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SOUTHEASTERN GEOLOGY

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MAGADI-TYPE CHERT, INDICATOR OF A LACUSTRINE ENVIRONMENT IN THE MIDDLE EOCENE MCBEAN FORMATION, SOUTH CAROLINA

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ABSTRACT

Magadi-type chert occurs in the basal part of the middle Eocene McBean Formation at three localities in Orangeburg and Calhoun Counties, S.C. These occurrences imply that in this area the basal part of the McBean Formation was deposited in an alkaline lake with a high silica content. Formation and filling of the lacustrine depression could have resulted from a timely combination of the following factors: an initial nearshore to estuarine depositional environment; falling sea level; minor tectonic uplift of the Cape Fear arch and (or) the northern boundary of the South Georgia rift basin; intense chemical weathering in a tropical climate; and seasonal rainfall.

INTRODUCTION

Magadi-type chert is a variety of chalcedony that has been found in sedimentary deposits of lakes and is believed to have formed from a hydrous sodium-silicate precursor such as magadiite ($NaSi_7O_{13}(OH)_3 \cdot 3H_2O$). Magadiite and Magadi-type chert were first described by Eugster (1967, 1969) and Hay (1968) in Pleistocene lacustrine beds in the vicinity of Lake Magadi, East Africa. Magadiite has been found at Alkali Lake, Oregon (Rooney and others, 1969) and in spring deposits in California (Gude and Sheppard, 1969, 1972; Starkey and Blackmon, 1979). Magadi-type chert has been recognized in Jurassic through Pleistocene age continental sedimentary rocks in the western United States (Surdam and others, 1972; Sheppard and Gude, 1974, 1980) and in a Cambrian playa-lacustrine sequence in Australia (White and Youngs, 1980).

This report describes the occurrence and nature of chert, tentatively identified as Magadi-type chert, in the basal part of the middle Eocene McBean Formation in Orangeburg and Calhoun Counties, South Carolina (fig. 1). The history of stratigraphic nomenclature for Eocene age rocks in the southeastern Atlantic Coastal Plain has been long and complex. Huddlestun and Hetrick (1979) proposed revision of part of the nomenclature in Georgia, and Huddlestun (1982) reviewed the terminology for eastern Georgia and western South Carolina. In the geologic map of the Woodford quadrangle, South Carolina, Kite (1984) included the McBean Formation in a unit termed undifferentiated middle Eocene sediments. In the area of this report the McBean Formation, together with the Congaree, Warley Hill, and Barnwell Formations, constitute the middle and upper Eocene Orangeburg Group of Siple and Pooser (1973). These formations are a sequence of dominantly clastic rocks northwest of the Orangeburg scarp in the Atlantic Coastal Plain of South Carolina (Cooke, 1936; Siple, 1959; Pooser, 1965). The stratigraphic relationships of some of the units are shown in the cross section accompanying figure 1.

The Congaree Formation consists of poorly sorted sand, interbedded sand and mudstone, and indurated siltstone and sandstone layers. The Warley Hill Formation is characteristically glauconitic with the glauconite content increasing upsection. Where it interfingers with the Congaree to the northwest, the Warley Hill is a glauconitic quartzose sand. Downdip to the southeast, it interfingers with the Eocene Santee Limestone and is a glauconitic calcarenite.

The McBean Formation is chiefly a silty, clayey, fine- to medium-grained sandstone with vague bedding. Siltstone and claystone laminae are a minor component. Cooke (1936) stated that the contact of the McBean with the underlying units is an erosional unconformity, and according to Pooser (1965, p. 19), the contact of the McBean with the underlying Congaree and Warley Hill Formations (fig. 1, cross section) is "generally marked by boulders and beds up to 2 ft. (0.7 m) thick of silica-cemented sandstone and chertlike material". The upper Eocene Barnwell Formation unconformably overlies the McBean Formation to the northwest and southwest and has

