO, just as it failed to show the difference in Figure 5. The relation of sigma to CaO (Figure 7) seems to be best defined in the granites.

From the regional data presently available no firm petrogenic interpretation seems possible from the sigma of monazites. Seemingly three important factors need to be known with greater certainty than they are at present before sigma can be employed with confidence in interpretation: (1) the form of solution of the rare-earth elements in magmatic and metamorphic fluids, which is a critical factor in the direction of fractionation of these elements, (2) the role of calcium-bearing silicate minerals and apatite in the fractionation of the rare-earth elements in magmatic and metamorphic cycles, and (3) whole-rock CaO at the sources of analyzed monazites.

POSSIBLE PETROGENIC SIGNIFICANCE OF THE MONAZITE

Stewarts Creek

The detrital monazite from Stewarts Creek is probably from the granodiorite around Mount Airy, but its composition as reflected by the ratio Ce/ (Nd+Y) only faintly reflects the remarkable manner in which monazite occurs in that rock. The monazite was reported by Dietrich (1961, p. 10) to cluster with magnetite, zircon, apatite, sphene, epidote, and biotite, and to be "surrounded by thin corona-like coatings of epidote... (and) monazite grains within relatively large grains of epidote are typically smaller than those elsewhere."

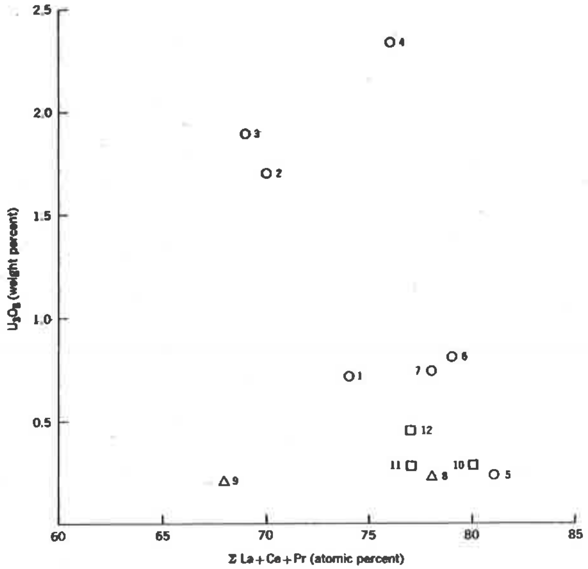

Figure 6. Relation between sigma and U_3O_8. Locations of sources of monazite shown in Table 3 and/or Figure 4. Granitic rocks (circles): 1, Stewarts Creek, N. C.; 2, Big Creek, N. C.; 3, Big Creek, N. C.; 4, Cherryville Quartz Monzonite, N. C.; 5, Whiteside Granite, N. C.; 6, Toluca Quartz Monzonite, N. C.; 7, Toluca Quartz Monzonite, N. C. Pegmatite (triangles): 8, Pegmatite of Toluca Quartz Monzonite, N. C.; 9, Amelia County, Va. Metamorphic rocks (squares): 10, Biotite gneiss, N. C.; 11, Biotite schist, N. C.; 12, Sillimanite schist, N. C.

Unrimmed monazite tends to occur between plates of biotite, but even some of these grains of monazite have epidote between them and the biotite. The epidote rimming the monazite was regarded by Dietrich (1961, p. 42) to be pyrogenic.

Dietrich (1961, p. 6-7, 56-60) regarded the rock to be leucogranodiorite of postkinematic mesozonal emplacement, possibly once to have been under great pressure. The presence of monazite possibly attested to the crystallization of the magma at great depth, but Dietrich regarded this as merely permissively corroborative data--even uncertain in the absence of analyses of the monazite--a position to be commended then and not much to be changed now.

