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SOUTHEASTERN GEOLOGY

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LIGNITE AND THE MIDWAY-WILCOX STRATIGRAPHIC BOUNDARY

MISSISSIPPI AND ALABAMA

C. R. MEISSNER, JR. *U. S. Geological Survey, Reston, Virginia 22092*

S. E. HEERMANN *U. S. Geological Survey, Reston, Virginia 22092*

ABSTRACT

The stratigraphic boundary between the Midway Group of Paleocene age and the overlying Wilcox Group of Paleocene-Eocene age needs clarification in the border area along east-central Mississippi and west-central Alabama. A lignite bed of the Oak Hill Member of the Naheola Formation, upper Midway Group, in Alabama appears to trace directly into a lignite bed in the Fearn Springs Member of the Nanafalia Formation, lower Wilcox Group, in Mississippi. If the lignite bed is the same between the two states then this would imply that it crosses the Midway-Wilcox boundary, which is an impossible situation. A core drilling and sampling program is recommended in the Mississippi-Alabama border area to help define the true relationship of the Oak Hill-Fearn Springs lignite beds and the positioning of the Midway-Wilcox stratigraphic boundary.

INTRODUCTION

The boundary between the Midway Group of Paleocene age and the overlying Wilcox Group of Paleocene-Lower Eocene age needs to be defined in the border area along east-central Mississippi and west-central Alabama (Fig. 1). A lignite bed described in the Oak Hill Member of the Naheola Formation, upper Midway Group, in Alabama (Daniels, 1973) appears to be the same bed as that described in the Fearn Springs Member of the Nanafalia Formation, lower Wilcox Group, just across the border in Mississippi (Foster, 1940) (Fig. 2). If the lignite bed actually is continuous from Alabama into Mississippi, then it is impossible for it to cross a major stratigraphic boundary such as that between the Midway and Wilcox Groups, especially as this boundary has been described as a major unconformity (Hughes, 1958). The U. S. Geological Survey Lexicon of Geologic Names of the United States from 1936-1960 (Keroher, 1966) cites eight authors, none of whom completely agree on the definition of the Fearn Springs Member and its stratigraphic position. MacNeil (1946) gives the option of three different positions for the Midway-Wilcox boundary, one at the base of the Naheola Formation (upper part of Midway), one at the base of the Coal Bluff Marl Member of the Naheola, and one at the base of the Fearn Springs Member of the

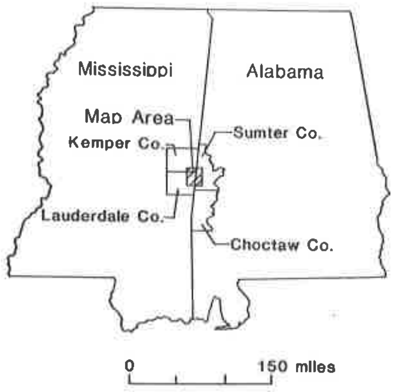

Figure 1. Index map of Mississippi and Alabama showing the area where the lignite bed was mapped.

Series	Group	East-Central Mississippi	Western Alabama
Paleocene	lower Wilcox	Nanafalia Fm. Fearn Springs Mbr.	Nanafalia Fm. Grampian Hills Mbr. "Ostrea Thirsae" beds Gravel Creek Mbr.
	upper Midway	Naheola Fm. Porters Creek Clay	Naheola Fm. Coal Bluff Marl Mbr. Oak Hill Mbr. Porters Creek Clay

Figure 2. Chart showing stratigraphic nomenclature of formations of the lower Wilcox and upper Midway Groups presently used in east-central Mississippi and western Alabama.

Nanafalia Formation (lower part of Wilcox). In later work, MacNeil (1951) traced the Fearn Springs into Alabama where it merged with the upper part of beds assigned to the Naheola (Coal Bluff Member). He then considered the upper beds equivalent to the Fearn Springs and assigned them to the basal Nanafalia (lower part of Wilcox), and the lower beds (now called the Coal Bluff Marl Member which overlies the Oak Hill Member) were assigned to the Naheola (upper part of Midway). Recent paleontological work with micro and macro fossils in eastern Alabama by Norman O. Frederiksen, Thomas G. Gibson, and others, of the U. S. Geological Survey, suggested that the Fearn Springs belongs in the upper part of the Midway Group (oral communication, 1982), but this has not been confirmed in western Alabama or Mississippi.

None of the above cited information explains the apparent continuity of the Oak Hill lignite of Alabama into the Fearn Springs lignite of Mississippi and for this reason the authors of the present study conducted surface investigations in the region in an attempt to confirm the apparent continuity of the lignite bed and to establish its relationship with the Midway-Wilcox boundary. These investigations are summarized below.

SURFACE RECONNAISSANCE

During the period January 12-23, 1982, the lignite bed described in the Fearn Springs Member of the Nanafalia Formation (lower part of Wilcox) was traced from southern Kemper County, Mississippi, across the northeastern corner of Lauderdale County, Mississippi, into Sumter County, Alabama, a distance of about 20 miles (Fig. 3). The topography is gently rolling on poorly consolidated sandstone, siltstone, and claystone and although exposures are few and far between, about 20 outcrops were visited. These include measured sections or isolated outcrops cited in the literature and those found on the recent reconnaissance. All of these exposures do not contain lignite, but at some places lignitic sediments and/or underclays are found at the same stratigraphic horizon indicating continuity of the horizon of the lignite bed. Elevations of the lignite outcrops were determined by altimeter. These surface elevations were supplemented by available drill hole data obtained from the Kemper County, Mississippi, report (Hughes, 1958), the Lauderdale County, Mississippi, report (Foster, 1940), and lignite investigations in Sumter County, Alabama (Daniels, 1973). In all, elevation data were obtained for 40 points (Fig. 3).

STRUCTURE-CONTOUR MAP

Structure-contour lines at 20-foot intervals were drawn on top of the lignite bed (Fig. 4). These contours show a regional southwest dip averaging about 26 feet per mile, or about 1/4°. This dip is similar to that reported in the literature by earlier workers. A small structural dome with over 40 feet of closure was contoured in Sumter

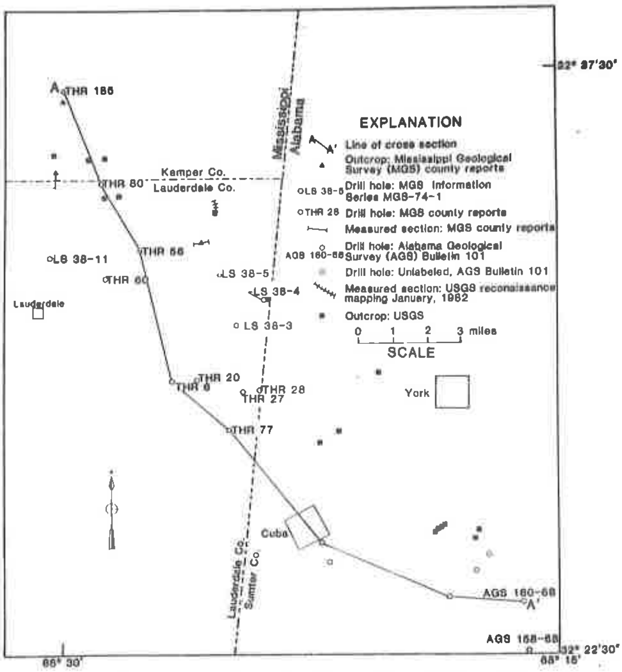

Figure 3. Location map showing outcrop and drill hole data points.

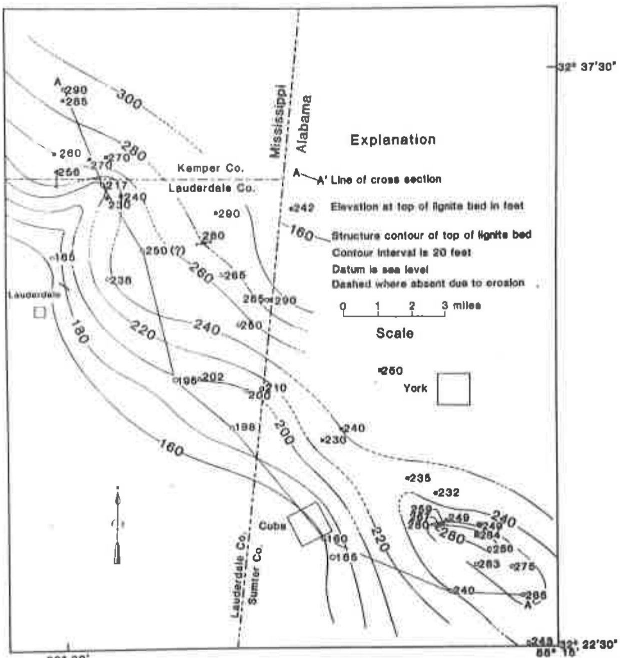

Figure 4. Structure contour map of the Oak Hill-Fearn Springs lignite bed.

County south of York, Alabama. This dome is as much as 6 miles long and 2 1/2 miles wide. Surface evidence indicates a northeast reversal of dip as much as 4^{o} along a cut on the St. Louis-San Francisco railroad about 4 miles south of York, Alabama. Mapping along a creek a mile or so west of this cut also indicates northeast dip. The lignite bed at this railroad cut is part of the Oak Hill Member of the Naheola Formation (upper part of Midway-Daniels, 1973).

ISOPACH MAP

An isopach map of the Oak Hill-Fearn Springs lignite bed was drawn from the thicknesses recorded in the same outcrop and drill hole data used in constructing the structure-contour map (Fig. 5). The isopach map indicates a nearly continuous bed from Sumter County, Alabama, to Lauderdale County, Mississippi. Apparently, small parts of the bed in northeastern Lauderdale County may thin to zero or grade into lignitic and carbonaceous clay and silt. The thickest part of the bed is in Sumter County, Alabama, south of York where it is as much as 9 feet thick in elongate pods. The bed trends northwestward into Lauderdale County, Mississippi, where its thickness is as much as 5.5 feet. The northeastern boundary of the bed is where it pinches out, or changes into lignitic and carbonaceous clay and silt, or where it crops out at the surface at its up-dip end. The down-dip boundary of the bed as contoured on the isopach map is simply limited to the available shallow well data. Actually, it is reasonable to assume the bed extends well down-dip into the subsurface to the southwest.

GEOLOGIC CROSS SECTION

A geologic cross section was constructed using published drill hole data of eight shallow holes extending along a line from Sumter County, Alabama into Kemper County, Mississippi (Fig. 6). One of these holes is about 95 feet deep, whereas the rest are not more than 45 feet deep. A single bed of lignite was penetrated in each hole which is interpreted by the authors to be the same bed. This means, from southeast to northwest, that the bed continues from the Oak Hill Member of the Naheola Formation (upper Midway Group) into what has been included in the Fearn Springs Member of the Nanafalia Formation (lower part of Wilcox). In the construction of the cross section at a manageable size, the difference between the vertical and horizontal scale resulted in a vertical exaggeration of approximately 200 times normal. This

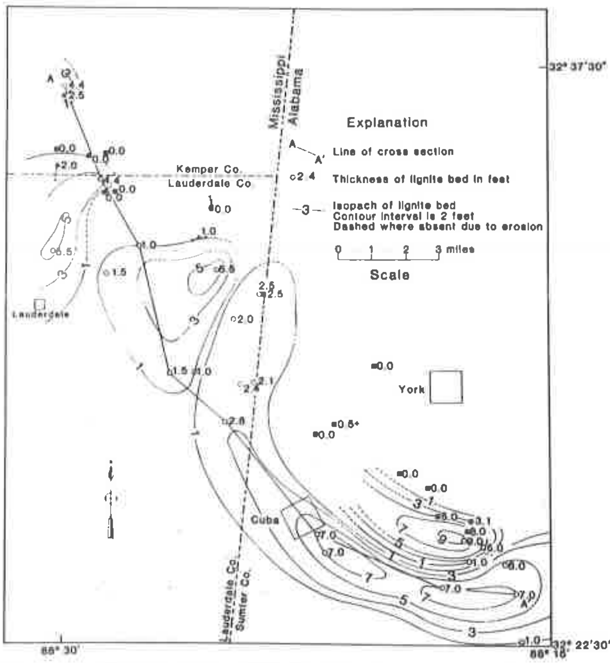

Figure 5. Isopach map of the Oak Hill-Fearn Springs lignite bed.

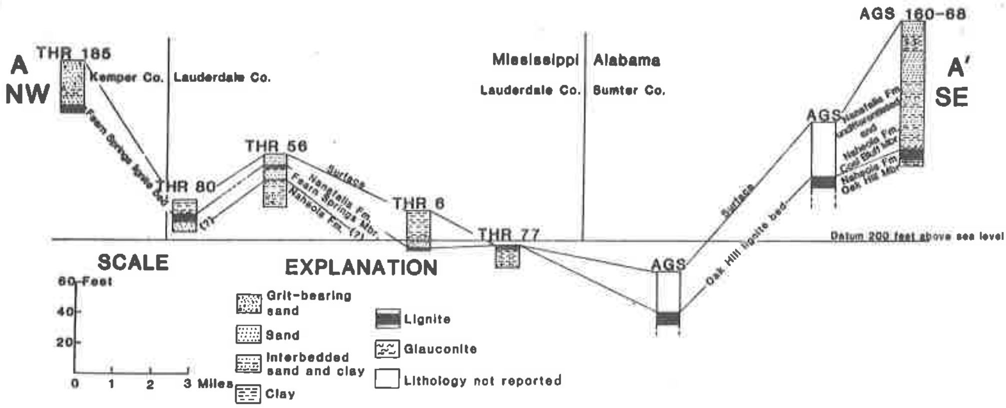

Figure 6. Geologic cross section of the Oak Hill-Fearn Springs lignite beds, Alabama-Mississippi.

exaggerated the dip between drill holes, which if drawn to the same vertical and horizontal scale, would show a very gentle dip between sections. None of the drill holes are deep enough to adequately reveal the stratigraphic relationship of the overlying sedimentary formations with the underlying formations. One hole, THR 56, does show the contact of Fearn Springs with Naheola. This contact, if accurate, shows the position of the lignite bed near the base of the Fearn Springs. None of the other holes show this contact, which also is supposedly the contact between the Wilcox Group and underlying Midway Group, so that the continuity and exact position of the Midway-Wilcox Group contact cannot be ascertained. This is true in the area of other drill holes which are too shallow or widely spaced to satisfactorily trace the Midway-Wilcox contact or to determine the relationship of this contact with the Oak Hill and the Fearn Springs Members. The continuity of the lignite bed indicates that the two members are equivalent, because the bed seemingly extends from one member into the other. However, the determination of the Midway-Wilcox boundary requires stratigraphic drilling of sufficient depth in the border area between Alabama and Mississippi to examine lower Wilcox-upper Midway sediments as a basis for defining and correlating formations in this area. Such a drilling program would be done best by a cooperative project involving the states of Alabama and Mississippi, and the U. S. Geological Survey.

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UPPER CRETACEOUS (CAMPANIAN-MAESTRICHTIAN)

MARINE STRATA IN THE SUBSURFACE OF NORTHERN DELAWARE

C. W. HOULIK, JR. *Ecological Analysts, Inc., Hunt Valley/Loveton Center, 15 Loveton Circle, Sparks, Maryland 21152*

RICHARD K. OLSSON *Department of Geological Sciences, Wright Laboratories, Rutgers University, New Brunswick, New Jersey 08903*

RICHARD W. AURISANO[1] *Department of Geological Sciences, Wright Laboratories, Rutgers University, New Brunswick, New Jersey 08903*

ABSTRACT

A sedimentologic, biostratigraphic, paleoecologic, and geophysical log study of the Merchantville, Englishtown, Marshalltown, and Mount Laurel formations in the subsurface of northern Delaware illustrates the utility of the integration of these data for subsurface studies. Age based on biostratigraphy of planktonic foraminifera and dinoflagellates, environment of deposition determined by foraminiferal analysis, and sedimentological analysis correlated with geophysical logs provides the framework for subsurface studies of the geologic history of the study area.

The Campanian Merchantville, Englishtown, and Marshalltown formations and the Maestrichtian Mount Laurel Formation were deposited during alternating transgressive and regressive changes in sea level. Deposition of these formations took place in mid to outer shelf, inner shelf, outer shelf, and mid shelf depths, respectively.

The lithologic character of these formations when identified on geophysical logs and integrated with the paleontologic data gives a dynamic geologic significance to the subsurface geophysical log correlations.

INTRODUCTION

A series of test borings and observation wells were completed in an area extending roughly two miles south of the Chesapeake and Delaware Canal in New Castle County, Delaware (Fig. 1). Surficial exposures in this vicinity (primarily along the Chesapeake and Delaware Canal) have received considerable attention for many years (e.g., Carter, 1937; Groot and others, 1954; Groot, 1955; Mumby, 1961; Jordan, 1962; Richards and Shapiro, 1963; Olsson, 1964; Pickett, 1970; Owens and others, 1970; and Owens and Sohl, 1973). A synopsis of these studies is contained in Benson, 1976; Pickett, 1976; and Owens and others, 1970. During the study described in this paper, samples (1.5 or 2.5 inch split spoon) were obtained from the borings at five-foot intervals or closer. Spontaneous potential, resistivity and natural gamma logs were also obtained. Samples from several of the borings were examined for microfaunal content.

A sedimentologic zonation of the Upper Cretaceous section penetrated was established on the basis of visual examination of the borehole samples, and it was found that this zonation was recognizable on the borehole geophysical logs. Biostratigraphic zonation on the basis of planktonic foraminifera was found to agree well with the sedimentologic contacts. Subsequently, an independent evaluation of the biostratigraphy on the basis of dinoflagellates further substantiated the sedimentologic and foraminiferal interpretation of the Campanian-Maestrichtian stratigraphy.