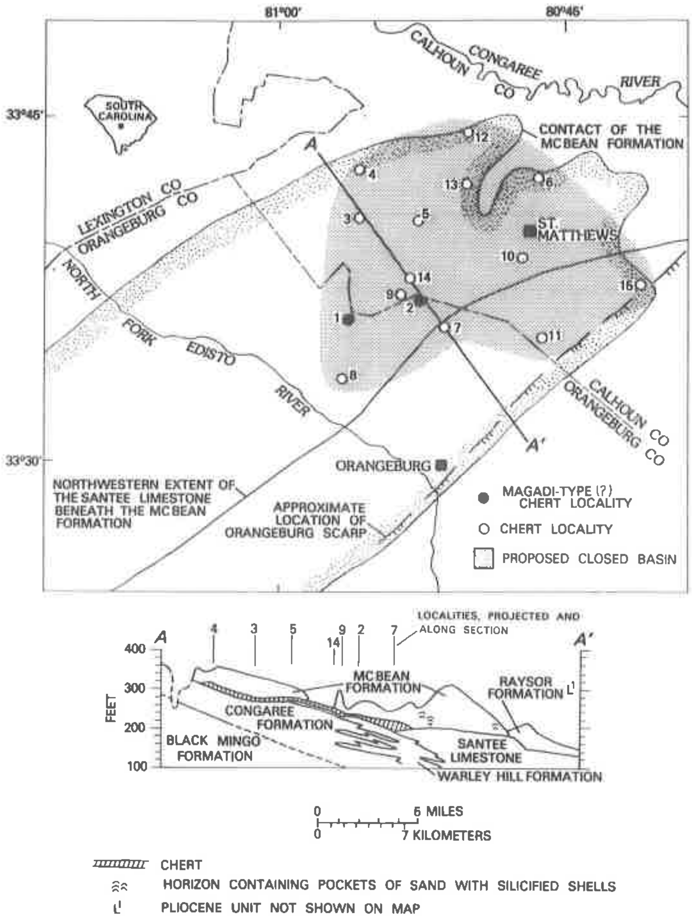

Figure 1. Map showing distribution of McBean Formation in Orangeburg and Calhoun Counties (modified from Pooser, 1965), Magadi-type (?) chert localities and other chert occurrences (table 2), and approximate area occupied by the proposed closed basin.

been mapped within the immediate study area (Kite, 1984). Cooke (1936) estimated the thickness of the McBean to be about 30 m.

Siple (1959) and Pooser (1965) assigned the Congaree and McBean Formations to estuarine and back barrier depositional environments. The glauconitic Warley Hill Formation apparently was deposited in a nearshore environment intermediate between the Congaree and the open- or normal-marine Santee Limestone.

A chert nodule collected from the McBean Formation in 1975 at an unrecorded location near St. Matthews, South Carolina, was tentatively identified as Magadi-type chert by Richard A. Sheppard and James A. Gude, 3d (oral commun., 1979). This report is based on a brief reconnaissance search, carried out in 1980, for similar appearing chert in roadcut exposures of the Congaree and McBean Formations between Aiken and St. Matthews, South Carolina.

OCCURRENCE AND DESCRIPTION OF CHERT

Chert, both in the form of chalcedony and as opal C-T, is a relatively common constituent of the Orangeburg Group (Cooke, 1936; Pooser, 1965; and Siple and Pooser, 1973) and of Tertiary sedimentary rocks of the southeast Atlantic Coastal Plain in general (Carver, 1980). Fragments of chalcedonic chert, opal-cemented sandstone and fullers earth (opal C-T and smectite) were found in the Congaree and McBean Formations at several localities in the reconnaissance area. Magadi-type (?) chert, however, was found only in the McBean Formation, at two localities north-northwest of Orangeburg (fig. 1). Neither locality is the unrecorded location from which Magadi-type (?) chert was collected in 1975. Kite (written commun., 1984) has identified chert similar in appearance to Magadi-type (?) chert in undifferentiated middle Eocene sediments at four localities in the southeast and south-central part of the Woodford 15-minute quadrangle. These localities are not shown on figure 1.

Locality No. 1

Magadi-type (?) chert occurs east of Little Limestone Creek on county road 77. This occurrence corresponds to Cooke's locality 118 (1936, p. 65) and Pooser's locality 38-8 (1965, p. 72). The present exposure is about 2 m high and 6 m long, and consists of highly weathered, mottled medium-orange-brown, very fine grained sandy silt containing irregularly shaped chert nodules. Most of the chert nodules occur in a lenticular bed 2 m long and 0.5 m thick; however, nodules several centimeters long are present as obscure layers throughout much of the outcrop. The nodules are horizontally flattened, 1-20 cm long and as much as 8 cm thick. The chert appears white and chalky throughout, and many nodules have the reticulated pattern common to most Magadi-type chert on at least one surface (fig. 2). Some chert nodules in the lenticular bed contain molds of mollusk shells and cheilostome bryzoans.