Although Dietrich's thin sections disclosed unusual amounts of monazite, actually the rock is rather lean in monazite, as is shown by the distribution of monazite-free concentrates (Figure 2). The central part of the intrusive mass near Mount Airy contains monazite, but the peripheral parts, which are somewhat finer grained than the core,

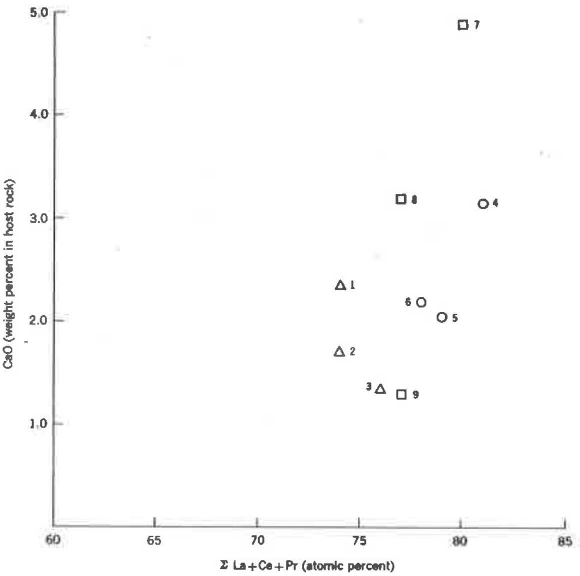

Figure 7. Relation between whole-rock contents of CaO and sigma of monazite. Postkinematic granitic rocks (triangles; localities shown in Table 3 and/or Figure 4): 1, Granodiorite, Mount Airy, N. C. CaO from Dietrich (1961, p. 28); 2, Granodiorite, Mount Airy, N. C. CaO by U. S. G. S.; 3, Cherryville Quartz Monzonite, N. C. Synkinematic granitic rocks (circles): 4, Whiteside Granite, N. C.; 5, Toluca Quartz Monzonite, Rutherford County, N. C.; 6, Toluca Quartz Monzonite, type locality, Cleveland County, N. C. Metasedimentary rocks (squares): 7, Biotite gneiss, N. C.; 8, Biotite schist, N. C.; 9, Sillimanite schist, N. C.

contain little or no monazite (Mertie, 1953, p. 24-25). From this sparsity of monazite we infer that the rock did not consolidate under unusually high pressure and temperature or at great depth; thus, Dietrich's estimate of mesozonal emplacement seems appropriate.

Only a small part of the thorium, uranium, and rare-earth elements in monazite-bearing igneous rocks are in the monazite. Early-formed minerals like apatite, and various calcium silicate minerals, including even the plagioclase feldspars are the preferred hosts of these elements (Goldschmidt, 1954, p. 317, 427, 562; Vlasov, 1966a, p. 216).

Associations of monazite with epidote, sphene, and apatite have seldom been described from thin sections of igneous rocks in the eastern United States. A notable exception is Daly's (1903, p. 56) description of monazite mantled by primary allanite in hornblende-biotite nordmarkite. Megascopic examples of monazite with rims of epidote, allanite, apatite, and sphene have been reported in a number of descriptions of pegmatite. The possible genetic implications of the monazite with rims of epidote or allanite, as seen in thin sections of igneous rocks, have not been considered beyond setting a position in the paragenetic sequence. Megascopic examples of monazite rimmed with allanite in pegmatites have been interpreted as evidence of a process of alteration of monazite to allanite related to reactions associated with declining pressure and temperature.

Epidote and allanite form an isomorphous and isostructural series in which the rare earths, thorium, and uranium are enriched in allanite over epidote (Deer and others, 1963, p. 212-216; Vlasov, 1966b, p. 302-305). The conditions under which these minerals form have been studied with conflicting results. Epidote has been said to form under oxidizing conditions and allanite under reducing conditions (Khvostova, 1961, p. 1307). In metamorphic rocks, allanite was thought to be a higher temperature mineral than rare-earth-bearing epidote (Vas'kovskii, 1965). Neither generalization satisfactorily explains the common occurrence of parallel intergrowths of epidote and allanite with epidote forming the rim (Deer and others, 1962, p. 217). From these data it cannot be said why epidote and not allanite rims the monazite at Mount Airy.