These analyses, summarized in Figure 2, demonstrate that the Merchantville, Englishtown, and Marshalltown formations of the Matawan Group and at least the lower part of the Mount Laurel Formation of the Monmouth Group are recognizable in the subsurface of northern Delaware.

[1]Present address: Research Center, Amoco Production Company, Tulsa, Oklahoma.

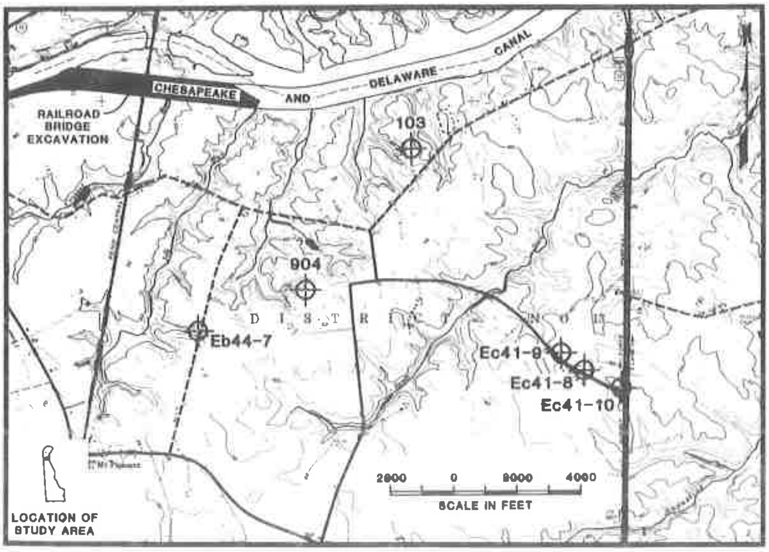

Figure 1. Location map.

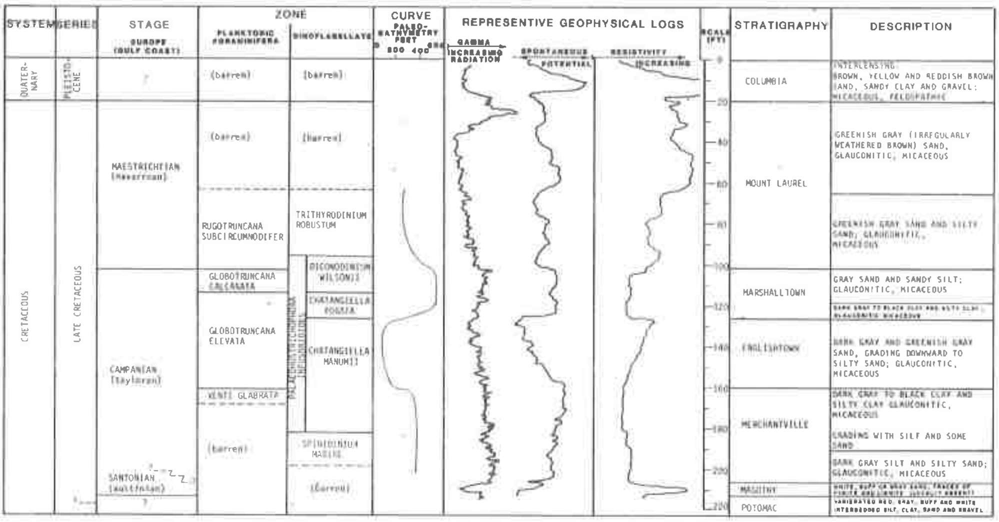

Figure 2. Lithology and stratigraphy of Upper Cretaceous marine strata in northern Delaware.

SEDIMENTOLOGY - STRATIGRAPHY

Figure 3 is a fence diagram illustrating the subsurface distribution of the Campanian and Maestrichtian strata in the study area. Correlation with the surface exposure described by Owens and others (1970) is also illustrated.

Strata identified as the Magothy Formation in this study consist of white, buff or gray, fine-grained, subangular quartz sand. Most samples contain a trace of silt size cubes of pyrite and fine particles of lignite. This unit is discontinuous within the study area and, where present, ranges in thickness from 3 to 16 feet. It is possible that the lagoonal facies (gray or black clayey silt) of the Magothy Formation exists within the study area. However, silt samples obtained from these borings from the interval directly overlying the red or light gray clay or silt of the Potomac Group are glauconitic and were assigned to the Merchantville Formation on that basis.

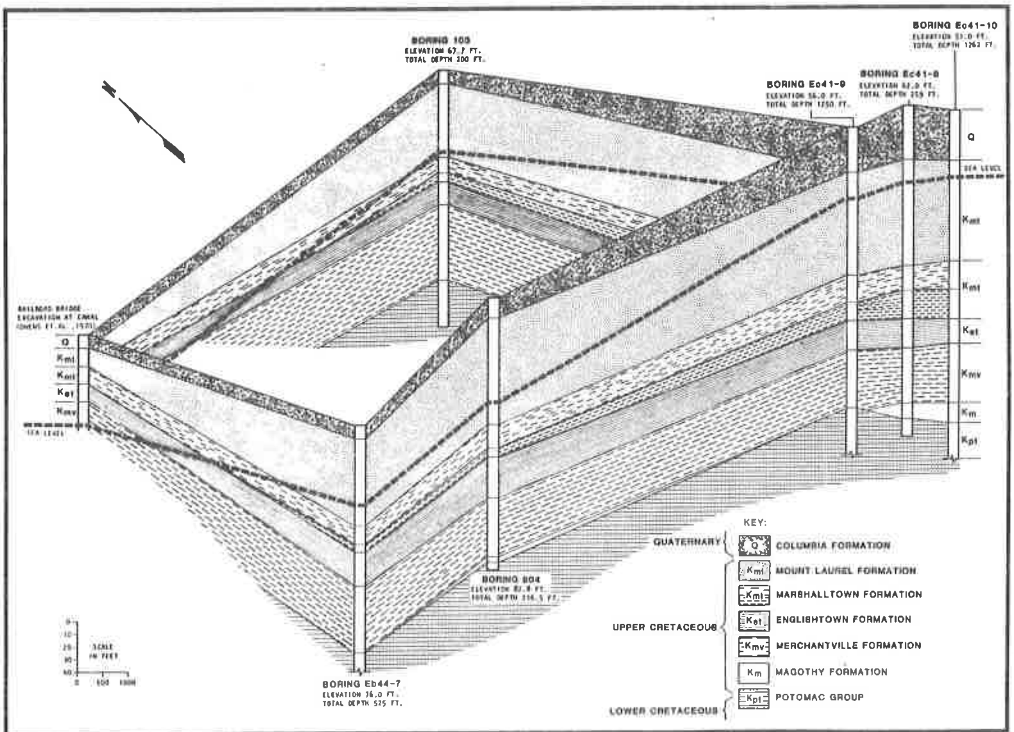

Figure 3. Fence diagram of Upper Cretaceous marine strata in northern Delaware.

Both the upper and lower beds of the Merchantville Formation are exposed in the banks of the Chesapeake and Delaware Canal. However, the section is not continuous and the central portion of the formation is concealed. The total thickness of the Merchantville Formation at the railroad bridge is about 50 feet (Owens and others, 1970). The upper 23 feet of the formation is composed of poorly to well sorted, very fine to fine, very silty, and clayey sand. This sand is pale gray in the upper 8 feet, the remainder being dark grayish black. Quartz, glauconite, feldspar, chlorite, and muscovite are the major minerals. Pyrite and siderite are important accessory minerals, particularly in the upper beds. Fossils are abundant but mostly consist only of shell imprints. All calcareous material has been leached. A thin zone containing abundant crab claws is found near the top of the formation. Where they are exposed, the lower Merchantville beds apear to be more clayey and less micaceous and to have a higher glauconite content than the upper beds.

In the subsurface, the Merchantville Formation is generally on the order of 45 feet thick within the study area. The basal 15 to 20 feet of the formation consist of dark gray and greenish sandy silt. This is overlain by approximately 25 feet of dark gray silty clay. Highly micaceous (muscovite) and glauconitic zones are found throughout the formation but appear to be more consistently present in the lower 20 feet. Some depositional laminae are present, but most of the sediment has been homogenized by burrowing organisms. Glauconite tends to be concentrated in the burrows; large discrete burrows are filled with glauconite, sand, and silt. Pelecypod fragments are commonly found in the lower part of the formation.

The top of the Merchantville Formation is readily identifiable on electric logs due primarily to the characteristic shape of the overlying Englishtown Formation (Figure 2). The contact occurs at the break in slope between the relatively straight, near vertical log of the Merchantville clay and the inverted triangle form of the Englishtown Formation, which results from its coarsening upward in grain size.

The Englishtown Formation is primarily a fine to very fine-grained, well-sorted, silty sand having a thickness of 14 feet (Owens and others, 1970). The sands are white to pale buff in color, the upper part being pale gray in the lower few feet. Many of the beds are less than one inch in thickness and small-scale crossbedding is present, the crossbedding being thicker in the lower few feet. In outcrop the entire formation is extensively stained orange-brown by iron oxides along the bedding planes and penetrated by excellently preserved burrows.

In the subsurface, the Englishtown Formation comprises a coarsening upward sequence of gray sandy silt, which grades upward to a green-black silty sand. The sand is fine grained, subangular to subrounded quartz with varying amounts of botryoidal pellets and angular grains of glauconite. Muscovite, chlorite, and pelecypod fragments are found throughout the unit.

In outcrop a 14 foot section of the Marshalltown Formation described by Owens and others (1970) consists of a grayish green, clayey, silty sand, which grades downward into a greenish-black massive sand. Glauconite is the major sand constituent, although quartz and feldspar are abundant in the lower few feet. Borings filled with glauconitic sand are common, as are shell fragments and molds.

In the subsurface, the Marshalltown Formation consists of a basal 9 to 20 feet of dark gray to black clay and silty clay overlain by 14 to 20 feet of greenish-gray, medium-grained quartz and glauconite sand. Muscovite and pelecypod fragments are common throughout the formation.

The basal clay of the Marshalltown Formation pinches out in the up dip direction between the borings and the canal exposure and thickens in the down dip direction (Fig. 3). It has been tentatively identified in a well (Delaware Geological Survey, Open File log no. Ec15-28) about five miles northeastward along strike at Delaware City. The driller's log of this well identifies seven feet of clay above the Englishtown Formation. The top of Englishtown pick was made on the electric log of the well.

The Marshalltown Formation is also identifiable on borehole geophysical logs (Fig. 2). Its low response on spontaneous potential and resistivity logs has been utilized by Petters (1976) in mapping the Marshalltown Formation throughout New Jersey. Woodruff (1976a and 1976b) described the characteristic peak on gamma logs at the top of the Marshalltown Formation in Delaware.

Owens and others (1970) described the basal 15 feet of the Mount Laurel Formation as a somewhat argillaceous and silty, yellowish to reddish brown sand. Quartz, feldspar, and glauconite are the major sand sized constituents; muscovite and apatite also are present in significant concentrations. The sand is moderately to well sorted and mostly fine to medium grained.

In the subsurface, the Mount Laurel Formation ranges up to 85 feet in thickness within the study area. The basal 20 feet of the formation consists of dark greenish-gray argillaceous fine sand which grades upward into 15 to 18 feet of silty fine sand. This is in turn overlain by 35 to 45 feet of fine to medium grained, occasionally silty dark green, gray, or brown (weathered) sand. The Mount Laurel sand is subangular to subrounded quartz with botryoidal pellets and angular grains of glauconite. Muscovite flakes are common. Discrete burrows are recognizable throughout the formation, pelecypod fragments are common to abundant, and a few belemnites were recovered in borehole samples.

BIOSTRATIGRAPHY

A biostratigraphic analysis of planktonic foraminifera contained in core samples from boring 904 (Figs. 3, 4) shows that the upper Merchantville Formation, the Englishtown Formation, the Marshalltown Formation, and the lower Mount Laurel Formation can be placed in the zonal scheme utilized in the New Jersey coastal plain (Olsson, 1964; Petters, 1976, 1977). Furthermore, a study of the dinoflagellate stratigraphy in this boring reveals an identical zonation to that established in New Jersey by Aurisano (1980). A Campanian to lower Maestrichtian age for the formations in this study is based upon these zonations (Fig. 2).

The uppermost part of the Merchantville Formation contains the middle Campanian *Ventilabrella glabrata* planktonic foraminiferal zone (Figs. 2, 4). The lower part of the Merchantville Formation, although barren of foraminifera, can be placed in the lower Campanian on the presence of the dinoflagellate *Spinidinium mariae* Zone (Figs. 2. 6), which, according to Aurisano (1980), is the lowermost dinoflagellate zone of the Campanian.

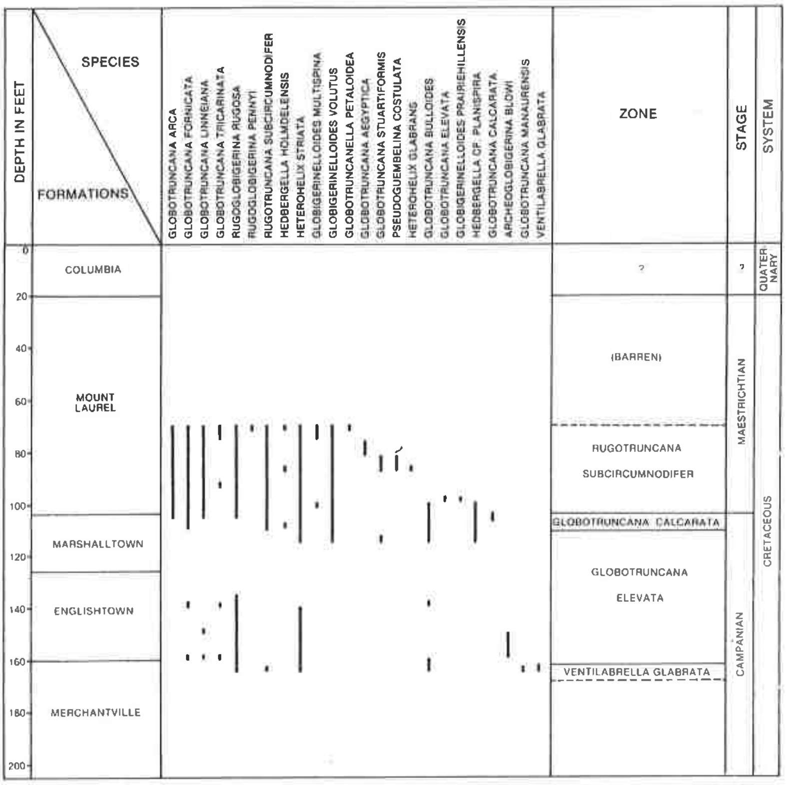

Figure 4. Distribution of planktonic foraminiferal species in boring 904.

The overlying Englishtown Formation lies within the upper Campanian *Globotruncana elevata* planktonic foraminiferal zone (Figs. 2, 4). A well-defined dinoflagellate assemblage also occurs in the Englishtown Formation (Fig. 6) and places the formation within the upper Campanian *Chatangiella manumii* Subzone of the *Palaeohystrichophora infusorioides* Zone (Aurisano, 1980).

In boring 904 foraminifera are absent in the lower clay unit of the Marshalltown Formation. A dinoflagellate assemblage is present, however, placing this unit in the upper Campanian *Chatangiella porata* Subzone of the *P. infusorioides* Zone (Figs 2, 6). Foraminifera are abundant in the upper sand unit of the Marshalltown Formation. This unit contains the uppermost Campanian *Globotruncana calcarata* planktonic foraminiferal zone, which allows the Campanian-Maestrichtian boundary to be placed between the Marshalltown Formation and the overlying Mount Laurel Formation (Figs. 2, 4). The dinoflagellate *Diconodinium wilsoni* Subzone of the *P. infusorioides* Zone also occurs in the sand unit (Fig. 6), but this zone also extends into the lower Maestrichtian and, therefore, is not useful for defining the Campanian-Maestrichtian boundary.

Only the lower portion of the Mount Laurel Formation contains microfossils; the upper part is devoid of microfossils, apparently due to weathering. The lower part of the Mount Laurel Formation contains the lower Maestrichtian *Rugotruncana subcircumnodifer* planktonic foraminiferal zone (Figs. 2, 4). In addition to the *D. wilsoni* Subzone, it also contains the lower Maestrichtian *Trithyrodinium robustum* dinoflagellate zone (Fig. 6).

The planktonic foraminiferal biostratigraphy of the upper Campanian-lower Maestrichtian was examined in two other borings (103 and Eb44-7) used in this study in order to determine whether the lithologic and zonal relations observed in the 904 boring are similar. In these borings the uppermost Campanian *Globotruncana calcarata* Zone and the lowermost *Rugotruncana subcircumnodifer* Zone lie within the Marshalltown Formation and the Mount Laurel Formation, respectively, thus demonstrating that the paleontological and physical characteristics of these units are correlatable in the subsurface of the study area.

PALEOECOLOGY

The formations penetrated in the study area are part of two depositional phases recognized by Petters (1976) in New Jersey to reflect major transgressive-regressive cycles. The foraminiferal assemblages of these formations (Fig. 5) are indicative of the changing bathymetry of these cycles.