One chert mass about 30 cm long and 10 cm thick (vertical axis) displays features believed to indicate soft-sediment deformation. It consists chiefly of an aggregation of individual lenticular, curved, and flattened pieces of chert (2-3 cm across, less than 1 cm thick), subhorizontal to inclined, closely interleaved, and separated by thin septa of green and brown, waxy-textured smectite. The surfaces of many individual chert pieces in the aggregation show parallel striations 2-3 mm wide, inferred to be extrusion grooves produced during soft-sediment deformation. This internal fabric of the chert mass is similar to that which could be expected to result from subequal vertical loading of a layer of material having the consistency of toothpaste or putty, and resting on a slightly inhomogeneous substrate.

Locality No. 2

This locality corresponds to Pooser's locality 38-14 (1965, p. 73). It is a roadcut along U.S. Route 21, 0.5 km south of the Calhoun-Orangeburg County line and 6 km east-northeast of locality no. 1. The outcrop is about 60 m long and 3-5 m high. White, chalky appearing chert nodules occur in at least three lenticular layers 0.3-0.6 m thick and as much as 10 m long. The layers are interbedded with 0.6- to 1.0-m-thick beds of medium-gray, silty, very fine to medium sand and mottled red, sandy and silty clay. The nodules at this locality also contain shell molds, but no soft-sediment deformation features were seen, except for the general flattening of the nodules. Most nodules are reticulated on all surfaces. They are 1-15 cm long and about 5 cm thick, slightly smaller than those at locality no. 1, and are more regularly shaped.

Description of chert

Four Magadi-type (?) chert nodules, typical of samples collected at localities 1 and 2, are shown in figure 2. In these samples, the cracks which comprise the characteristic surface reticulation of Magadi-type chert (thought to be a result of the greater than 25 percent volume decrease associated with the conversion from magadiite to chert) vary considerably within a single sample, both in depth and in closeness and regularity of spacing. This variation in the reticulated surface pattern is atypical of many Magadi-type chert nodules (for example, see illustrations in Hay, 1968; Surdam and others, 1972; Sheppard and Gude, 1974); however, the surface reticulation and general form of two samples of Magadi-type chert shown in the report by Surdam and others (1972, fig. 5) appear to be quite similar to the Magadi-type (?) chert from South Carolina.

The chert is white to pale yellowish orange, is commonly iron stained in the

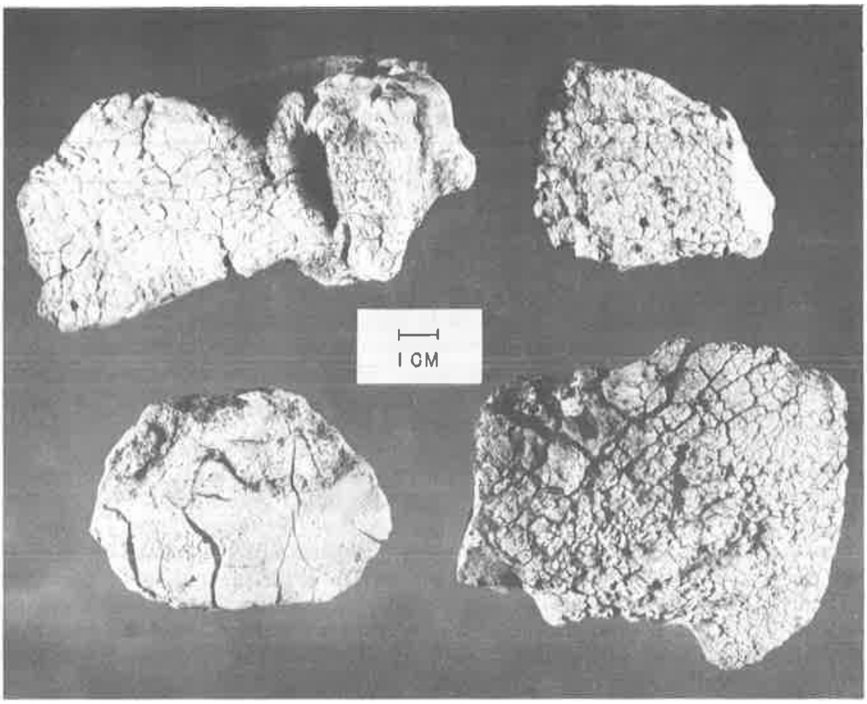

Figure 2. Magadi-type (?) chert nodules from McBean Formation, S.C. (locality 1) showing characteristic surface reticulation.