Allanite and monazite are associated only in a restricted range, and allanite is present in many rocks that lack monazite. Allanite is a common accessory mineral in extrusive silicic rocks whereas monazite is unreported. Allanite is common in shallow granitic rocks where monazite is sparse, but allanite is an extremely rare mineral in plutonic silicic rocks where monazite is a common accessory mineral. Both allanite and monazite are found as accessory minerals in granitic pegmatites, and allanite commonly replaces the monazite; locally monazite replaces allanite in pegmatite. These relations have been interpreted in several ways. Allanite is thought to be generally more stable than monazite in silicic magmatic rocks which consolidate at low temperature or pressure, and monazite is thought to be more stable than allanite at high temperature and pressure (Overstreet, 1967, p. 23). In the intrusive granitic rocks at Mt. Wheeler, Nev., allanite was found by Lee and Bastron (1967, p. 340-341) to be present to the exclusion of monazite in rocks containing more than about 2.0 weight percent CaO, and monazite was present to the exclusion of allanite in rocks with less than about 0.7 percent CaO. At intermediate percentages of CaO, both allanite and monazite are present. These percentage relations of CaO to the relative distribution of allanite and monazite may be of only local application. For example, rholites, shown by Daly (1933, p. 9) to average 1.2 percent CaO, are not known to contain monazite although they commonly contain allanite. Analyses, quoted by Dietrich (1961) of the monazite- and epidote-bearing granodiorite at Mount Airy show as much as 2.96 percent CaO.

The coronalike mantles of epidote on monazite in the granodiorite at Mount Airy may show arrested crystallization of the monazite and replacement of the monazite by epidote owing to incomplete reaction between monazite and magma under conditions of accelerated crystallization related to reduced temperature and pressure. The mechanism for this reaction might be the early crystallization of accessory monazite under conditions of temperature and pressure too great for the formation of epidote. Reduction of pressure and temperature into a range less favorable for monazite and more favorable for epidote would arrest the crystallization of monazite and promote that of epidote, allowing epidote rims to form over monazite. Further growth of the coated monazite would be prevented, and actual replacement of monazite by epidote might take place. Reaction between already formed monazite and the magma could lead to the elimination of uncoated monazite as a mineral phase when the magma passed into a pressure-temperature range unfavorable for the precipitation of monazite. The rare earths, thorium, uranium, and phosphate of the replaced monazite would enter newly forming calcium silicates and phosphate, the earlier formed of which already harbored the bulk of these elements. Had the magma remained in this PT environment possibly most of the originally formed monazite would have disappeared and the crystalline rock would be virtually monazite free but epidote bearing. Rapid lowering of PT conditions might preserve some of the monazite.

Presently available data on the distribution of monazite in the rock at Mount Airy seem to contradict the above supposition. If the magma rose rapidly from depth and crystallization of monazite was arrested with concomitant acceleration of crystallization of epidote, the finer grained marginal parts of the pluton that cooled rapidly should be richest in monazite. The central, coarser grained part, where cooling was probably slower than at the walls, seemingly should have provided a better opportunity for replacement of monazite by epidote. Thus, the walls should be richer in monazite. Obviously, a detailed study of the distribution of monazite in the pluton is needed to see to what extent the distribution of monazite is related to the cooling history of the body.

The percentage of thorium in the Stewarts Creek monazite is appropriate for monazite that has crystallized under higher PT conditions than that reached in regional metamorphism by the wallrocks of the pluton at Mount Airy. Such monazite is inferred to have been moved upward to its present site, relative to the wallrocks, through magmatic intrusion.

The uranium content of the Stewarts Creek monazite is not petrogenically distinctive; however, that of the Big Creek monazites is.

The ratio of Ce/ (Nd+Y) of 2.12, giving a sigma of 74, lies between a possibly strongly fractionated and poorly fractionated distribution of the rare earths. A much more definite indication of lack of fractionation might be expected than is shown here if crystallization of the monazite had been arrested early. However, too many uncertainties attach to sigma to permit its use with confidence in this petrogenic interpretation.

Big Creek

The Big Creek monazites are probably relatively unfractionated as suggested by sigmas of 69 and 70 and their unusual uranium contents (Figure 3). Early arrest of the precipitation of monazite through rapid decrease in PT conditions might account for uranium-rich monazites with low sigmas like those from Big Creek.