The Merchantville Formation was deposited during a widespread marine transgression which followed a period of major regression. The Magothy Formation, which is a nearshore facies of the Merchantville Formation (Spoljaric, 1972; Petters, 1976), lies disconformably on the formation below due to the erosion caused by the major regression. The foraminiferal assemblage in the upper part of the Merchantville Formation (Fig. 5) indicates mid to outer shelf depths (300-400 feet) during deposition of this formation. Although foraminifera are absent in the lower part of the Merchantville Formation, due most probably to dissolution, the palynodebris belongs to a micrinitic palynofacies (Aurisano, 1980), which Habib (1979) interprets as material accumulated under marine conditions with minor terrigenous influence. The palynofacies thus is consistent with the foraminiferal evidence of mid to outer shelf depths during deposition.

The Merchantville Formation was deposited during a major upper Santonian to lower Campanian transgressive deposition phase (Petters, 1976). The succeeding Englishtown, Marshalltown, and Mount Laurel formations were deposited during the Campanian to Maestrichtian oscillation phase of Petters (1976), a phase that includes minor regressions interrupted by minor transgressions. Major unconformities are absent.

The coarsening upward sand deposit which comprises the Englishtown Formation is indicative of the shallowing water depths associated with the regression that ended deposition of the Merchantville Formation. A low diversity of benthonic foraminiferal

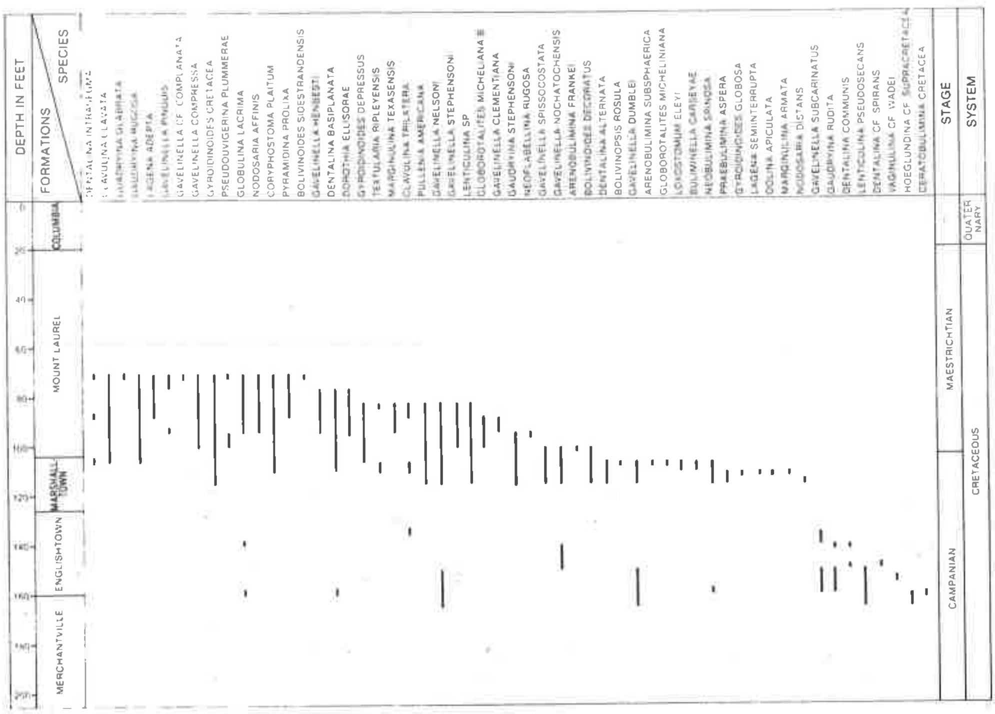

Figure 5. Distribution of benthonic foraminiferal species in boring 904.

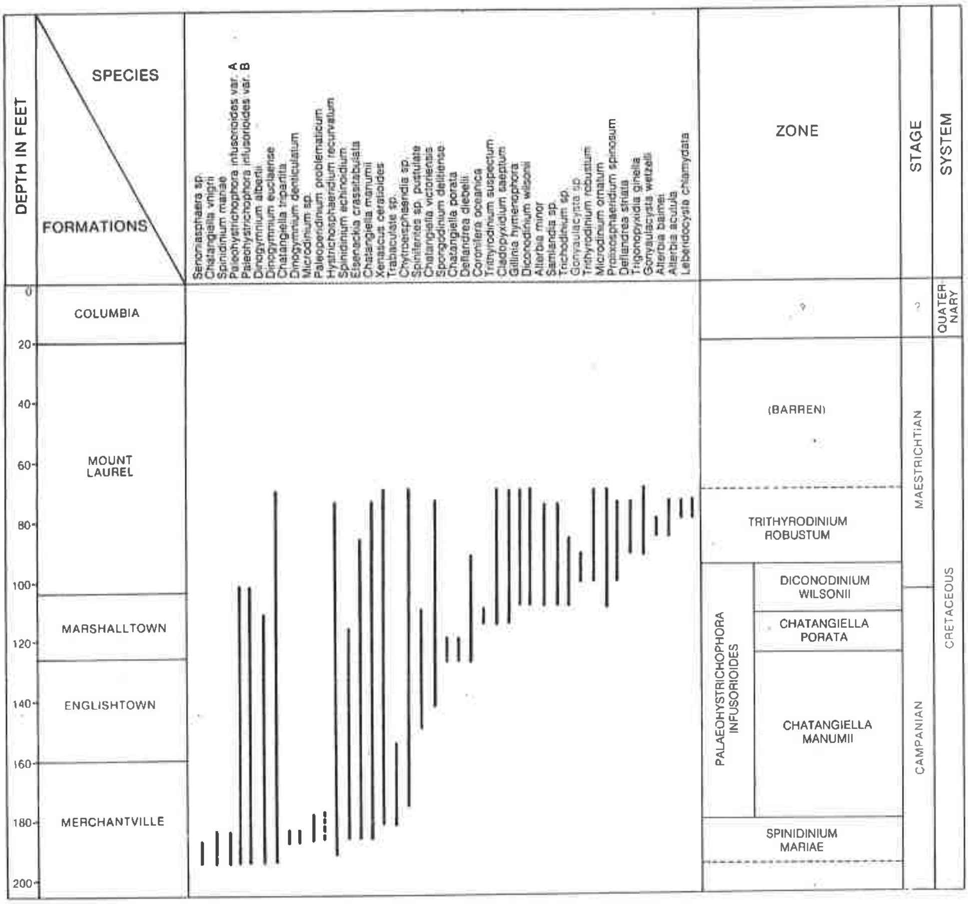

Figure 6. Distribution of dinoflagellate species in boring 904.

species (Fig. 5), which include *Gavelinella subcarinatus* (Cushman and Deaderick) and *Gaudryina rudita* Sandidge as prominent members, suggests inner shelf depths of less than 150 feet during deposition of the Englishtown Formation (Fig. 2). Very low numbers of small-sized specimens of a few species of planktonic foraminifera are also indicative of such shallow depths.

In sharp contrast to the Englishtown Formation, the foraminiferal assemblage of the Marshalltown Formation reflects a substantial increase in water depth during Marshalltown sedimentation. An abundant and very diverse assemblage of benthonic foraminiferal species (35 species) indicates outer shelf depths (400-600 feet) during deposition of the Marshalltown Formation (Figs. 2, 5). Prominent members of the benthic assemblage include *Gavelinella dumblei* (Applin), *G. nelsoni* (Berry), *Globorotalites micheliniana* (d'Orbigny), *Gyroidinoides cretacea* (Carsey), *G. depressus* (Alth) and *Neobulimina spinosa* (Cushman and Parker). Also indicative of outer shelf depths for the Marshalltown Formation is the well developed and diverse (22 species) assemblage of planktonic foraminiferal species, reflecting strong oceanic influence.

Regressive conditions returned during deposition of the Mount Laurel Formation; however, the shoaling did not equal that attained during deposition of the Englishtown Formation. The benthonic foraminiferal assemblage (25 species) as well as the planktonic assemblage (15 species) of the Mount Laurel Formation, while of low diversity, are more diverse than those of the Englishtown Formation (Figs. 4, 5). *Gavelinella compressa* (Olsson), *G. pinguis* (Jennings), and *Gyroidinoides cretacea* (Carsey) are species characteristic of the benthic assemblage, which suggests a midshelf (circa 300 feet) environment of deposition for the Mount Laurel Formation

(Fig. 2). The less diverse planktonic assemblage reflects the weaker oceanic influence of these depths.

CONCLUSIONS

Deposition of the Merchantville, Englishtown, Marshalltown, and the Mount Laurel formations was influenced by alternating transgressive and regressive cycles of sea level change. Paleoecological analysis of foraminiferal assemblages indicates that depths shifted between inner, mid, and outer shelf environments of deposition. These formations developed under distinct environments of deposition during each sea level change and are dated by biostratigraphic zonation of planktonic foraminifera and dinoflagellates. The integration of the sedimentologic, biostratigraphic, and paleoecologic data with the geophysical logs of boreholes in the study area allows recognition and correlation of these formations to be extended beyond the study area on the basis of geophysical correlation.

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PETROLOGY AND GEOCHRONOLOGY OF PERALKALIC METAGRANITE AND METARHYOLITE DIKES, FOUNTAIN QUARRY, PITT COUNTY, NORTH CAROLINA

RICHARD L. MAUGER *Department of Geology, East Carolina University, Greenville, North Carolina 27834*

RICHARD K. SPRUILL *Department of Geology, East Carolina University, Greenville, North Carolina 27834*

MICHAEL T. CHRISTOPHER *Versar Inc., 6621 Electronic Drive, Springfield, Virginia 22151*

M. SHAFIQULLAH *Laboratory of Isotope Geochemistry, Department of Geosciences, University of Arizona, Tucson, Arizona 85721*

ABSTRACT

Fountain Quarry is in an isolated outcrop of peralkalic metagranite that is 30-40 km east of the Piedmont-Coastal Plain boundary. The granite includes quartz, alkali feldspars, aegirine and NaFe-amphiboles and a rare Ba(Fe,Mn)Ti silicate, bafertisite. Major element analyses and the rare-earth pattern are typical of other silicic peralkalic igneous rocks. Metamorphism is demonstrated by the following features: 1) very pure end-member compositions and low structural states of the alkali feldspars; 2) dimensional- and lattice preferred-orientations of quartz and alkali feldspars; 3) metamorphism of rhyolite and basalt (amphibolite) dikes that cut the granite; 4) greenschist-grade mineral assemblages in pelitic rocks that cut(?) the granite; 5) dimensional-preferred-orientation textures in pegmatitic quartz veins which cut the granite; and 6) a N12E vertical foliation in the granite and other rocks in the quarry. This foliation is parallel to local magnetic patterns. Regional magnetic patterns indicate that the granite was involved in the complex folding that affected the eastern Carolina volcanic slate belt.

The original alkali feldspars have completely recrystallized, resulting in redistribution of minor elements such as Rb, Sr, and Ba. Although the scatter is high, a Rb/Sr whole-rock scatterchron indicates an early Paleozoic original crystallization age. A single whole-rock sample not on the scatterchron probably recrystallized and lost Sr about 340 my ago. K-Ar ages on biotite (309 my) and microcline (257 my) are uplift-cooling ages that indicate times when the temperature fell below 250^{o}C and 200^{o}C respectively. Albite in the granite may have up to 92 x 10^{-12} (moles/gram) excess ^{40}Ar if its closure temperature is assumed to be the same as microcline. A hornblende age (474 my) may record an Ordovician age for peak metamorphic temperatures; but, an adjustment for probable excess ^{40}Ar would lower the hornblende age to the Acadian-Hercynian range. Metamorphism, major compressive deformation, and uplift at Fountain predate Alleghanian deformation and cooling ages in the Raleigh Belt and immediately adjacent areas.

The Fountain granite and Crossnore-type granites with peralkalic affinities in the Sauratown Mountains (NC) and Mount Rogers-Crossnore area of the Blue Ridge (NC and VA) were emplaced in association with late Precambrian and early Paleozoic rift zones which formed along the edge of the North American and African(?) continents. The Fountain silicic peralkalic magmas developed in association with large-scale rhyolitic volcanism which was initiated and sustained by sub-crustal emplacement of large basaltic magma bodies. The mafic magmas crystallized at depth to give the dominantly siliceous Carolina volcanic slate belt its more dense mafic keel.

INTRODUCTION

Fountain Quarry is located 0.8 km southeast of Fountain, North Carolina (Fig. 1).

Early efforts to quarry dimension stone ended in failure (Councill, 1954). Eventually the property was acquired by Martin Marietta Corporation. Production of crushed stone commenced in 1961, and this activity is continuing at the present time. Fluvial sands and Pleistocene marine-estuarine sediments (Snyder and Katrosh, 1979) overlie the crystalline rocks along the perimeter of the quarry (Fig. 2); however there was an original outcrop in what is now the southern half of the quarry (Fig. 2, this report; Richards, 1950, Parker, 1968). The nearest surface exposures of other crystalline rocks are a few tens of kilometers northwest and southwest (Fig. 1) of Fountain. Thus the quarry marks the easternmost surface exposure of crystalline rock in the central North Carolina Coastal Plain (Fig.1; see also Gleason, 1981).

At Fountain the crystalline rocks form a steep-sided, elongate ridge that trends N12E. The ridge trend is parallel to a pervasive northeast foliation in the quarry rocks, and it is also roughly parallel with magnetic anomaly trends in the area (Zietz and others, 1980). The ridge is about 400 m across (distance between 70 ft depth-of-overburden contours; Fig. 2) and its length exceeds 1000 m. Brown (1959) characterized the feature as a buried monadnock that rose about 120 m above the crystalline rocks in nearby areas.

PETROLOGY OF THE FOUNTAIN GRANITE

The most abundant rock in the quarry is a medium-grained, equigranular, light-grey peralkalic metagranite which we informally call the Fountain granite (Fig. 3). The granite is cut by dikes of amphibolite, metarhyolite, and basalt. The metarhyolites and the amphibolites (originally basaltic dikes) had intruded the granite prior to regional metamorphism, and the basalt dikes are presumed to be Mesozoic in age. There is a pervasive foliation in the granite and other quarry rocks (N12E, nearly vertical; line symbol marked f, Fig. 3) that is defined by the following features: aligned aegirine and NaFe-amphibole crystals; fine-grained, dark-colored, flattened xenoliths; tabular feldspar mineral aggregates; foliation in the amphibolites and in minor schistose metamorphic rocks; and schlieren-like zones in the granite. A detailed study of the non-granitic rocks from the quarry is in progress (Mauger, in preparation).

Major minerals of the granite (Table 1) include quartz, albite, microcline, aegirine, and NaFe-amphibole; minor phases include magnetite, calcite, fluorite, and bafertisite, a rare Ba(Mn,Fe)Ti fluorosilicate mineral (Mauger, in press). The alkali feldspars have very uniform, nearly pure end-member compositions, and they have small but significant iron contents (FeYYCa; atomic; Table 2, analyses 13 and 14). Both feldspars exhibit low structural states (Table 3). Rare perthitic grains are patchy,

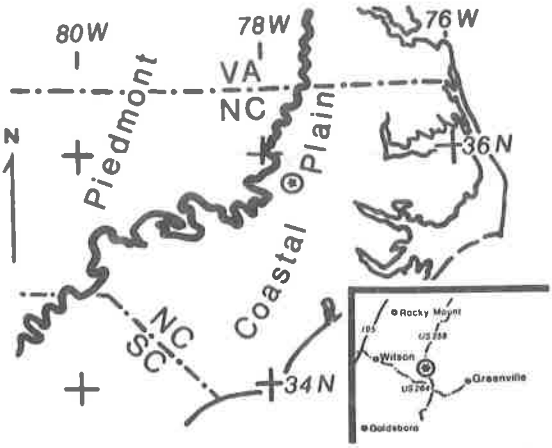

Figure 1. Index map showing the location and geologic setting of the Fountain granite. Fountain Quarry is marked by the ellipse with the star inside. Crystalline rocks are exposed at the surface west and north of the Piedmont-Coastal Plain boundary (heavy black line).

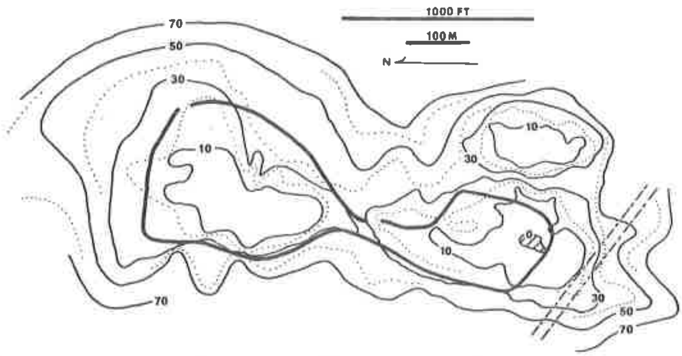

Figure 2. Contour map of the crystalline-rock surface at Fountain Quarry. The contours show depths from the surface (horizontal) to the crystalline rocks. Supplementary contours (10 foot intervals) are shown by dotted lines. The heavy line shows the perimeter of the quarry. Note the original outcrop of granite in what is now the south pit (area inside the zero contour line). The parallel dashed lines show county road 1240.