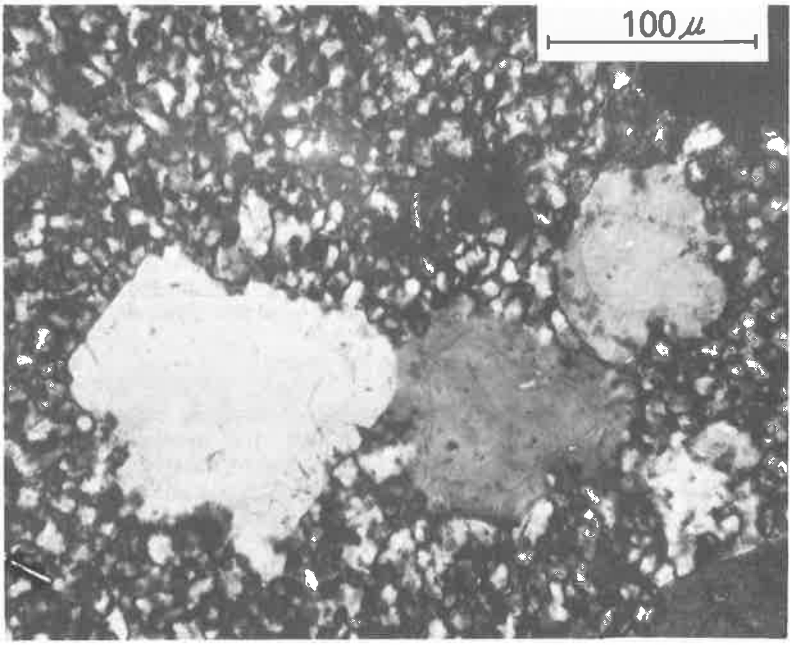

Figure 3. Photomicrograph of Magadi-type (?) chert from locality 1 showing unoriented fabric of equigranular chalcedony, and quartz overgrowths on silt and sand size detrital quartz grains.

surface cracks, and on fresh breaks appears chalky throughout. Only surfaces exposed to weathering are actually soft, however. The soft weathered layers are 1-5 mm thick and consist of powdery masses of disaggregated chalcedony crystals. At low magnification, numerous small pits can be observed within the soft layers. The pits are equidimensional or slot shaped and probably were the sites of detrital quartz grains and mica flakes. The soft layers are interpreted to result from weathering because they are present on old broken surfaces of the chert in addition to the original exterior surfaces of the nodules. Thus, these soft outer layers probably do not represent the opaque white coating commonly observed on unweathered Magadi-type chert nodules (Hay, 1968, p. 267; Sheppard and Gude, 1974).

The chalky appearance of the chert is attributed to the effects of weathering in the humid temperate to subtropical climate that has prevailed in the southeastern United States since the middle Eocene. As an example, in southwest Virginia I have observed many Paleozoic chert fragments in residual limestone soils of late Tertiary age that have been altered from the normal vitreous luster of fresh Paleozoic chert to white, light-gray, or yellow chalky-looking chert. Similarly, it may be assumed that the Magadi-type (?) chert of the McBean Formation may originally have had a vitreous luster. Because of the relatively high porosity and permeability of the sandy McBean Formation, however, it is not likely that any fresh chert remains.

Although chalky in appearance, the interiors of the Magadi-type (?) chert nodules are hard and consist of mosaics of unoriented, equigranular chalcedony crystals 2-10 μ across (fig. 3). Some nodules are composed mostly of smaller crystals, others chiefly of larger crystals, and still others are mixtures of finely crystalline areas and coarsely crystalline areas. Chalcedony crystals in pore fillings and adjacent to cracks are as much as 30 μ across.

The chert contains about 4 percent detrital quartz and 4 percent pore space. Inasmuch as many pores probably represent quartz grains which were plucked out during grinding of the thin sections, the amount of detrital quartz initially present may have been nearer 6 percent. Concentrations of quartz grains, unequally distributed throughout the chert, do not appear to define bedding planes. Detrital heavy minerals consisting chiefly of hornblende, chlorite, muscovite, and biotite are present in trace amounts. No feldspar was observed.