The similarity of the composition of the Big Creek monazites to the composition of monazites from the Cherryville Quartz Monzonite in Cleveland County, N. C., is remarkable. Further evidence for correlation between the posttectonic Cherryville and the postmetamorphic plutons at Big Creek and Mount Airy needs to be sought. One sample of monazite from the Cherryville Quartz Monzonite gave a lead-alpha age of 260 m. y. (Overstreet and Bell, 1965, table 7), which is within the group of isotopic ages reported for the granodiorite at Mount Airy (Butler and Dunn, 1968, table 3).

Speculative Summary

Considerations of depth of emplacement of the granodiorite around Mount Airy and the interpretation of pyrogenic origin for the epidote and muscovite led Dietrich (1961, p. 58) to postulate that the magma was once under great pressure but that emplacement may have involved appreciable magmatic movement. Arrested crystallization of monazite owing to rapid lowering of PT conditions in the magma chamber was offered above as a possible explanation for the association of monazite and pyrogenic epidote in the granodiorite and for the low sigma and high content of uranium in the Big Creek monazites. Magmatic movement was postulated as the probable process for bringing the thorium-rich Stewarts Creek monazite into the level of the staurolite-kyanite subfacies wallrocks around the pluton at Mount Airy. Rapid lowering of PT conditions might have occurred concurrently with appreciable movement of the magma.

For example, relatively unfractionated monazite has been reported from the incipient cryptovolcanic structure at Hicks Dome, Ill. (Trace, 1960). The facts that the Hicks Dome monazite is relatively unfractionated, that it has a large percentage of ThO_2 for monazite in calcareous rocks, and that it occurs in a cryptovolcanic structure were interpreted to mean that the Hicks Dome monazite formed with little fractionation at depth and was transported with volcanic rapidity toward the surface (Overstreet, 1967, p. 162). Rapid upward transport of the monazite probably prevented reactions that would have led to lower abundances of thorium and yttrium in the monazite or to a possible elimination of the monazite phase.

Are the association of monazite and epidote in the granodiorite at Mount Airy and the exotic compositions of the monazites from Big Creek evidence of magmatic movement of volcanic rapidity in the post-tectonic plutons in Surry and Stokes Counties, N. C.? Can these plutons be the roots of Paleozoic felsic volcanic centers? Stratigraphic evidence to answer these questions, as suggested by G. H. Espenshade, USGS (oral commun., 1968), might be sought in the rocks immediately under the sedimentary rocks of the Dan River Triassic Basin, 35 miles to the east (not shown on Figure 2), and in cobbles in Triassic conglomerates. Are little-metamorphosed or unmetamorphosed felsic volcanic rocks, of the same age as the pluton at Mount Airy, present under the Triassic sediments? Do the Triassic conglomerates and fanglomerates contain pebbles derived from hypabyssal rocks, such as granite porphyry, or from felsic lava representing volcanic sequences now removed by erosion? Relicts of such sequences, if in fact they existed, might be preserved as fragments in the conglomerates or as erosional remnants under the lowest Triassic sediments. Isotopic ages of minerals in rock fragments in the conglomerate would aid in possible correlation with the pluton at Mount Airy. Thin sections of cobbles of granite porphyry might show the distinctive mantles of epidote on monazite seen in the granodiorite at Mount Airy.

Structural evidence, particularly the shape and probably thicknesses of the plutons, would be useful in learning whether the plutons could be root zones of felsic volcanic centers. The gravity data developed by V. I. Mann and presented by Dietrich (1961, p. 6) suggest that the pluton at Mount Airy is one of the few bodies of granitic rock in North Carolina to extend in depth. Hooker and Johnson (1967, p. 5, 15; 1969, p. 10, 14, fig. 3) have determined large near-surface horizontal stresses with apparent orientation of the principal stress at N. 87° E., in the granite at the quarry of the North Carolina Granite Corporation at Mount Airy. The magnitude of the principal compressive stress measured in the quarry at Mount Airy is the greatest recorded for the Appalachian Piedmont. Although the maximum compressive stress at each of eight sites seems to be aligned with major structures, the geologic interpretation of this feature is uncertain, particularly when one of the measured sites is in diabase of Triassic age in Virginia (Hooker and Johnson, 1969, fig. 3).

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