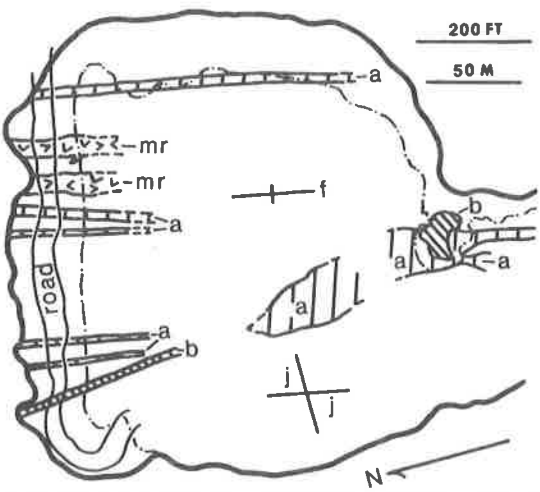

Figure 3. Geologic map of the north pit, Fountain Quarry. The very heavy line is the surface perimeter of the quarry. In most places it coincides with the contact between the Pleistocene sediments and the crystalline rocks. The dash-dot line marks the perimeter of the quarry floor. Rocks and symbols are as follows: amphibolite (a); basalt (b); metarhyolite (mr); foliation attitude (f); and joint-set attitudes (j). The remaining area (no pattern) is Fountain granite.

irregularly shaped intergrowths of albite in microcline. This type of perthite is generally ascribed to replacement rather than to exsolution (Smith, 1974, p. 415). The alkali feldspars form monomineralic aggregates of equidimensional granoblastic grains (Fig. 5). The aggregates, from 3 to 5 mm in length, have their long dimensions parallel to the foliation. Individual grains are generally from 1 to 2 mm in diameter. Quartz shows a strong dimensional-preferred orientation. Grains (1 to 2 mm in diameter) and grain aggregates (up to 5 mm in length) have their long dimensions aligned with the plane of foliation. Strain-induced subgrain boundaries are irregular in detail. They generally correspond to unit-prism forms and they form angles of 0 to 30 degrees with the direction of maximum grain elongation.

Long dimensions of acicular amphibole grains (1 to 3 mm) and of amphibole-pyroxene aggregates lie in the foliation plane. Although aegirine is commonly overgrown by amphibole, each mineral also occurs separately, and no sense of a

Table 1. Modal analysis of the Fountain granite.

quartz	33.0	NaFe-amphibole	4.1
albite	27.6	magnetite	0.3
K-feldspar	28.7	fluorite	0.1
aegirine	5.4	bafertisite	Trace

based on 700 counts of stained thin-sections

paragenetic sequence is evident. A few grains of bafertisite and fluorite also typically occur with the mafic silicates.

The pyroxene (Table 2, analyses 9 and 10) is aegirine (84 to 95 mole percent acmite) with a minor Wo content. The amphiboles (Table 2, analyses 1 through 8) are bluish NaFe-varieties with slightly different contents of K, Ca, Mg, Mn, and Ti. They are classified as ferroeckermannite, ferrorichterite, arfvedsonite, and riebeckite (Papike and others, 1974; Papike and Cameron, 1976). Some (ferroeckermannite, Table 2, analysis 1) have fluorine (0.5 wt%); others (arfvedsonite and riebeckite, analyses 5 and 6, Table 2) contain only minor fluorine (approx. 0.1 wt%). These amphiboles have 0.1 percent or less of ZnO and much lower levels of ZrO_2 (analyzed by microprobe). Similar amphiboles in Nigerian peralkalic granites have substantial ZnO and ZrO_2 contents (Borley, 1976).

Magnetite occurs as small anhedral blebs. It has a very pure end-member composition (Table 2, analyses 11 and 12), and the mole percentage of jacobsite ($MnFe(3^+)_2O_4$) exceeds that of ulvospinel. Calcite occurs as small blebs in the NaFe-silicate clumps and as thin intergranular films where it is associated with quartz and alkali feldspar. Calcites in the granite (Table 2, analysis 15), metarhyolite (Table 7, analysis 18), and amphibolites (Mauger, in preparation) typically contain substantial manganese, and lesser, but variable amounts of iron and magnesium.

CHEMICAL FEATURES OF THE FOUNTAIN GRANITE

The Fountain Granite has a comenditic composition according to the classification of Macdonald (1974a). Silica contents range from 72 to 76 percent (Table 4). High combined alkalis (8 to 10 wt%), relatively low Al_2O_3 contents, and Fe_2O_3 contents of 3.5 to 4.5 percent are typical features of comendite analyses (Table 4). The granite is quite homogeneous chemically; of course, some of the variations (Table 4) are the result of analytical factors. However most variations are probably due to differences in modal mineral composition; in particular, quartz and mafic silicate contents vary slightly from sample to sample. The variations in CaO probably reflect modal variations in minor phases such as calcite and fluorite. Note the uniformity in the TiO_2 contents (Table 4).

The average agpaitic index of the granite is 1.09. Strontium contents are very low; rubidium contents are from 105 to 130 ppm, and Rb/Sr ratios are high (Table 5). The granite exhibits extreme enrichment in LREE (C/C_0~100), a large negative europium anomaly, strong enrichment in HREE (C/C_0~40 to 50), and a relatively flat profile in the pattern of the HREE (Table 6 and Fig. 4). These are common features of rare-earth patterns from other comendites and peralkalic granites (Nicholls and Whitford, 1976; Ewart and others, 1976; Barberi and others, 1975; and Bowden and others, 1976).

METARHYOLITE

Two dikes of fine- to very fine-grained metarhyolite are exposed in the north-northeast part of the quarry (Fig. 3). The dikes are 3 to 4 m wide, and they are separated by about an equal thickness of granite. The dikes are essentially vertical, and their walls are roughly parallel to the N12E foliation (Fig. 3). The metarhyolite exhibits a much higher fracture density and a less regular fracture pattern than the granite. Locally small stringers of metarhyolite cut the granite, and xenoliths of granite are visible in the metarhyolite. Metarhyolite and amphibolite were not seen in mutual contact (Fig. 3).

Table 2. Microprobe analysis of minerals from the Fountain granite, Fountain, North Carolina.

Amphiboles (atoms per 23 oxygens)

	1	2	3	4	5	6	7	8
Si	7.866	7.967	7.902	8.000	8.000	8.000	8.000	8.000
Al(4)	0.134	0.033	0.098	-	-	-	-	-
Al(6)	0.200	0.329	0.253	0.236	0.231	0.174	0.457	0.404
Fe(2+)	4.350	4.397	4.415	4.014	3.165	2.837	4.081	4.312
Fe(3+)	-	-	-	0.528	0.903	1.354	0.216	0.008
Mg	0.325	0.314	0.435	0.425	0.745	0.572	0.289	0.357
Mn	0.224	0.194	0.199	0.128	0.116	0.134	0.197	0.202
Ti	0.190	0.065	0.064	0.021	0.011	0.014	0.023	0.064
Ca	0.303	0.305	0.345	0.145	0.279	0.186	0.263	0.247
Na(M4)	1.409	1.397	1.289	1.503	1.550	1.731	1.442	1.374
xoct	0.288	0.299	0.366	0.352	0.172	0.084	0.294	0.379
Na(A)	0.680	0.674	0.737	0.592	0.338	0.148	0.467	0.587
K(A)	0.283	0.296	0.270	0.104	0.056	0.027	0.257	0.247
SiO2	47.27	50.94	49.23	49.37	49.08	51.25	49.35	52.07
Al2O3	1.70	1.96	1.86	1.24	1.20	0.94	2.39	2.23
FeO	31.25	33.62	32.89	33.52	29.85	32.10	31.69	33.62
MgO	1.31	1.35	1.82	1.76	3.07	2.46	1.19	1.56
MnO	1.59	1.46	1.46	0.93	0.84	1.01	1.44	1.55
TiO2	1.52	0.56	0.53	0.18	0.09	0.12	0.19	0.55
CaO	1.70	1.82	2.01	0.84	1.60	1.11	1.52	1.50
Na2O	6.47	6.83	6.51	6.67	5.97	6.21	6.08	6.58
K2O	1.33	1.49	1.32	0.50	0.27	0.14	1.24	1.26
sum	94.14	100.02	97.63	95.01	91.96	95.34	95.32	101.18
QUAD	3.7	2.9	-	24.9	22.5	13.5	27.6	16.6
Ca	6.1	6.1	-	3.2	6.7	5.2	5.7	5.0
Mg	6.5	6.3	-	9.3	17.8	15.9	6.2	7.3
Fe2	87.4	87.7	-	87.6	75.6	78.9	88.1	87.7
OTHERS	96.3	97.1	100.0	75.1	77.5	86.5	72.4	83.4
A	38.4	40.4	42.1	31.7	20.3	9.2	33.4	37.8
NaM4	56.2	58.2	53.8	68.3	79.7	90.8	66.6	62.2
Al4	5.3	1.4	4.1	-	-	-	-	-
NAME	Fe eck	Fe eck	Fe rich	arfve	arfve	rieb	Fe eck	Fe eck

For abbreviations see Papike and others (1974) and Papike and Cameron (1976).

pyroxenes
atoms per 6 oxygens

	9	10
TET		
Si	1.981	1.978
Al(4)	0.019	0.013
OCT		
Al(6)	0.013	-
Fe(2+)	0.107	0.044
Fe(3+)	0.820	0.895
Mg	0.020	0.020
Mn	0.023	0.020
Ti	0.006	0.003
M2		
Ca	0.186	0.149
Na	0.826	0.888
Chemical Analyses		
SiO2	53.07	52.94
Al2O3	0.72	0.29
FeO	29.70	30.06
MgO	0.36	0.36
MnO	0.72	0.63
TiO2	0.22	0.09
CaO	4.64	3.72
Na2O	11.41	12.25
K2O	0.02	0.02
sum	100.86	100.36
Summary		
QUAD	16.1	10.2
Wo	59.4	69.9
En	6.4	9.4
Fs	34.2	20.7
OTHERS	83.9	89.9
Ti	0.7	0.3
NaM2	97.0	98.3
Al4	2.3	1.4
NAME	Acmite	Acmite

magnetite
atoms per 4 oxygens

	11	12
Ca	0.004	0.005
Mg	-	0.001
Mn	0.014	0.015
Ti	0.008	0.007
Al	-	0.003
Fe(2+)	0.990	0.985
Fe(3+)	1.984	1.983
Summary		
mag	97.4	97.2
ulvo	0.8	0.7
*Mn	1.4	1.5
*Ca	0.4	0.5
*Mg	-	0.1
	0.14	0.26
	0.10	0.60
	94.04	94.40
	0.00	0.02
	0.45	0.46
	0.28	0.26
	0.09	0.14
	0.01	0.03
	0.06	0.07
	95.17	96.24

*Mn is $MnFe(3+)_2O_4$
*Ca is $CaFe(3+)_2O_4$
*Mg is $Mg[Al,Fe(3+)]_2O_4$

For pyroxene abbreviations, see Papike and others (1974) and Papike and Cameron (1976).

feldspars
mole percentages

	13	14
Ab	97.5	2.5
Or	0.7	96.3
An	0.1	0.2
*Fe	1.7	1.0
Chemical Analyses		
SiO2	67.10	64.59
Al2O3	18.63	17.82
FeO	0.47	0.27
CaO	0.03	0.03
Na2O	11.33	0.28
K2O	0.12	16.58
sum	97.67	99.58

*Fe is $FeAl_2Si2O8$

calcite
mole percentages

	15
cal	91.9
rhod	6.1
sid	2.0
Chemical Analyses	
FeO	1.5
MnO	4.4
CaO	52.9
sum	58.9

Table 3. Structural states of alkali feldspars from the Fountain granite.

reflection	2θ microcline	*2θ max microcline
$\bar{2}01$	21.0	21.0
060	41.8	41.8
$\bar{2}04$	50.5	50.5(5)

reflection	2θ albite	*2θ low albite
$\bar{2}01$	22.1	22.0(5)
060	42.5	42.5
$\bar{2}04$	51.2	51.1(5)

microcline Ab 3.5, Or 95.3, An 0.1, *Fe 1.1
albite Ab 97.4, Or 0.6, An 0.0, *Fe 2.0

*two-theta values from Wright (1968)

the microcline locally shows the typical cross-hatched twinning pattern; triclinicity (Δ) based on the 130 and $1\bar{3}0$ reflections is slightly greater than one, which also confirms the low structural state (maximum microcline) for the K-feldspar.

two-theta values are for Cu K_{α} radiation.

Table 4. Chemical analyses of the Fountain granite and metarhyolite.

	Fountain Granite								Metarhyolite	
	10-25-5	11-10-5	11-10-3	6-1-15	2-10-1	2-10-9	6-1-2	10-7-4	1-4-78a	1-4-78b
SiO_2	74.9	74.5	73.9	74.5	75.7	74.0	75.0	72.2	77.1	78.4
TiO_2	0.21	0.21	0.20	0.20	0.22	0.24	0.20	0.23	0.18	0.16
Al_2O_3	10.6	10.5	11.0	11.6	11.2	--	--	--	12.9	12.9
Fe_2O_3	3.6	4.5	3.7	3.8	3.4	3.8	3.7	3.8	2.4	1.9
MnO	0.11	--	0.12	--	0.14	0.13	0.15	0.03	0.2	0.1
MgO	0.06	0.10	0.16	0.22	0.04	0.06	0.08	0.02	0.20	0.09
CaO	0.26	0.78	0.89	0.47	0.69	0.61	0.75	0.25	0.61	1.23
Na_2O	4.6	4.8	4.3	3.7	4.5	4.8	4.2	4.0	4.1	3.7
K_2O	3.9	4.3	4.6	4.7	4.1	4.2	4.1	4.2	3.5	4.2
P_2O_5	<0.02	--	<0.02	--	--	--	--	--	<0.02	0.03
CO_2	<0.1	<0.1	0.2	--	--	--	--	--	0.3	0.6
F	0.044	0.034	0.034	0.036	--	--	--	--	0.020	0.018
LOI	0.5	0.5	[0.3]	0.4	--	--	--	--	[0.3]	[0.6]
sum	98.8	100.2	99.2	99.6	100.1	--	--	--	101.5	103.3

Analysis 10-25-5 and MgO, Na_2O, P_2O_5, MnO, CO_2, F, and LOI in 11-10-5, 11-10-3, 6-1-5, 2-10-1, 1-4-78a and 1-4-78b were done by Skyline Labs, Inc., Wheat Ridge, Colorado (AA and wet chemical methods). Analyses 2-10-9, 6-1-2 and 10-7-4 were done at the Department of Geology, University of North Carolina at Chapel Hill (AA and spectrophotometer, T. Sando, analyst). The other results are from the Department of Geology, East Carolina University (X-ray fluorescence). [LOI] indicates that the sums were corected for the LOI that was CO_2. Total iron is reported as Fe_2O_3, and the dash (--) indicates that the oxide (element) was not determined.

*Table 5. Rb-Sr *data for the Fountain granite.*

sample	Rb(ppm)	Sr(ppm)	$^{87}Sr/^{86}Sr$	$^{87}Rb/^{86}Sr$
2-10-1	106.7	35.8	0.7746	8.67
2-10-9	110.7	27.8	0.7891	11.64
6-1-2	111.7	26.5	0.7921	12.29
10-7-4	127.4	7.9	0.9700	48.10
F-1	106.4	20.5	0.8236	15.15
F-12	111.6	20.4	0.8257	16.04
F-10**	4.2	58.7	0.7604	0.21

* R.K. Spruill, analyst

$\lambda = 0.142 \times 10^{-10}\ yr^{-1}$

** aegirine - NaFe amphibole mixture;

P. Fullagar, analyst

Rb and Sr by isotope dilution; routine experimental errors (at 1 level) are estimated as < 1% in Rb/Sr and < 0.05% in 87Sr/86Sr. Slope (Figure 5) calculated using least-squares regression model of York (1969).

Table 6. Rare-earth elements in 2-10-1 sample of the Fountain granite, Fountain, North Carolina.

elem	ppm	C/Co	elem	ppm	C/Co
Ce	103.08	126.79	Dy	14.96	46.03
Nd	71.79	120.25	Er	8.51	39.95
Sm	16.57	86.30	Yb	8.02	38.56
Eu	2.71	37.53	Lu	1.23	38.08
Gd	16.01	61.81			

C/Co is chondrite normalized ratio using Leedey chondrite values (Masuda and others, 1973) divided by 1.20. Analyses were done by isotope dilution at the Department of Geology, University of North Carolina, Chapel Hill, North Carolina; R.K. Spruill, analyst.

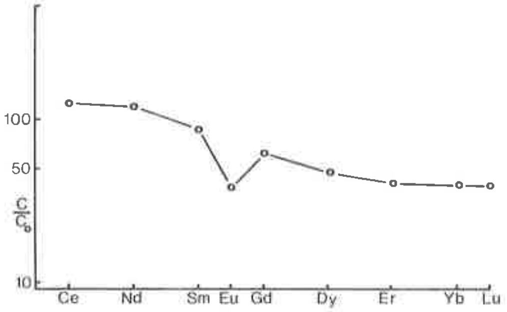

Figure 4. Rare-earth element pattern for the Fountain granite.

PETROLOGY OF THE METARHYOLITE

Alkali feldspars in the metarhyolite have fairly pure end-member compositions (Table 7), but the albite has a small An-component (2 to 5 mole percent) and Fe<Ca (atomic). The microcline composition is very similar to that in the granite (Table 7, analyses 1 through 5). Very coarse poikilitic albite (Table 7, analysis 13), associated with quartz veinlets in the metarhyolite, has a lower An content than the typical fine-grained albite. Minor phases in the metarhyolite include calcite with notable Mn, Mg, and Fe contents (Table 7, analysis 18), magnetite with virtually a pure end-member composition (Table 7, analysis 21), phengite, chlorite, biotite, and stilpnomelane (Table 7, analyses 22 through 27). These latter phyllosilicates amount to only a few percent of the rock. However they do exhibit a distinctive planar alignment that is parallel to the pervasive foliation in the granite (Fig. 6). Quartz and alkali feldspars also exhibit the same textural fabrics as they show in the granite.