Most detrital quartz grains are subangular, but sparse well-rounded grains were observed. About 60 percent of the detrital quartz is silt size (25-64 μ) and 40 percent is very fine sand size (64-125 μ); the mean grain size is 60μ. Quartz grains in the more coarsely crystalline chert nodules are encased by quartz overgrowths as thick as 20 μ (fig. 3). No overgrowths were observed on quartz grains in finely crystalline (2 μ) chert nodules except within pockets of more coarsely crystalline chert.

Chemical Composition

Chemical analysis by X-ray fluorescence spectroscopy indicates that chert from locality no. 1 consists of 98.3 percent silica (table 1). The principal impurities--aluminum, iron, magnesium, and sodium--may be contained chiefly in detrital heavy minerals and small amounts of smectite.

GENESIS OF THE CHERT

Magadiite, the precursor of Magadi-type chert, precipitates from alkaline brines of the type Na-CO_3-Cl-SO_4 with high silica content (Hardie and others, 1978). The pH of modern brines in Lake Magadi, Kenya and Alkali Lake, Oregon, is as high as 11.0 and the silica content is as much as 2,700 ppm (Jones and others, 1967). Both lakes are in areas of extensive volcanic-rock outcrop--the best source terrane for surface water of the appropriate chemistry to produce high-silica alkaline brine.

Precipitation of magadiite may be triggered by lowering of pH brought about by an influx of fresh water into an alkaline lake as, for example, by runoff following heavy rains. The unaltered precipitate is a white, powdery material composed of crystalline spherulites. When wet, magadiite has a putty-like consistency, but dry samples are hard and porcelaneous. At Lake Magadi, it occurs as finely laminated beds as thick as 0.6 m interbedded with lacustrine units; and at Alkali Lake, it occurs as veins 10-

Table 1. Major-oxide composition of Magadi-type chert from locality 1 [X-ray spectroscopic analysis made in U.S. Geological Survey laboratories. Analysts: J. S. Wahlberg, J. Taggart and J. Baker]

Oxide	Weight percent
SiO_2	98.3
Al_2O_3	0.23
$FeO+Fe_2O_3$	0.19
MgO	0.13
CaO	0.09
Na_2O	0.18
K_2O	0.05
TiO_2	0.03
P_2O_5	0.08
MnO	< 0.02
LOI	0.56
TOTAL	99.86

30 cm wide cutting playa sediments. Soft-sediment deformation features are common in bedded magadiite.

Magadiite begins to transform to chalcedonic chert rapidly, perhaps within a few hundred to a few thousand years of being precipitated (Hay, 1968; Eugster, 1969). Flushing by ground water has been suggested as the conversion mechanism (Eugster, 1969). The resulting chert is most commonly nodular and has a reticulated surface pattern that may be a result of the volume loss that accompanies the magadiite to chert conversion. The chert nodules are variously flattened, irregular, lobate, or spinose, and generally range from 3 to 20 cm in their longest dimension. Soft-sediment deformation features formed in the earlier magadiite phase are commonly preserved in the later chert phase. Crystal molds ascribed to evaporite minerals such as trona, pirssonite, gaylussite, and calcite have been noted in some Magadi-type chert (Eugster, 1969; Surdam and others, 1972; Sheppard and Gude, 1974).

Surdam and others (1972) thought the features of Magadi-type chert that indicate soft-sediment deformation in conjunction with volume reduction are uniquely characteristic. It is on the basis of these features that the McBean chert is tentatively identified as Magadi-type. Factors against the chert being Magadi-type are that criteria related to the depositional environment of magadiite, such as fine-grained lacustrine sediments or evidence of other evaporite minerals (trona, gaylussite, and pirssonite, for example), have not been recognized in the McBean Formation. Water chemistry poses an additional problem. The fauna of the McBean indicate a brackish to nearshore marine depositional environment (Cooke, 1936; Siple, 1959; Pooser, 1965). The relatively high content of SO_4 ions in normal marine and brackish water results in brines saturated with respect to sulfate minerals, rather than the Na-CO_3-Cl-SO_4 brine type from which magadiite and associated minerals, such as trona, are formed (Hardie and others, 1978).