The metarhyolites are locally cut by quartz veinlets that contain epidote, coarse-grained albite, and a few poikiloblastic pinkish-orange garnets (Fig. 6). These garnets are "post-tectonic" (Spry, 1969, p. 257). The garnets are rich in spessartine; the other major components are almandine, andradite, and grossularite (Table 7, analyses 19 and 20). Garnets with similar compositions are reported (Speer and others, 1980) in the Castalia granite (36.02°N, 78.05°W; Fig. 7) and in greenschist grade metamorphosed rhyolitic rocks from localities in the main Carolina slate belt (Fodor and others, 1980; Seiders, 1978).

CHEMISTRY OF THE METARHYOLITE

Normative corundum (approx. 2 mole%; Table 4) and aluminous phases such as garnet and micas indicate that the metarhyolite is peraluminous. If the metarhyolite had been metasomatically altered by fluids derived from the peralkalic granite, sodium would have been added in excess of aluminum, and normative corundum in the metarhyolite would have been reduced. Thus, the original rhyolite was almost certainly peraluminous. Although the SiO_2 contents (Table 4) include a contribution from the

Table 7. Compositions of minerals from the metarhyolite at Fountain Quarry, Fountain, North Carolina.

Feldspars (mole percentages)

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17
Ab	2.4	3.9	2.4	3.5	3.6	3.2	95.1	93.6	97.1	93.5	96.1	95.1	98.4	2.5	95.8	2.5	96.4
Or	95.7	95.6	95.7	96.2	96.0	95.8	0.9	0.5	0.6	1.0	0.6	0.7	0.6	97.5	1.0	97.2	1.3
An	0.3	0.2	0.2	0.2	0.1	0.2	3.7	5.3	1.8	5.2	2.6	3.7	0.9	0.0	3.1	0.1	2.1
Fe	1.7	0.3	1.7	0.1	0.3	0.8	0.3	0.6	0.6	0.2	0.6	0.5	0.1	0.0	0.0	0.2	0.2

1 to 5 are matrix microcline; 6 is average of 1 to 5; 7 to 11 are matrix albite; 12 is average of 7 to 11; 13 is coarse-grained poikilitic albite; 14-15 and 16-17 are perthitic pairs; 14 and 16 are the exolved phase, 15 and 17 are the host phase.

Calcite (mole percentages) 18; 90.0 cal, 8.0 rhod, 1.4 sid, 0.6 magn

Garnet (atoms per 12 oxygens) — mole percentages end members

	X	Y	Z	gros	pyro	spes	alma	andr
19	0.074 Mg 1.773 Mn 0.500 Fe2+ 0.653 Ca	0.205 Fe3+ 1.786 Al 0.041 Ti +0.032 *R	2.903 Si 0.097 Al	12.2	2.5	58.6	16.5	10.2
20	0.070 Mg 1.639 Mn 0.802 Fe2+ 0.488 Ca	0.146 Fe3+ 1.860 Al 0.011 Ti +0.017 *R	2.951 Si 0.049 Al	9.4	2.3	54.4	26.6	7.3

Ti is assumed to be 3+. For the end-member calculations, Ti is combined with Al(6), and Al(4) is combined with Si. Residuals (*R) are accumulated in the Y-site.

Magnetite (mole percentages) 21; 98.2 mag, 0.6 ulvo, 0.5 *Mn, 0.6 *Ca, 0.1 *Mg

Micas (atoms per 22 oxygens) and Chlorite (atoms per 28 oxygens)

	K	Na	Ca	Mg	Fe	Ti	Mn	Al(6)	Si	Al(4)
22	1.896	0.047	0.005	0.374	0.718	0.082	----	2.819	6.359	1.641
23	2.066	0.023	0.027	1.975	2.934	0.303	0.142	0.327	5.560	2.440
24	0.009	0.018	0.018	4.503	4.687	0.012	0.107	2.561	5.612	2.388

In 22 (phengite), iron is calculated as 3+; in 23 (biotite) and in 24 (chlorite), iron is calculated as 2+.

Stilpnomelane (atoms per 8 silicons)

	K	Na	Ca	Mg	Fe	Ti	Mn	Al	Al/Al+Si
25	1.762	0.129	0.019	1.869	3.268	0.004	0.852	1.244	0.135
26	0.795	0.065	0.026	1.961	2.985	0.004	0.767	1.212	0.132
27	0.575	0.064	0.033	1.965	2.888	0.003	0.761	1.240	0.134

Note the common Al/Al+Si and the large differences in potassium.
Al/Al+Si in biotite ranges from 0.25 to 0.41.

*Table 7 (continued). *Chemical analyses of minerals from the metarhyolite at Fountain Quarry.*

	2	10	18	19	20	21	22	23	24	25	26
SiO_2	64.07	64.96	0.00	34.74	35.28	0.00	46.76	34.98	25.86	42.71	46.59
TiO_2	0.00	0.00	0.00	0.66	0.17	0.21	0.80	2.54	0.07	0.03	0.03
Al_2O_3	18.34	20.65	0.00	19.11	19.37	0.07	27.82	14.76	19.35	5.63	5.99
FeO	0.08	0.07	1.08	10.08	13.55	92.92	6.31	22.07	25.83	20.87	20.79
MnO	0.00	0.00	6.03	25.05	23.14	0.15	0.04	1.05	0.59	5.38	5.27
MgO	0.00	0.00	0.26	0.60	0.56	0.02	1.85	8.34	13.92	6.70	7.66
CaO	0.04	1.08	53.57	7.30	5.45	0.15	0.03	0.16	0.08	0.09	0.14
Na_2O	0.44	10.59	0.00	0.05	0.11	0.04	0.18	0.07	0.04	0.36	0.20
K_2O	16.30	0.18	0.00	0.03	0.04	0.06	10.93	10.19	0.03	7.37	3.63
H_2O	-	-	-	-	-	-	(5)	(5)	(14)	(11)	(8)
sum	99.27	97.53		97.62	97.67	93.62	94.72	94.16	85.77	89.14	92.17

The values for (H_2O) are estimated by difference from 100%.
0.00 means that sample counts ≤ background counts. Total iron is reported as FeO

* numbers at heads of columns refer to samples (numbers underlined) for which mole percentages or ration proportions are given earlier in the table

small quartz-carbonate veinlets, the rocks are still highly siliceous. In sample 1-4-78a (Table 4), Na_2O exceeds K_2O. This is a typical characteristic of silicic volcanic rocks from the main Carolina slate belt (Seiders, 1978; Butler and Ragland, 1969). Potash exceeds Na_2O in Sample 1-4-78b, but Na>K (atomic) (Table 4).

XENOLITHS AND SCHLIEREN

The granite contains a small percentage of fine- to very fine-grained dark inclusions (Fig. 5B). These have minerals and mineral compositions like those in the granite; however magnetite and amphibole are more abundant. The inclusions show the N12E foliation. They are probably made-over xenoliths of mafic volcanic rocks that were incorporated while the granite was being intruded. Some darker streaks that resemble schlieren are also visible in the granite. These generally are aligned with the foliation.

PEGMATITIC QUARTZ VEINS

A few pegmatitic quartz veins occur in the granite. The most prominent ones were found at the base of the east wall of the narrow zone that joins the north and south pits (Figs. 2 and 3). The veins are mostly medium-grained quartz, and the grains show a strong dimensional-preferred orientation (Fig. 5D) similar to that which is typically seen in metaquartzites. The veins contain a few very coarse-grained crystals of ilmenite and light-pink microcline. Locally thin films of highly foliated micaceous minerals occur in the quartz. There are also quartz veins in the amphibolites. They contain foliated micaceous clots and the quartz grains show a dimensional-preferred orientation. These and the pegmatitic quartz veins are interpreted as pre-deformational.

OTHER QUARTZ VEINS

Numerous other small quartz veins and pods in the quarry have very coarse grains ($\geq$5 mm) which don't show a dimensional-preferred fabric (Fig. 5E and F). These veins also contain angular fragments of the foliated granite. Thus we believe that they post-date the major period of recrystallization and deformation. The post-deformational veins contain a complex assemblage of minor phases. These include coarse bladed crystals of NaFe-amphibole, pale-yellow sphalerite, molybdenite, fluorite, calcite (Fe- and Mn-bearing), bafertisite and an unidentified grey-brown BaFeMnTi silicate. This latter material has substantial Y (1500 ppm), Sr (1000 ppm), La (500 ppm), and Zr (500 ppm) as estimated by emission spectrographic methods. Locally streaked-out and lineated minerals along rupture surfaces in these veins record a late stage of brittle deformation (Mauger, in prep).

JOINTING PATTERNS IN THE QUARRY

Christopher (1979) measured the attitudes of more than 350 joints and closely spaced groups of joints in the quarry. He studied local joint patterns in various parts of the quarry and found only minor differences in the different areas. Thus his data are probably representative of joints in the whole granite body. There are two major, nearly vertical, joint sets striking N12E and N89E (Fig. 3, line symbols marked j). These sets are defined by major, through-going fractures. The N12E set is essentially parallel to the foliation. Many of the main near-vertical joint surfaces have coatings of quartz, albite, carbonates, NaFe-amphiboles, biotite, and other minerals. A third, less prominent set of shallow-dipping to horizontal joints is evident. These are not mineralized, and they probably are sheeting-type fractures that formed before the granite was buried in Plio-Pleistocene time.

The two near-vertical joint sets at Fountain have similarities to longitudinal joints (N12E set) and cross joints (N89E set) as defined by Balk (1937). He deduced that cross joints in plutonic rocks open soon after solidification because they typically carry veneers of late-stage igneous or hydrothermal minerals. Longitudinal joints open along planes of flow foliation in a pluton, and the cross joints open at right angles thereto

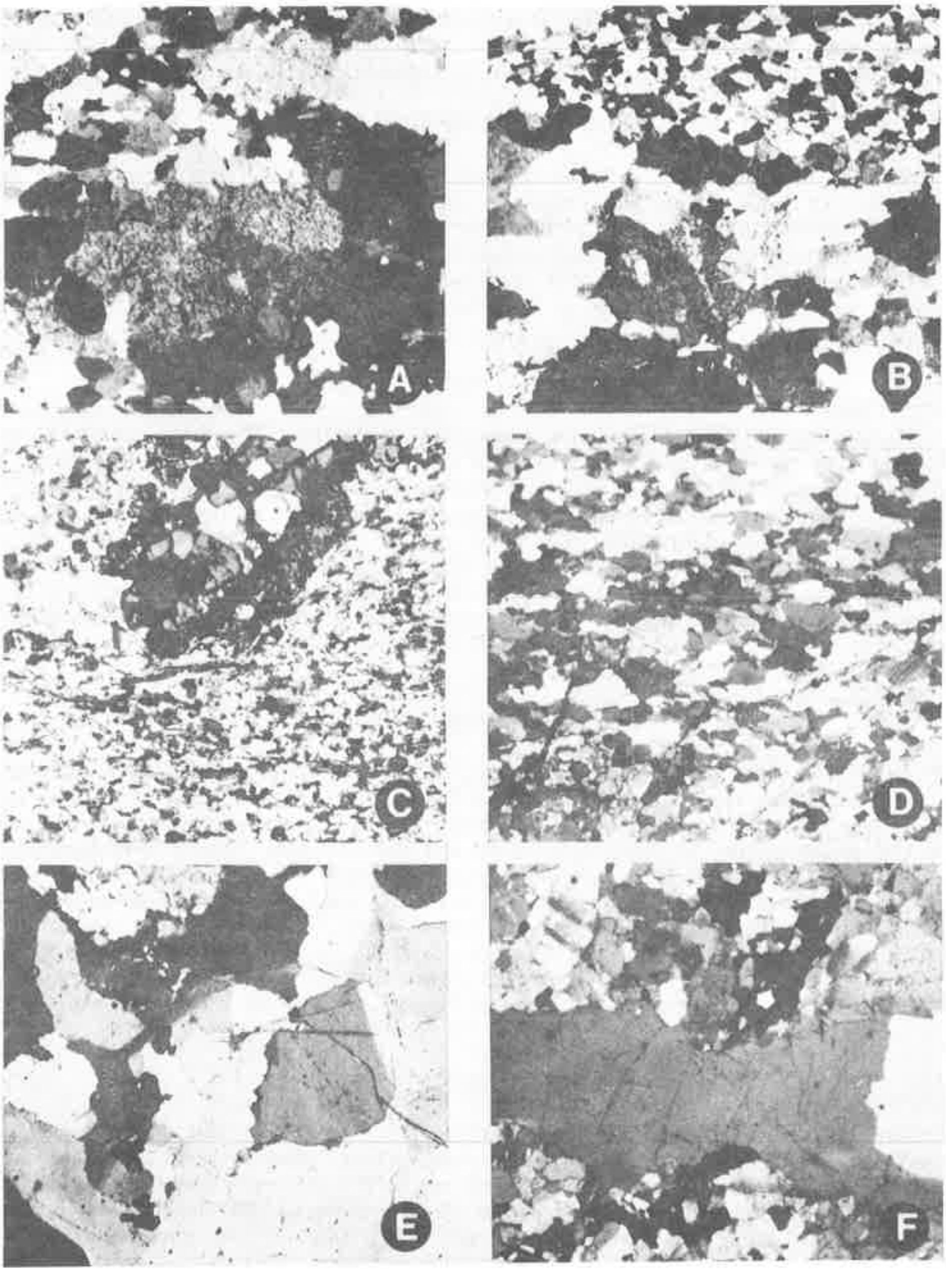

Figure 5. Photomicrographs of Fountain granite and associated rocks. A. Fountain granite. Grey to dark-grey grains (lower one-half) are microcline in a multi-grain aggregate. Twinning is faintly visible. Other smaller grains (white to dark grey) are quartz or albite. Black grains (right-center top and center-left edge) are aegirine or NaFe-amphibole. Photo width, 2.49 mm; diameter of black circle with white A, 0.285 mm. B. Fountain granite with fine-grained inclusion. Inclusion (top) tapers to point at upper left. Light-colored grains are albite, microcline and quartz; dark grains are aegirine, NaFe-amphibole, and magnetite. Large microcline aggregate (center to lower center) is part of normal granite (coarser grains). Black grain at lower edge is aegirine. Photo width, 2.49 mm; diameter of black circle with white B, 0.285 mm. C.

(Balk, 1937). During metamorphism and recrystallization of the Fountain granite, maximum extensional strain was in the N12E foliation plane and maximum compressive stresses were oriented (SE-NW) at right angles to the foliation. Although the foliation in the granite is of metamorphic origin, the N89E joints may be similar to cross joints because they opened at roughly right angles to the direction of maximum extension. The N12E set probably opened when the compressive stresses were relaxed. These near-vertical joints were opened and mineralized before the region was uplifted and cooled, about 260 my ago as recorded by the K-Ar date on microcline (Table 9).

CONDITIONS OF METAMORPHISM AT FOUNTAIN

Mineral assemblages at Fountain (Table 8) were probably equilibrated at temperatures in the middle to upper greenschist range (300 to 450°C). Diopsidic augite (Wo 50, En 30, Fs 20) in the amphibolites may indicate that peak temperatures were close to 500°C. Pressures are not well constrained (Table 8). Solvus alkali feldspar compositions approach pure end-member compositions with increasing pressure (Goldsmith and Newton, 1974) and with lower temperatures. The Fountain alkali feldspars could have formed at high pressures if equilibration temperatures were about 300°C. Conversely the pure end-member compositions could have originated by equilibration at lower temperatures and much lower pressures. Other criteria (Table 8) are compatible with moderate pressures, except for the acmite-jadeite proportions in aegirine (see Fig. 4, Popp and Gilbert, 1972). The Fountain aegirine has little, if any, Al in six-fold coordination (Table 2, analyses 9 and 10), so the Popp and Gilbert data give unrealistically low pressures (<<1 kbar) of metamorphism. Perhaps the Wo content of the aegirine and/or the presence of NaFe-amphibole invalidate applying this experimental data to the Fountain granite.

POTASSIUM-ARGON RESULTS

Mineral separates from the quarry have different K-Ar ages (Table 9). Although the ages (hornblende > biotite, "plagioclase" > microcline) are in the order that is generally accepted for cooling ages (Hanson and others, 1975; Harrison and others, 1979), there are some important discrepancies. First, the difference between the hornblende and biotite ages is very large. Second, the low structural states and pure end-member compositions suggest that the alkali feldspars continued to equilibrate

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Metarhyolite with garnet-bearing quartz veinlet. Edge of veinlet is marked by transition from massive to poikiloblastic garnet (upper center). Finer grains in metarhyolite are quartz, microcline, and albite. Note the foliation imparted by the long, thin muscovite (phengite) flakes (lower one-half). Photo width, 2.49 mm; diameter of black circle with white C, 0.285 mm. D. Metamorphosed pegmatitic quartz vein. All grains are quartz. Note tendency for long dimensions of grains to lie more-or-less parallel to the long dimension of the photo. Dark crack (from lower-left corner to middle left) is a seam of foliated stilpnomelane. Compare with 5E and 5F. Photo width, 5.72 mm; diameter of black circle with white D, 0.66 mm. E. Post-deformational quartz vein. Larger grains are all quartz. Finer-grained aggregates are Fountain Granite (angular fragment in quartz vein, left center to lower-left center; granite in vein wall, upper left to left center). Black grains in granite are aegirine or NaFe-amphibole. On the left side of the granite fragment, a vein branches off the main vein (toward the upper left). Note healed microcracks and minor strain-induced subgrain texture in quartz. Larger grains (not shown) in this vein are up to 5 mm in diameter. Photo width, 5.72 mm; diameter of black circle with white E, 0.66 mm. F. Post-deformation quartz vein. This vein branches off from the main vein shown in figure 5E. Minerals in granite include albite (prominent grid twinning), microcline, quartz, and aegirine (black). Note large single-crystals and healed microcracks in quartz vein. Tracings of opposing vein walls fit reasonably well by shifting lower wall to right; however, 3-dimensional fit is not known. Thus vein is probably dilational in origin. Photo width, 5.72 mm; diameter of black circle with white F, 0.66 mm.

Table 8. Estimated P,T conditions for metamorphism of the crystalline rocks at Fountain.

assemblage/relationship	P(kbars)	T °C	reference
alkali feldspar compositions in granite and metarhyolite	high(?)	200 to 400	1,7,8
albite-pargasite (Fe pargasite) diopside-sphene-epidote-biotite-magnetite(pure) assemblage in amphibolites	-	350 to 500	
stabilities of albite + quartz/jadeite	≤*12	300 to 400	1
upper thermal stability limit of stilpnomelane with muscovite and biotite	≤10	<*480	2
biotite-chlorite-phengite-stilpnomelane assemblage in pelitic clots	-	350 to 500	
phengite thermal stability (0.3 mole% FeMg mica)	>*2	350	3
100% acmite in acmite-jadeite portion of aegirine	"low"	-	4
Mn-rich garnets in blue schist terrains but with more almandine than at Fountain	"high"	"low"	5,6
thermal stability limits of ferrous amphiboles	-	<500	9
NaFe-amphiboles in low to middle green-schist rank rocks elsewhere	-	300 to 450	11
features that indicate extensive cold working of quartz in granite and in pegmatitic veins	-	400	10

*indicates derived P (or T) associated with a chosen value or range in T (or P)

references

(1)Goldsmith and Newton, 1974; (2)Winkler, 1976; (3)Velde, 1965; (4)Popp and Gilbert, 1972; (5)Lee and others, 1963; (6)Newton and Fyfe, 1976; (7)Robin, 1974; (8)Champness and Lorimer, 1976; (9)Charles, 1977; (10)Liddell and others, 1976; (11)Onaki and Ernst, 1969.

down to temperatures (approx. 200°C) where argon is retained. Thus it is unlikely that the albite age (292 my) should be significantly older than the microcline age. Third, there is a major difference between potassium contents in the albite mineral separate (1.14%, Table 9) and in the albite grains (0.07 to 0.10%; measured by microprobe, Table 10). Microscopic inspection of sodium cobaltinitrite-stained grains of the albite mineral separate showed the presence of about 9 percent microcline in the form of albite-microcline aggregates. Potassium mass-balance calculations (Table 10, line b) give 8.8 percent microcline in the albite separate. Thus the albite K-Ar age (Table 9) is for a mixture of albite (91%, 0.07 to 0.10% K) and microcline (9%, 12.12% K). Of the radiogenic Ar in the albite separate (630.6×10^{-12} m/g; Table 9; Table 10, line c), 509.8×10^{-12} m/g was contributed by the microcline impurity. The pure albite thus contributed 132.5×10^{-12} m/g. Using a potassium content of 0.085% (Table 10) would give the albite a K-Ar age of 729 my. This age is obviously much too high and it suggests that extra radiogenic ^{40}Ar is present in the pure albite phase. If the two alkali feldspars had exactly the same closure temperatures, corresponding to the microcline age of 257 my, the albite would contain about 92×10^{-12} m/g of excess ^{40}Ar (Table 10, line d). If the albite closed at the same temperature as biotite (309 my ago), the albite would contain about 82×10^{-12} m/g excess ^{40}Ar. These amounts

Table 9. K-Ar data for minerals from the Fountain Quarry, Fountain, North Carolina.

sample	K%	rad ^{40}Ar x 10^{-12} m/g	^{40}Ar % atm	Age(my)
microcline UAKA 79-156	12.15	5797	1.6	257 ± 5
	12.14	5791	1.6	
	12.06	5790	1.7	
	12.12	5793	1.6	
albite UAKA 79-158	1.129	634.4	5.0	292 ± 6
	1.130	628.4	5.4	
	1.189	629.5	5.0	
	1.149	630.2	4.9	
		630.6	5.1	
hornblende UAKA 79-157	0.237	223.2	12.5	474 ± 10
	0.237	223.4	13.1	
	0.238	222.6	25.1	
	0.236	223.1	16.9	
	0.237			
	0.237			
biotite UAKA 81-97	7.888	4644	1.2	309 ± 7
	7.965	4640	2.0	
	7.910	4666	1.7	
	8.038	4637	1.1	
	7.950	4645	1.1	
		4646	1.4	

average values are underlined; the albite and microcline are from the Fountain granite. The hornblende is from a thin (approx 0.5 cm) vertical vein of coarse-grained hornblende that cuts the granite. This vein is interpreted as a thin basalt dike that was metamorphosed along with the granite. The biotite is from a thin (approx 10cm) marble-pelitic vein of unknown origin. It is possibly a metamorphosed lamprophyre dike (Mauger, in prep)

constants

$\lambda_{\beta} = 4.963 \times 10^{-10}\ yr^{-1}$; $\lambda_{\epsilon} = 0.581 \times 10^{-10}\ yr^{-1}$

$\lambda = 5.544 \times 10^{-10}\ yr^{-1}$; $^{40}K/K = 1.167 \times 10^{-4}$ atom/atom

Table 10. Experimental and calculated values used in the evaluation of extra radiogenic ^{40}Ar in albite at Fountain Quarry.

a. potassium contents
microcline mineral separate 12.120% (Table 9)
albite mineral separate 1.149% (Table 9)
*albite (pure phase) 0.085%
*average of numerous microprobe determination on albite from different samples

b. mineral composition of albite separate
**91.2% albite, **8.8% microcline

c. ^{40}Ar (radiogenic)
albite mineral separate 630.6 x 10^{-12} m/g (Table 9)
microcline mineral separate 5793 x 10^{-12} m/g (Table 9)
albite (pure phase) **132.5 x 10^{-12} m/g

d. extra ^{40}Ar in albite (pure phase) assuming that the albite and microcline ages are equal (257 my).

**92 x 10^{-12} m/g or **69.4% of total (line c, this table)

**means values are calculated; other values cited in this table are experimental (measured).

are comparable to those found in albite (20 to 170 x 10^{-12} m/g) and quartz (100 x 10^{-12} m/g) from Amelia, Virginia (Laughlin, 1969); similar quantities of extra ^{40}Ar in the Fountain biotite and hornblende would bring the ages into closer agreement because the hornblende age is much more sensitive to excess ^{40}Ar (Table 11).

The dated hornblende occurs as slabs of long (10 cm or more) slightly bent to locally contorted, acicular, bladed crystals with closely aligned long dimensions. This material forms a thin (a few centimeters thick) vein that cuts the granite. The vein is roughly parallel to the amphibolite dikes in the quarry (Fig. 3). It may represent a metamorphosed mafic dike, or it may be mechanically deformed vein-filling material that crystallized originally under static conditions following the main deformational and metamorphic event (Mauger, in preparation). There was a very late stage of brittle deformation in the quarry that produced slickensides and lineated acicular mineral grains on some rupture surfaces (Mauger, in preparation) and this deformation affected the dated hornblende. In either case, the hornblende crystallized during or following the major deformational and metamorphic event. Interpretation of the K-Ar age (Table 9) involves many considerations (Harrison and McDougall, 1980a; 1980b). The most important are probable excess environmental ^{40}Ar (Damon, 1968) and the fact that hornblende retains argon at higher temperatures (450 to 500°C) than does biotite (250°C). The Mg/Fe ratio also affects argon retention in amphiboles (O'Nions and others, 1969). If the hornblende age (474 my) is assumed to be real, the hornblende-biotite age difference would have to represent a prolonged interval of slow cooling from about 500°C to about 250°C, or it could be due to a thermal pulse, about 300 to 350 my ago, that degassed the biotite without affecting the hornblende (Harrison and others, 1979; Dallmeyer, 1975). Most of the difference between the two ages could be due to excess environmental ^{40}Ar.

Rb-Sr ISOTOPIC STUDIES

A five-point scatterchron (Table 5, Fig. 6) gives a probable early Paleozoic original age (531 my) for the Fountain granite. Uncertainties in the age ($\pm$60 my) and initial ratio are far in excess of those attributable to routine experimental errors (Fig. 6). The large scatter is due to geologic factors, the most important being feldspar recrystallization and mobilization of Rb and Sr. The original granite would have contained quartz, alkali feldspars that may have been unmixed to some extent, NaFe-amphibole and/or NaFe-pyroxene, and titanomagnetite as major phases. Most of the Rb and some of the Sr were probably in the original alkali feldspar. As recrystallization proceeded, Rb and part of the Sr were incorporated into the microcline. Highly radiogenic Sr (Table 5) was incorporated into the NaFe-silicates. The magnetite recrystallized, releasing Ti. Ba and the remaining Sr were available to form the rare BaFeTi silicates (approx. 0.1% SrO, Mauger, in press) in the granite and post-deformational quartz veins. The whole-rock samples were open systems with respect to Sr and Rb. Pre- and post-recrystallization Rb/Sr ratios and Sr isotopic compositions were roughly preserved for most whole-rock samples, at least to the extent that the whole-rock isochron was evidently retained as a scatterchron. Other whole-rocks were affected more drastically. For example 10-7-4 (Table 5) does not even fall close to the scatterchron. It has lower Sr (8 ppm), MnO, MgO, and CaO than the other granite samples (Table 4), and the $^{87}Sr/^{86}Sr$ is relatively higher. These data suggest that 10-

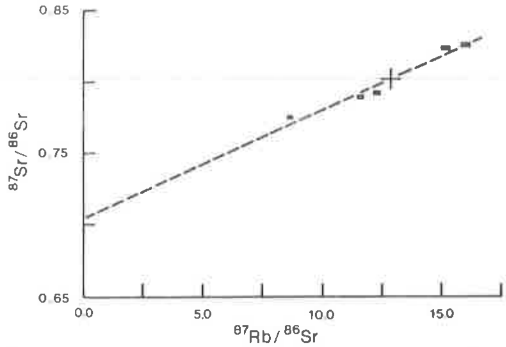

Figure 6. Rb/Sr scatterchron plot for whole-rock samples of Fountain granite. Cross is centroid (12.89, 0.8020). Slope is 0.007571 ($\pm$0.000859) and age is 531 $\pm$ 60 my. Initial ratio is 0.70447 $\pm$ 0.01129. Least squares line is after York (1969). Sizes of black rectangles give estimates of experimental errors associated with the points.

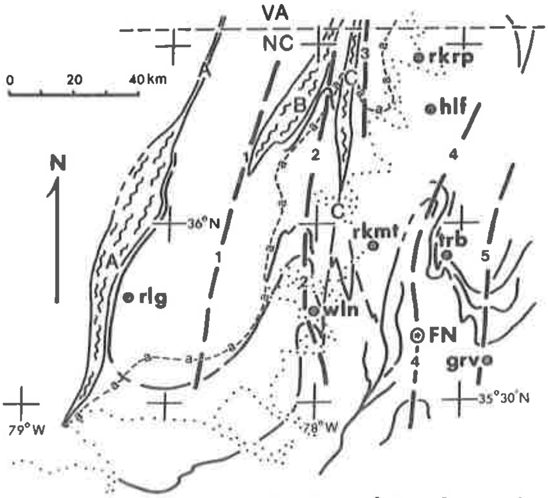

Figure 7. Geologic map showing the relationship of the Fountain granite to structural and metamorphic elements in the eastern Piedmont. Towns include Fountain (fn), Raleigh (rlg), Wilson (wln), Rocky Mount (rkmt), Halifax (hlf), Roanoke Rapids (rkpd), Tarboro (trb), and Greenville (grv). The western edge of the coastal plain cover is shown by the dotted line. Fault zones and mylonite (wavy-line pattern) are as follows (Farrar, in press): Nutbush Creek zone (A); Macon zone (B); and Hollister fault and mylonite zone (C). Fold axes (heavy lines with numerals) are as follows: Wake-Warren antiform (1); Spring Hope synform (2); Thelma antiform (3); Fountain synform (4); and Greenville antiform (5). The line --a-- shows the boundary between the chlorite-biotite zone (to the east and south in the eastern Carolina slate belt) and higher grade metamorphic rocks (staurolite-sillimanite, staurolite-kyanite, and biotite-almandine) in the Raleigh Block between the Nutbush Creek zone and line --a--. The Nutbush Creek zone marks the eastern boundary of the main Carolina volcanic slate belt (Triassic rocks are not shown). The structural form lines show the traces of linear magnetic anomalies (Farrar, in press; Zietz and others, 1980).

7-4 preferentially lost alkaline earth elements (including Sr) during recrystallization, and that its Rb/Sr ratio may be three to four times the pre-crystallization value.

AGE OF METAMORPHISM AND DEFORMATION

The K-Ar age on microcline (257 my) gives the date when the quarry rocks cooled below 200°C for the last time, so it is interpreted as an uplift-cooling age. The dated biotite sample (Table 9) is from a strongly foliated and banded marble-pelitic schist which may have been a thin pre-metamorphic lamprophyre (?) dike (Mauger, in preparation). Although we are not sure how those pelitic rocks originated, they show the same foliation as the granite and the amphibolites, so they record the major deformational-metamorphic event. The biotite age (309 my; a few million years less if excess argon is present, Table 11) also represents a time of cooling and uplift following prolonged (?) burial at temperatures above 250°C. Thus the major period of metamorphism and deformation occurred prior to 300 my ago, and it is older than Alleghanian deformation that affected the Butterwood Creek granite in the area of the Hollister zone and Thelma antiform (Fig. 7) (Farrar, in press; Farrar and others, 1981).

Isotopically, the main period of metamorphism and deformation is constrained to be within the broad limits imposed by the actual K-Ar ages of hornblende and biotite. Our preferred interpretation is that excess environmental ^{40}Ar in the hornblende would bring its age closer to that of the biotite. Thus the major metamorphic and deformational event was probably Hercynian or Acadian in age and the hornblende age does not necessarily record an Ordovician metamorphic event. Extrapolating $^{87}Sr/^{86}Sr$ ratios back in time (Table 12) indicates that sample 10-7-4 and average Fountain granite would have had the same $^{87}Sr/^{86}Sr$ ratio between 335 and 340 my ago. This

Table 11. Effect of excess ^{40}Ar on K-Ar ages of hornblende and biotite from the Fountain Quarry.

biotite (UAKA-81-97)			hornblende (UAKA-79-157)		
δ	%	Age, my	δ	%	Age, my
0	0.0	309	0	0.0	474
20	0.4	308	20	9.0	437
40	0.9	307	40	17.9	398
60	1.3	306	60	26.9	359
80	1.7	304	80	35.9	318
100	2.2	303	100	44.8	277
*92	2.0	304	*92	41.2	294

δ is excess $^{40}Ar \times 10^{-12}$ m/g

% is percentage of excess ^{40}Ar in total radiogenic ^{40}Ar

* is 92×10^{-12} m/g excess ^{40}Ar estimated for albite from the Fountain granite

Table 12. Estimated $^{87}Sr/^{86}Sr$ ratios of sample 10-7-4 and average Fountain granite as a function of age.

time	87Sr/86Sr	
my ago	10-7-4	*ave. granite
0	0.9700	0.7968
330	0.7441	0.7397
335	0.7406	0.7389
340	0.7372	0.7380
345	0.7338	0.7371
350	0.7303	0.7363

$*^{87}Rb/^{86}Sr$ is 12.15

should represent the time when 10-7-4 recrystallized and lost strontium. This result is compatible with our preferred interpretation of the K-Ar results, given the uncertainties involved in both estimates.

REGIONAL GEOLOGIC SETTING OF THE FOUNTAIN GRANITE

The Carolina and eastern Carolina volcanic slate belts (CVSB and ECVSB) are separated by higher grade metamorphic rocks and late Paleozoic granites of the Raleigh Belt (Fig. 7). South of Raleigh, the higher grade metamorphic rocks die out rather abruptly, and greenschist-grade rocks of the two volcanic slate belts are separated only by a 10 to 20 km-wide strip of Triassic rocks in the Wadesboro-Deep River-Durham Basin. Lithologic and magnetic patterns in the southeastern and eastern Raleigh Belt are continuous into the ECVSB, showing that the contact between the belts represents a very steep metamorphic gradient, not a major fault zone (Farrar, in press). Farther east, lithologic and magnetic patterns show marked discontinuities along the southward projection of the Hollister fault and mylonite zone (Fig. 7). Farrar used this fault zone to divide the ECVSB into zones I (on the west) and II (on the east). Zone II is buried by Coastal Plain sediments except for small areas of metavolcanic rocks and late Paleozoic granites that are exposed in the major river valleys. ECVSB II probably extends to the area just east of Greenville, where changes in magnetic patterns indicate the presence of a buried Triassic basin (Daniels and Zietz, 1978). Thus the Fountain outcrop is probably in ECVSB II.