In spite of these negative factors, I consider the reticulated surface pattern of the nodules and the soft sediment deformation features displayed by the chert mass at locality 1 to be compelling evidence in favor of an identification as Magadi-type (?) chert. Assuming that the McBean chert is Magadi-type (?) chert, then one of two inferences is possible based on known occurrences of magadiite; (1) some part of the McBean Formation is lacustrine, and a closed basin containing high-silica alkaline brine was present in the vicinity of Orangeburg and Calhoun Counties during the middle Eocene, or (2) high-silica alkaline springs were present in the area during the middle Eocene. Both inferences require that the area was subaerially exposed or at least isolated from marine water.

In the absence of additional field work, it is not possible to determine if the Magadi-type (?) chert was deposited in a lake or in springs; however, I prefer a lacustrine environment for the following reasons: water with the composition of high-

Table 2. List of data for localities shown on figure 1. All elevations given in feet for ease of comparison with topographic maps and published stratigraphic data; ? indicates data not available

Locality number					Elevation (ft)		
This study	Cooke (1936)	Pooser (1965) Exposure	Pooser (1965) Auger hole	Johnson and Heron (1965)	Chert interval	Base of McBean Formation	Underlying formation (beneath McBean)
1	118	L.38-8			248-265	245	Warley Hill
2		L.38-14			230-233	230	Warley Hill
3		L.9-47	A.9-8		269-272	269	Congaree[1]
4			A.9-9		297-302	297	Congaree
5		L.9-38			272-277	272	Warley Hill
6		L.9-19			253-254	253	Warley Hill
7		L.38-101	A.38-34		197-216	197	Santee
8	112	L.38-7	A.38-10		205-215	205	Warley Hill
9	120	L.9-35	A.9-1		233-235	233	transition[2]
10				Calhoun 19	223-237	223	Congaree
11			A.9-12	Calhoun 12	221-226	221	Santee
12	123				?	?	?
13		L.9-44			?	?	Warley Hill
14	121	L.9-43			241-246	?	Warley Hill
15	128				190-195	190	Warley Hill[3]

[1]This unit should perhaps be assigned to the Warley Hill Formation on the basis of glauconite content.

[2]Warley Hill-Congaree transition zone.

[3]Cooke (1936) assigned the sediments at this exposure to Pleistocene underlain by the McBean. On the basis of Cooke's descriptions, however, in this report these units are assigned to McBean underlain by Warley Hill.

silica alkaline brine is more common in lakes than in springs; lacustrine deposits are more apt to be preserved than spring deposits; and the relatively large area in which the chert occurs is more compatible with a lacustrine depositional environment. As stated above, a lacustrine environment has not been recognized previously for the McBean Formation; however, both the lithology of the McBean and the middle Eocene paleogeography of the area are permissive of this interpretation. The final section of this report is a speculative reconstruction of the proposed closed basin in Orangeburg and Calhoun Counties and of conditions that might have favored its development in this particular area during the middle Eocene.

PROPOSED MIDDLE EOCENE LAKE

Size and location

The estimated areal extent of the proposed lake is shown in figure 1. The estimate is based in part on Magadi-type (?) chert localities 1 and 2 and the unrecorded locality near St. Matthews. Other points shown on figure 1 are exposures and auger holes where chert was observed by Cooke (1936), Johnson and Heron (1965), and Pooser (1965) that, based on their descriptions (table 2), may be similar to the Magadi-type (?) chert of localities 1 and 2. Because these two localities were found in a single day, the probability is high that chert at some of the other locations also may be Magadi-type (?). The size of the proposed alkaline lake thus defined by the occurrence of chert is about 450 km^2. The area of the lake could be extended to the west, across the North Fork of the Edisto River, if the Magadi-type (?) chert localities of Kite (written commun., 1984) were included in figure 1. The stratigraphic position of Kite's localities, relative to the base of the McBean Formation as defined by Cooke (1936) and Pooser (1965), is unclear, however.