Fossil and isotopic evidence indicates that the volcanic rocks of the CVSB in south central North Carolina are Cambrian. An older late Precambrian volcanic assemblage is recognized from Chapel Hill northward into southern Virginia. The rocks in the younger southern portion are bimodal in silica, with siliceous varieties being dominant over mafic ones (Seiders, 1978). In the ECVSB, the volcanic and volcaniclastic rocks are inferred to be late Precambrian to early Paleozoic (Farrar, in press). The probable Cambrian age for the Fountain granite supports this age assignment. Although the sampling is indeed meager, the amphibolite (basalt) and metarhyolite dikes at Fountain suggest that the volcanic assemblage was also bimodal in silica. Cuttings from wells farther east along the coast give ages that range from latest Precambrian to mid-Permian (Denison and others, 1967). However interpretations of these results are complicated by small sample sizes and by a lack of detailed knowledge of pertinent geologic factors concerning the samples (Denison and others, 1967). Their data support the idea that other distinctive lithologic-tectonic belts occur east of ECVSB II under the Coastal Plain and continental shelf (Glover, 1980).

Structural grain in the ECVSB can readily be interpreted from magnetic patterns (Daniels and Zietz, 1978; Zietz and others, 1980; Farrar, 1980; and Farrar, in press). Three episodes of folding are recognized (Farrar, in press). The youngest folds trend northeast-southwest; axial planes dip steeply, and locally the folds are overturned. This complex polyphase folding history contrasts sharply with the relatively simple, open fold patterns in the Cambrian part of the CVSB farther west.

Magnetic patterns due to tight folding and multiple fold episodes can also be recognized in ECVSB II (Fig. 7). The Greenville structure (axis 5) opens to the north; thus it is identified as an antiform by analogy with known structures in ECVSB I (Farrar, in press). Our interpretation suggests that a very tight synform (axis 4, Fig. 7) is located between Tarboro and Halifax (aeromagnetic map, Zietz and others, 1980). Southwest of Fountain, the west limb of this structure exhibits a complex dendritic pattern of linear high-intensity anomalies which may reflect relatively thin mafic flows and/or dikes in a pile of felsic volcanic or volcaniclastic rocks. This inferred synform passes through Fountain, and the outcrop form (Fig. 2), the foliation in the granite, and the magnetic signature of the granite (Zietz and others, 1980) conform with its northeast trend.

The youngest fold axes in ECVSB I (F_3 of Farrar, in press) were identified as Alleghanian on the basis of studies in and near to the Raleigh Belt (Farrar and others, 1981). In view of the K-Ar date on biotite, the NNE-trending regional fold pattern at Fountain probably had developed prior to 300 my ago.

Granitic plutons intrude the ECVSB (Speer and others, 1980). These give whole-rock Rb/Sr ages from 313 my (Castalia pluton, Fullagar and Butler, 1979) to 285 my for the Sims granite 40 km west of Fountain (Wedemeyer and Spruill, 1980). The Sims granite is not deformed; other 300 my-old plutons in the Carolina slate belt and in the eastern part of the Charlotte belt are generally regarded as undeformed (Fullagar, 1971; Butler and Ragland, 1969). However, the Rolesville pluton is foliated (Farrar, in press) and specific deformational features such as bulbous myrmekite, mortar texture, and mylonitic textures have been identified in 300 my-old granites near Lake Gaston (Koehler, 1982; Grundy, 1982; Farrar and others, 1981) slightly west and north of Roanoke Rapids, North Carolina (Fig. 7). In the east-central Virginia piedmont, late Paleozoic granites thought to be equivalent to the Petersburg granite (330 my; Wright and others, 1975) are cut by late Pennsylvanian-early Permian (Alleghanian) mylonitic zones (Bobyarchick and Glover, 1979). Associated amphibolites and other rocks reached peak amphibolite-grade metamorphic temperatures and were subjected to strong compressive deformation about 340 my ago. Slow cooling through greenschist-grade temperatures from 340 to 300 my ago was accompanied by retrograde metamorphism and by one minor episode of compression (Bobyarchick and Glover, 1979). A similar metamorphic history would be compatible with the isotopic data and with the metamorphic history of the rocks from Fountain.

In the Kiokee belt of South Carolina, Snoke and others (1980) identified late Paleozoic plutonic rocks and older rocks that were metamorphosed and strongly deformed in a late Paleozoic (Alleghanian) event. The Kiokee and Raleigh belts are similar in that late Paleozoic deformation of CVSB rocks is recognized in both areas (Hatcher and others, 1980). The Augusta fault zone, the eastern boundary of the Kiokee belt, is shown as being continuous with the Hollister zone (Hatcher and others, 1980). The Belair zone east of the Augusta line contains a similar lithologic assemblage to CVSB. It is thus equivalent to ECVSB II of Farrar (in press). The Fountain outcrop and the Belair belt were evidently only slightly affected by the intense Alleghanian metamorphism and deformation in neighboring parts of the Raleigh-Kiokee belt. Perhaps the zones of intense Alleghanian deformation and metamorphism are restricted to areas where very large late Paleozoic granitic plutons intruded the slate belt. These areas may have still been hot when the Alleghanian compressive crunch was initiated.

EMPLACEMENT OF THE FOUNTAIN GRANITE - THE TECTONIC IMPLICATIONS

The magnetic patterns in ECVSB I and in the area between Wilson and Greenville have about the same proportion of high- and low-intensity zones as are shown in the CVSB in south-central North Carolina, suggesting that the proportions of felsic to mafic rock are roughly equivalent in the two areas. Since this part of the CVSB is about 80 percent felsic (Seiders, 1978), the ECVSB is also probably dominantly felsic. The Fountain peralkalic occurrence is similar to those which are associated with large-scale, dominantly rhyolitic volcanism in rifted areas with thinned continental crust such as in the western United States. These rhyolites including the minor peralkalic varieties, are inferred to have been associated with deep-seated mafic magmatism (Hildreth, 1981). Gravity data suggest that the CVSB has a "mafic keel" at depth (Hatcher and Zietz, 1980; Seiders, 1978); however patterns in the ECVSB are complicated by the negative anomalies due to the late Paleozoic granites. A pattern of positive anomalies extending from Virginia southward toward Greenville is very similar to patterns in the CVSB, suggesting that the ECVSB also has mafic rocks at depth which have locally been displaced by younger granites. The Halifax County Mafic Complex (Stoddard and Teseneer, 1978; Farrar, in press; his Fig. 2) may represent a sliver of this mafic material.

Late Precambrian plutonic and volcanic rocks with peralkalic affinities occur in the Crossnore, North Carolina and Mount Rogers, Virginia area in the Blue Ridge Belt (Rankin, 1975; Novak and Rankin, 1980). Concordia intercept Pb-Pb zircon ages of the Crossnore-type granites are substantially older (about 800 my; Rankin and others, 1969) than the probable age of the Fountain granite. Rb/Sr whole rock ages from the same rocks are somewhat younger (700 to 500 my; Odom and Fullagar, 1971), and the youngest Rb/Sr age (500 my) is not significantly different from the Fountain result. Crossnore-type granites from the Sauratown Mountains (Espenshade and others, 1975) give generally discordant Rb/Sr whole-rock ages of from 700 to 600 my (Fullagar and Butler, 1980). Other possible Crossnore-type granites from near the Blue Ridge-Inner Piedmont boundary in Wilkes and Surry Counties, North Carolina, give discordant ages of from 700 to 300 my (Harper, 1977). In both areas, the discordant results can be attributed to changes in whole-rock Rb/Sr ratios resulting from metamorphism. None of Harper's (1977) possible Crossnore granites have $(Na+K) \geq Al$. However, some have high total alkalies, high Zr, and high Y; these are typical chemical features of silicic peralkalic rocks.

The close association of extensional tectonics and peralkalic volcanism (Macdonald, 1974b; Barberi and others, 1974), in part led Rankin (1975 and 1976) to propose that the Crossnore-Mount Rogers igneous assemblage was associated with a late Precambrian to early Ordovician rift zone in the Blue Ridge area, which at that time was situated at the edge of the North American craton. He regarded the Piedmont province as having been welded to North America at a later time. Recent COCORP seismic results support the concept that southern Appalachian piedmont terranes are part of a crystalline-rock thrust sheet that was emplaced over the late Precambrian-early Paleozoic continental margin of North America (Cook and others, 1979; 1980; 1981).

The peralkalic rocks at Fountain support a rift-related origin for the eastern volcanic slate belt, although regionally, geochemical data have been interpreted to favor an island-arc origin for parts of the main CVSB (Whitney and others, 1978; Black, 1980; Bland and Blackburn, 1980). Geophysical data have been interpreted to support rifting (Long, 1979).

The Crossnore-Mount Rogers area may have been near the western edge of a zone that was wider and more diffuse than that postulated by Rankin (1975). Theoretical studies (Le Pichon and Sibuet, 1981) indicate that crustal extension in zones 150 to 350 km wide may be enough to initiate diapiric intrusion of asthenosphere, which is followed by partial melting, volcanism, and emplacement of new oceanic crust at specific sites in the rifted area. Rifting of an area with continental crust would have produced "transitional crust" (Buffler and others, 1980; Anonymous, 1981) in which rift basins with thin continental crust (or none at all) are bordered by horst blocks of thicker crust. Emplacement of sub-crustal mafic magmas could have provided the heat to generate extensive rhyolitic volcanism at the surface. The voluminous rhyolites would have originated partly through partial melting of the more sialic crustal rocks and partly by magmatic differentiation. Small amounts of peralkalic rhyolite magma, represented by the Fountain granite, evolved toward the top of a deep, vertically-continuous magma system (Hildreth, 1981), which was connected with the sub-crustal mafic magmas. These deep-seated mafic magmas then crystallized to give the Carolina volcanic slate belt its characteristic mafic geophysical signature.

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PETROGRAPHY OF SOME ORTHOQUARTZITES FROM THE KEEFER FORMATION (SILURIAN), MONTGOMERY AND GILES COUNTIES, SOUTHWESTERN VIRGINIA

JON LIGHTNER Department of Geology, Radford University, Radford, Virginia 24142

ROBERT C. WHISONANT Department of Geology, Radford University, Radford, Virginia 24142

ABSTRACT

The Keefer Sandstone is a middle Silurian clastic formation prominently exposed in the Valley and Ridge Province of the central and southern Appalachians where it consists primarily of white orthoquartzites. Petrographic analysis of some Keefer orthoquartzites from the Montgomery and Giles Counties area of southwestern Virginia reveals a high degree of textural and mineralogical maturity. Our data concerning Keefer sandstone matrix versus framework percentages, framework sorting and rounding, quartz types and abundances, and accessory mineral composition support previous suggestions that these arenites formed in shallow shelf to intertidal settings. The Keefer source area most likely was a predominantly sedimentary terrane located beyond the present southeastern edge of the basin.

INTRODUCTION

The Keefer Sandstone (Middle Silurian) crops out in the Valley and Ridge Province of southwestern Virginia as a topographically resistant, ridge-forming sequence composed principally of hard, white to gray, fine- to medium-grained quartz-cemented orthoquartzites. Other lithologies, chiefly shale interbeds, are present in varying proportions. The Keefer Sandstone is underlain by the Rose Hill Formation which in turn is underlain by the Tuscarora Sandstone. The Rose Hill consists predominantly of reddish-colored shales and hematitic siltstones and sandstones with some beds very characteristically burrow-mottled. The Tuscarora is primarily orthoquartzite which, except for a greater abundance of conglomeratic zones, is virtually identical to the Keefer in the field. Field discrimination between the Keefer and Tuscarora sandstones, therefore, can be exceedingly difficult unless relatively complete exposures of the two formations and the intervening diagnostic Rose Hill lithologies are present.

The Keefer has been studied extensively in portions of Pennsylvania, Maryland, West Virginia, and Virginia (Butts, 1940; Woodward, 1941; Folk, 1960; Dennison, 1970; Woodfork and Patchen, 1970; Diecchio, 1973; Lampiris, 1975) but has received little detailed petrographic analysis in the southwestern Virginia area, where it is much thicker and more mature than to the north. Folk (1960) indicated that the Keefer in southwestern Virginia may have had a different source area and been deposited separately from the Keefer to the north. Diecchio (1973) and Lampiris (1975) thought the Keefer of southwestern Virginia should be given a separate formational name from the Keefer of Pennsylvania, Maryland, and West Virginia. Lampiris (1975) proposed the name "Eagle Rock Sandstone" after the location of the Keefer's thickest exposure. Inasmuch as very little information concerning the petrography of the Keefer Sandstone in southwestern Virginia is available, we undertook this investigation to obtain new petrographic data and to use these data to assess ideas concerning Keefer depositional environments and provenance area. Additionally, we intend our Keefer data to provide a further basis of comparison with the excellent detailed petrographic analyses of the Tuscarora available from Folk (1960) and Cotter (1982) to the north and Hayes (1974) in southwestern Virginia for purposes of discriminating between the two formations.

METHODS

This study is aimed principally toward a petrographic description of the white orthoquartzite lithologies found in the Keefer within a relatively small area of southwestern Virginia. (We consider the term "orthoquartzite" as equivalent to quartz arenite or quartzarenite, following the usage of Pettijohn, 1975, and Blatt, Middleton, and Murray, 1980). Petrographic analysis was restricted to the orthoquartzites because, first, this is by far the lithology most typical of the formation. Second, othoquartzite is also the most common lithology encountered in the Tuscarora in this area (Hayes, 1974), thus petrographic comparison of the two formations should be facilitated. No attempt was made to include any shales or other non-sand-sized clastic lithologies. We deliberately omitted from analysis rock types atypical of the Keefer orthoquartzite lithology such as the approximately 20 meters of pyrite-rich conglomerate and sandstone present at the base of the Keefer in the Gap Mountain section in Montgomery County, Virginia (Spencer, 1970).

In order to accomplish our goals, three relatively complete and continuously exposed Tuscarora-Rose Hill-Keefer sections in southwestern Virginia were chosen to insure sample selection over the Keefer section (Fig. 1). These exposures are (1) the road cut along U.S. 460 through Gap Mountain (sample designation GM) in Montgomery County, (2) the railroad cut also through Gap Mountain along the east side of the New River near McCoy (sample designation Mc) in Montgomery County, and (3) the road cut along U.S. 460 along the east side of the New River at the Narrows (sample designation N) in Giles County. A total of eighteen samples representative of the white orthoquartzite lithology were taken at irregular intervals throughout the Keefer succession.

One thin section was made from each of the samples and analyzed using standard petrographic techniques. Quartz was classified as vein quartz, pressure quartz, schistose quartz, or common quartz for comparison with Hayes' (1974) description of Tuscarora quartz types in the southwestern Virginia area. Textural calculations for mean grain size (ϕ) and sorting (ϕ) were performed using the formulas suggested by Folk (1968).

RESULTS

General Observations

Colors present in the Keefer samples range from white (N9) to grayish-red (5R 4/2) but most samples have an overall white, light gray (N7), or green-gray (5GY 6/1) color. Several samples displayed a distinct mottled pattern due to changes in color and

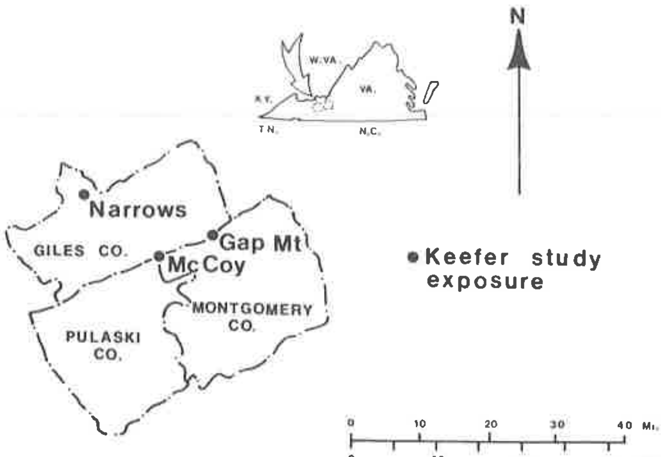


Figure 1. Index map of southwestern Virginia.

texture along irregular surfaces roughly parallel to bedding planes. Most samples showed a yellow to purple staining on the weathered surfaces from iron oxidation.

Only one sample contained any trace fossils. These are burrows, perpendicular to bedding, approximately one mm in diameter and 10 to 20 mm long. These burrows are encircled by green reduction halos, one to four mm wide. Such structures are of the *Skolithos* ichnofacies and have been described in the Tuscarora by Diecchio (1973) and Hayes (1974) and in the Keefer by Diecchio (1973). Numerous burrow structures, both horizontally and vertically directed, were observed in the Keefer in the field.

Texturally, the Keefer samples ranged from very fine sandstones with significant amounts of silt and clay to medium to coarse sandstones with very little silt and clay. Nearly all samples were well-cemented with quartz, but a few were slightly friable. Most samples contained at least a small amount of matrix material. The only framework mineral present in amounts greater than two percent on any slide was quartz. Indeed, on most slides quartz was the only framework mineral encountered during point counts, but other frameworks were observed and identified.

Matrix and Cement

Matrix in the Keefer ranged from none to 28 percent (hence, some of the samples qualify as quartz wackes) with a mean amount of 10 percent (Table 1). The matrix was commonly a hash of silt- and clay-sized particles, clay minerals, and sericite. A small amount of authigenic muscovite was present on many slides. Authigenic biotite was seen on two slides and authigenic chlorite on one. On about half the slides the matrix was colored yellow to red in ordinary light from abundant oxidized iron. Small hematite flakes were seen in the matrix on a few slides. One slide showed a significant amount of chert matrix, amounting to about one-third of the total matrix. Chert and clay matrix were concentrated in different areas of this slide. Hayes (1974) reported Tuscarora matrix compositions of varying proportions of clay, muscovite, sericite, chlorite, and biotite which are essentially identical to those observed in the Keefer by us.

We observed two Keefer samples that contained no matrix. In these instances, quartz overgrowths filled all available pore space. Overgrowths were seen on every slide but one. In several samples oxide coatings clearly outlined the detrital quartz grains, but even on slides where outlines were not seen, overgrowths were evident from the irregular shapes of the authigenic quartz masses extending into the interstices of adjacent grains. Hayes (1974) reported overgrowths as abundant in the Tuscarora. Oxide coatings are not characteristically preserved on detrital grains in the Tuscarora, however, apparently because the grains were subjected to a high energy environment which removed oxide coatings (Hayes, 1974).

Table 1. Texture of some Keefer orthoquartzites.

Sample	Meters Above Base	Matrix(%)	Mean Grain Size (ϕ)	Sorting (ϕ)
GM-1	21	12	1.03	0.50
GM-2	22.5	0	1.90	0.82
GM-3	33	0	1.90	0.86
GM-4	38	28	1.90	0.87
GM-5	38.5	23	3.07	1.12
GM-6	45	15	1.82	1.00
GM-7	47	2.5	2.48	0.63
Mc-2	1	26	2.40	0.97
Mc-3	2.5	11	1.18	0.59
Mc-4	13	14	0.97	0.58
Mc-5	25	5	1.74	1.00
Mc-6	30	6	1.79	0.96
Mc-7	31	3	1.15	0.74
N-2	1	9	0.53	1.13
N-3	6	14	1.95	0.64
N-4	9	3	1.56	0.65
N-5	11	14	2.56	0.75
N-6	16	2	0.16	0.70
		$\bar{x}$=10.4	$\bar{x}$=1.67	$\bar{x}$=0.81

Framework Components: Texture

Textural data concerning the Keefer framework components are given in Table 1 and in the Appendix which contains histograms showing the grain size distribution for each sample, composite distributions for each exposure, and the composite distribution for all Keefer samples analyzed during this study. Mean grain size of the framework components in the Keefer samples ranged from 3.07ϕ to 0.16ϕ with an average mean grain size of 1.67ϕ. Three of the samples were bimodal.

Sorting ranged from 0.50ϕ standard deviation to 1.13ϕ standard deviation with a mean of 0.81ϕ. Sorting derived from the composite grain size distribution was 1.02ϕ. Roundness estimates are difficult to obtain accurately in pressure-welded orthoquartzites such as the Keefer because of extensive overgrowths. Detrital grains in the Keefer are subrounded to well-rounded based on those samples where the grain outlines could be clearly seen. Some samples showed two distinct populations of rounded and angular to subangular grains.

Hayes (1974) recorded a range in mean grain sizes in the Tuscarora of 0.10ϕ to 3.29ϕ with an average of 1.90ϕ. His sorting values ranged from 0.30ϕ to 1.47ϕ with an average of 0.67ϕ. This suggests that the average Tuscarora sample is slightly finer grained and better sorted than our average Keefer sample. Hayes (1974) found the Tuscarora detrital grains to be typically subrounded to well-rounded, matching our Keefer observations.

Framework Components: Composition

Quartz is by far the preponderant detrital framework constituent in the Keefer samples, forming approximately 99 percent of the composition in this group. Accessory grains of varying types constitute the remaining one percent. Table 2 presents the different forms of quartz and species of accessory minerals observed in the Keefer during this investigation. Approximately 98 percent of the quartz grains were classified as common (following Hayes' [1974] usage of "common" as a catch-all term for quartz types that cannot be positively identified as belonging to any other group). In order of decreasing abundance, the quartz types in the Keefer samples are common, pressure, vein, and schistose. We discerned no consistent vertical trends in variation of quartz types.

Very little polycrystalline quartz seems to be present in the Keefer, although the tight suturing and interlocking of quartz grains make it very difficult to distinguish polycrystalline quartz reliably. Polycrystalline quartz content ranged from none to approximately five percent, supporting the observation of Blatt, Middleton, and Murray (1980, p. 291) that pure quartz sandstones tend to contain very minor amounts of polycrystalline quartz. From 61 to 93 percent of the 1ϕ to 2ϕ quartz grains in the Keefer were undulose with a mean of 80 percent. The undulose quartz might reflect original source rocks of low rank metamorphic grade (Basu and others, 1975) whereas the abundance of common quartz and the minor amount of polycrystalline grains could reflect high rank metamorphic or plutonic parent rocks (Blatt, Middleton, and Murray, 1980). As indicated in the following section, we believe that the physical and mineralogical maturity of the Keefer samples indicates reworking of previous sedimentary rocks, hence the derivation of detritus from freshly exposed crystalline parent rocks must have occurred several cycles prior to Keefer sedimentation.

Accessory minerals are found in nearly all of the Keefer orthoquartzites but constitute a very small percentage of the total framework components. Types of accessory minerals are shown in Table 2. The total abundance ranged from none to two percent with a mean of one percent. Tourmaline was the most commonly observed accessory mineral. It was seen on every slide except one, in amounts from one grain on a slide to an estimated two percent of the framework fraction. Both green and brown varieties were seen. Zircon, ilmenite, and rutile in that order were the next most commonly seen accessories, also in amounts of a few grains to two percent. One or two grains of finely foliated metamorphic rock fragemnts were observed on seven slides. Other accessory minerals seen on at least one slide were magnetite, chert, hornblende, hematite, garnet, and epidote. A very small amount of muscovite on some slides was possibly detrital.

Table 2. *Composition of some Keefer orthoquartzites.*

Sample	Meters Above Base	% of Total Quartz: Common	Vein	Pressure	Schistose	Accessory Minerals*
GM-1	21	99	1	0	0	T
GM-2	22.5	100	0	0	0	T, Z, I, Hb
GM-3	33	100	0	0	0	T, Z, Hb, Ep
GM-4	38	100	0	0	0	None
GM-5	38.5	98	1	1	0	T, I, Ma, Mu
GM-6	45	95	0	5	0	T, Z, R, Hb
GM-7	47	100	0	0	0	T, I, Ch
Mc-2	1	96	0	4	0	T, Z, I, R, Ma, He, MRF
Mc-3	2.5	99	0	0	1	T, Z, R, MRF
Mc-4	13	100	0	0	0	T, G, MRF
Mc-5	25	100	0	0	0	T, Z, I, R, G, MRF
Mc-6	30	100	0	0	0	T, Z, I, R, MRF
Mc-7	31	100	0	0	0	T, Z, R, Ch, MRF
N-2	1	97	3	0	0	T, Z, I, R
N-3	6	95	2	2	1	T, I, R, MRF
N-4	9	96	0	3	1	T, Z, I
N-5	11	96	0	4	0	T, Z, I, Hb, Ch
N-6	16	92	4	3	1	T, Z
		$\bar{x}$=97.9	$\bar{x}$=0.6	$\bar{x}$=1.2	$\bar{x}$=0.2	

*Accessory Minerals: T = tourmaline, Z = zircon, I = ilmenite, R = rutile, Hb = hornblende, Ch = chert, Ma = magnetite, Mu = muscovite, G = garnet, Ep = epidote, He = hematite, MRF = metamorphic rock fragment

Comparing our Keefer framework compositional data to that reported by Hayes (1974) for the Tuscarora in southwestern Virginia, we find virtually identical amounts of quartz and accessory grains (99 and one percent, respectively). Hayes (1974) found common quartz to be the predominant variety, followed by pressure, vein, and schistose (matching our ranking of quartz types in the Keefer). He reported a lower percentage of common quartz (approximately 87 percent of the total quartz population) than we found in the Keefer (about 98 percent). He also observed vertical changes in the percentage of quartz types whereas we saw none. Finally, Hayes (1974) recorded the following non-quartz framework components in his Tuscarora samples: a very few feldspar grains, some metamorphic rock fragments, and a heavy mineral assemblage consisting of tourmaline, zircon, ilmenite, and one grain of sillimanite. We can detect no important differences in the composition of the accessory fractions of the two formations.

SOURCE AREA AND DEPOSITIONAL ENVIRONMENTS OF THE KEEFER ORTHOQUARTZITES IN SOUTHWESTERN VIRGINIA

The Keefer orthoquartzites in the study area are characterized by well-developed compositional and physical maturity. This most probably reflects a polycyclic history for the Keefer detritus. Suttner, Basu, and Mack (1981, p. 1244) contend that virtually all ancient quartz arenites (orthoquartzites) are multicyclic in origin. Dickinson and Suczek (1979) analyzed modern and ancient sands and related their compositions to plate tectonics settings. They found quartz arenites to be derived from both cratonic and recycled orogenic provenances. Inasmuch as the Keefer clastic material is found in the latter setting, the conclusion that pre-Keefer arenaceous sediment must have been a major component of the source area seems very likely.

Investigations involving the Lower and Middle Silurian clastic sequence (Tuscarora-Rose Hill-Keefer) in southwestern Virginia invariably postulate a source area located toward the southeast (present coordinates) of the outcrop area (Dennison, 1970; Diecchio, 1973; Hayes, 1974; Lampiris, 1975; Whisonant, 1977). Chen (1977, p. 87) also stated that the land providing sediments for the Keefer lay on the present southeast side of the basin. Folk (1960) considered the northern Keefer to have the same southeastward-located source area as the Tuscarora. Mineralogically, the Keefer and Tuscarora orthoquartzites in southwestern Virginia are almost identical, further indicating a common source area that Tuscarora paleocurrent analysis in this area (Hayes, 1974; Whisonant, 1977) placed toward the southeast. Indeed, Diecchio (1973) has proposed that the Keefer in southwestern Virginia is actually a time-transgressive upward extension of the Tuscarora and that deposition was continuous to the southeast of the present outcrop belts.

Recyling of older sedimentary material from the provenance area and extensive exposure to high energy conditions during transportation and deposition in the Keefer basin is evidenced by the high quartz content and the rounding of the quartz grains observed during this study. Oxide coatings were seen on some quartz grains in samples which were quite texturally mature. This might indicate acquisition of an oxide coating during a phase of subaerial exposure (as, for example, along beach ridges) prior to final burial. A second possibility is that, although waves and currents were evidently strong enough to continue rounding quartz grains and removing labile grains surviving from previous cycles, physical energy may not have always been high enough to remove inherited oxide coatings. The rounding, extreme chemical stability, and minor amount of the accessory mineral fraction further support reworking of older sediment to produce the Keefer detritus.

The Keefer orthoquartzites show a greater range in textural maturity than mineralogical maturity, as indicated by the widely varying matrix percentages. Matrix content does not appear to correlate with roundness or sorting. Some samples with a mature framework composition show 10 to 20 percent matrix. Samples with higher matrix contents probably represent quieter water environments in which sandy detritus previously matured was deposited.

Several of the hand samples show changes in color or texture along irregular surfaces roughly parallel with the bedding plane. The slides made from these samples, as well as some others, show selective concentrations of larger and smaller grains which may be a very small scale graded bedding. On many slides there is an apparent parallel alignment of larger grains. This is probably a pressure solution effect in some cases, but on one slide the direction of alignment is the same as the long axis of the zone in which the larger grains are concentrated. These features and the mixture of rounded and subangular grains on some slides might reflect the mingling of two or more separate sediment populations during deposition.

Previous workers have postulated a variety of depositional settings for the Keefer Formation, ranging from nearshore shelf and intertidal environments through beach and barrier bars to coastal plain fluvial systems (Folk, 1960; Dennison, 1970; Diecchio, 1973; Lampiris, 1974, 1975; Dennison and Head, 1975; Chen, 1977; Smosna and Patchen, 1978). Our work on the Keefer orthoquartzites clearly supports a depositional regime in which the already matured Keefer detritus was driven toward a higher degree of mineralogical and textural maturity. We envision as the principal environments a wide shallow shelf and intertidal setting to achieve this maturity. The presence of *Skolithos* indicates an intertidal to littoral environment (Seilacher, 1967; Diecchio, 1973). Certainly a more comprehensive analysis of the entire Keefer interval throughout southwestern Virginia involving more detailed petrography and analysis of sedimentary structures will be necessary to identify the precise depositional conditions. Cotter's (1982) recent study of the Tuscarora in Pennsylvania using just such an approach identified a complex of depositional environments in that unit ranging from braided fluvial systems to shelf sand waves. We anticipate a similar complexity of depositional settings for the Keefer in southwestern Virginia.

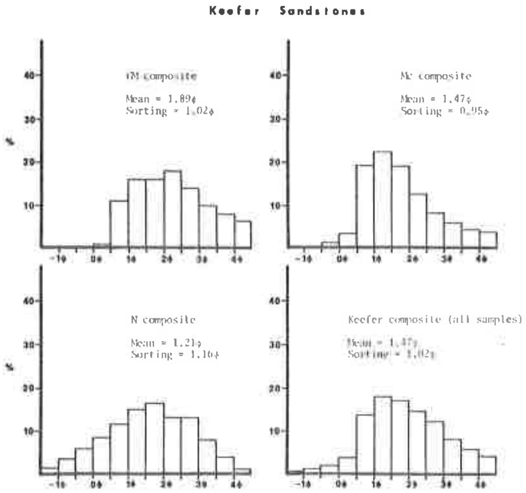
CONCLUSIONS

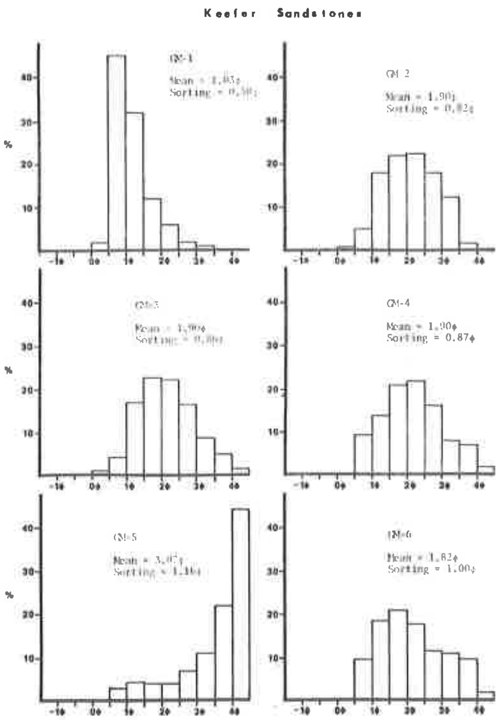
We conclude that the high degree of textural and mineralogical maturity displayed by the Keefer orthoquartzites in the study area supports an origin involving reworking of older sedimentary detritus in shallow shelf and intertidal settings, although other possible depositional environments are not ruled out. (The nature of the maturation processes affecting clastic material in pre-middle Paleozoic non-marine settings, particularly fluvial systems, when extensive terrestrial vegetation did not exist remains an unresolved problem.) Comparison of our Keefer petrographic data to numerous Tuscarora petrographic analyses indicates a common source land, well established by Tuscarora provenance studies (as well as by a few Keefer investigations) to be most likely located beyond the present southeast margin of the basin.

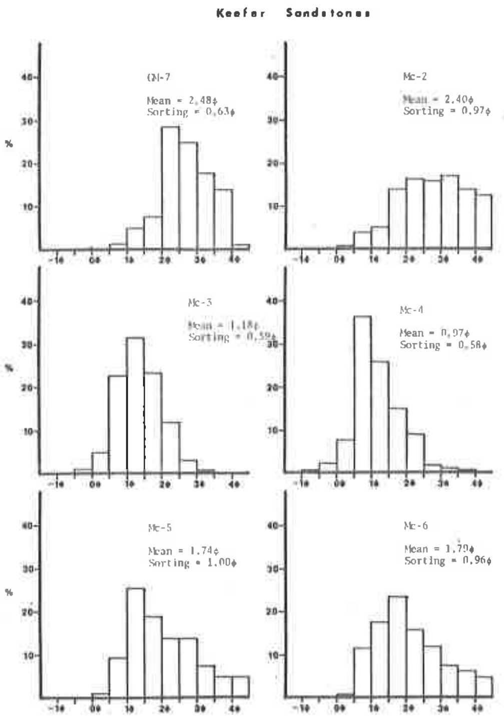
We wish to emphasize that our relatively limited study has presented more questions than conclusions. The full Keefer interval should be sampled over the entire area of exposure in southwestern Virginia and southern West Virginia. Samples should be taken at close-spaced, regular stratigraphic intervals and each section should be measured and described in detail. Sedimentary structures should be studied and facies mapped. A paleocurrent analysis especially would shed more light on the nature of the source area as well as the transport systems affecting the Keefer detritus.

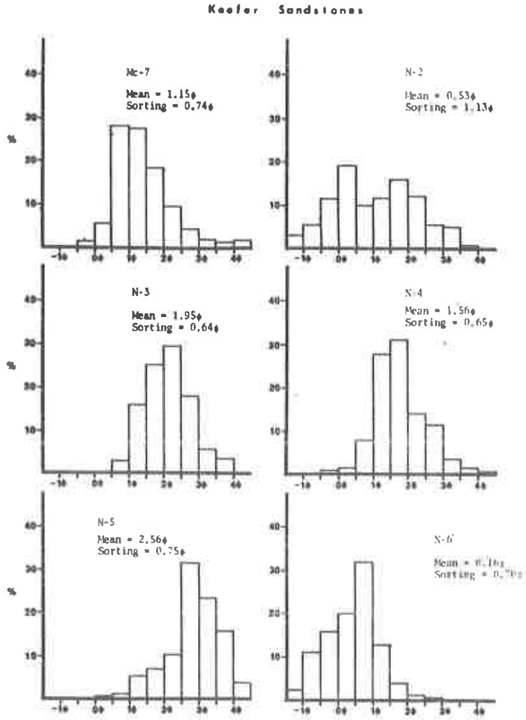
APPENDIX

Textural Data - Keefer Orthoquartzites (GM = Gap Mountain, Mc = McCoy, N = Narrows).









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Patchin Crandall Curtis