Age and duration of lacustrine environment

Within the estimated area of the proposed lake, chert has been reported only in the basal 0.3-6.5 m of the McBean Formation (Cooke, 1936; Johnson and Heron, 1965; Pooser, 1965). This occurrence suggests that magadiite was precipitated during only one relatively short period soon after the formation of the proposed closed basin. This, and the lack of abundant finely laminated, nonmarine clay or silt units in the McBean above the chert horizon, suggests that the proposed basin may have been a dry alluvial plain during much of the remainder of its history. Shell molds in the Magadi-type (?) chert are interpreted to have been a middle Eocene fauna living on the sea floor immediately prior to subaerial exposure and formation of the proposed closed basin. Thus the fauna is inferred to be a death assemblage and, therefore, would indicate only the maximum age of the lacustrine facies of the McBean Formation.

Formation of a closed basin

Mechanisms by which a closed basin could have been formed in this area are (1) closing off of a back-barrier lagoon from marine influence by growth of a barrier beach ridge, (2) eustatic lowering of sea level, (3) tectonic uplift, or (4) some combination of the three mechanisms. Because of the proximity to a shoreline, as indicated by the silicified fauna at the base of the McBean Formation and by the faunas of the underlying Congaree and Warley Hill Formations, any of the mechanisms is reasonably possible. The first mechansim by itself, however, is considered least likely to have led to the formation of the proposed closed basin. There are two reasons for this. First, basins formed in this way tend to be short lived along stable shorelines even if the shoreline is one of fairly low-energy. Second, the amount of isolation from marine and brackish water would probably not be adequate to produce an alkaline brine.

With the available data, it is not possible to argue in favor of a drop in sea level as opposed to tectonic uplift (or vice versa) as the more important of the two remaining mechanisms for formation of the proposed closed basin. Both were operative in the area during the middle Eocene. Tschudy and Patterson (1975) noted evidence in Coastal Plain sediments in central Georgia for at least one shoreline fluctuation during the middle Eocene. Vail and others (1977) showed drops in global sea level at the beginning and end of the middle Eocene. Ward and others (1979) interpreted the burrowed and beveled surface separating the middle Eocene Moultrie and Cross Members of the Santee Limestone (east of the area in Berkeley and Georgetown Counties, figure 4) to have formed during a marine regressive and transgressive cycle.

Two features are present in the region which could have been sites of tectonic uplift during the middle Eocene; the Cape Fear arch and the northern boundary of the South Georgia rift. Stratigraphic evidence indicates that the Cape Fear arch (fig. 4), a broad northwest-trending basement high near the South Carolina-North Carolina border (King, 1969), was tectonically positive during much of the late Mesozoic and Cenozoic. For example, Lower Cretaceous strata are present north of the arch but not to the south; the middle Eocene Santee Limestone pinches out on the southwest flank of the arch; Upper Cretaceous strata are exposed over the crest of the arch; and Pliocene and Quaternary shoreline features have been warped upward over the arch relative to contemporaneous shoreline features in South Carolina, Georgia, and Florida (Winker and Howard, 1977).

The South Georgia rift is a large east-northeast trending lower Mesozoic extensional basin that traverses South Carolina, Georgia, southeast Alabama, and the Florida panhandle (Popenoe and Zeitz, 1977; Popenoe, 1977; Gohn and others, 1978; Chowns, 1979; Daniels and others, 1983). The basin is filled with lower Mesozoic sedimentary rocks, basalt, and diabase and is overlain by Cretaceous sedimentary rocks. Although the presence of the basin is well documented, it is nowhere exposed; the rift boundaries were determined by Daniels and others (1983) using aeromagnetic and gravity maps and well data. In a study of the Charleston earthquake of 1886 edited by Gohn (1983), many authors cited evidence indicating that reactivation of faults associated with the rift in the Charleston area may be responsible for the historic seismicity of that region. Although Cenozoic tectonic movement has not been assigned previously to the northern boundary of the South Georgia rift in South Carolina, figure

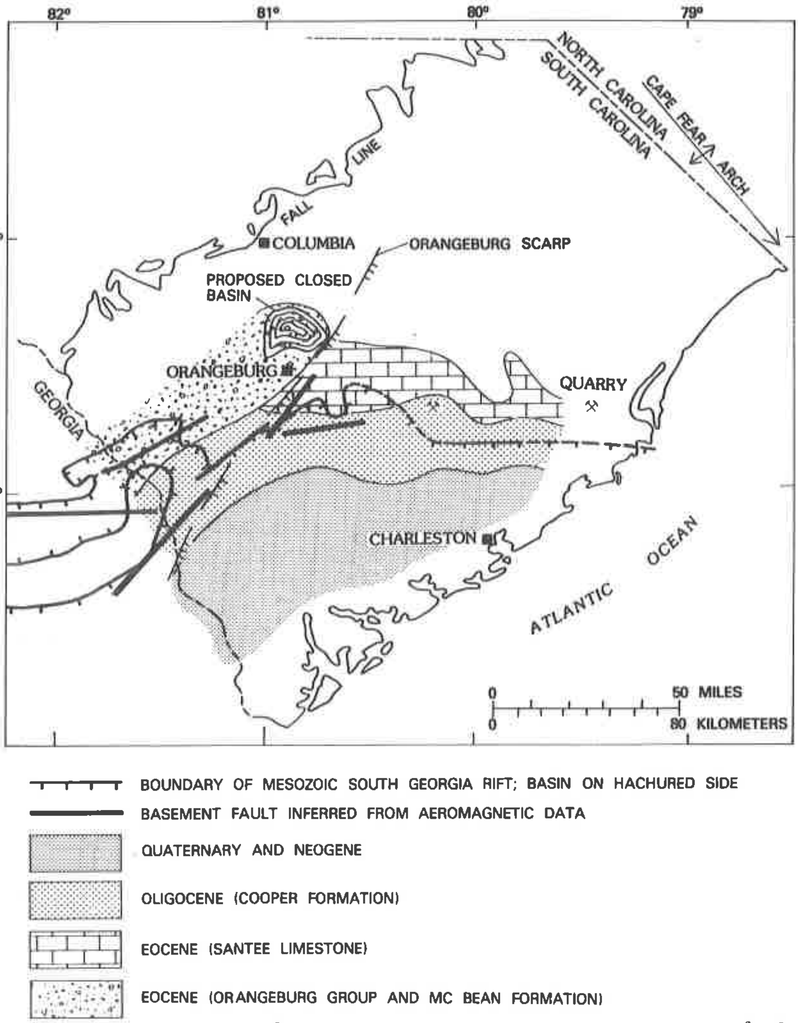

Figure 4. Map showing location of Cape Fear arch and northern boundary faults of the Mesozoic South Georgia rift basin (Daniels and others, 1983) in relation to the proposed closed basin, the Orangeburg scarp, and Eocene and Oligocene formation contacts (Cooke, 1936; Pooser, 1965). The quarries are localities where the beveled surface separating the Cross and Moultrie Members of the Santee Limestone is exposed (Ward and others, 1979).

4 shows that there is considerable parallelism and coincidence of the northern rift boundary and its associated inferred faults with Eocene and Oligocene formational contacts and with the Orangeburg scarp, a topographic feature. I suggest that the parallelism and coincidence could have resulted from syndepositional tectonic activity along this border of the South Georgia rift, from the middle Eocene through the Oligocene. Minor tectonic activity involving reactivation of the rift border faults could have controlled the shape and location of the open marine depositional basin on the southeast and the fluvial to nearshore marine platform on the northwest.

Because the mechanisms described above (sea level fluctuation and minor tectonic uplift) are not unusual events either in terms of location or time, I consider the probability to be fairly high that one or more of the events did occur during the middle Eocene and did result in the formation of the proposed closed basin.

Source of water

The problem of water chemistry was noted previously as an objection to the chert in the McBean Formation being Magadi-type (?). To counter this objection, it must be assumed that the marine and brackish water in which the underlying Congaree and Warley Hill Formations and Santee Limestone were deposited was flushed out of the near-surface ground-water system prior to development of a lake in the proposed closed basin. If very much of the original water had remained, a brine rich in SO_4 ions would have been the result of evaporation rather than an alkaline brine rich in sodium carbonate minerals. The mechanisms of sea level lowering and (or) minor tectonic uplift that was proposed for formation of the closed basin in the preceding section would probably have assured fairly complete flushing of marine water from the surface and near-surface ground-water systems.

A second problem involving water chemistry is the dominant lithology of the source terrane for the water that was ponded in the proposed closed basin. Eugster (1967) noted that volcanic or igneous terranes are obvious sources for surface water of the appropriate composition to produce an alkaline brine; that is, rich in sodium and carbonate ions and relatively low in calcium, chlorine, and sulfate ions. With the exception of the Magadi-type (?) chert in the basal part of the McBean Formation and an occurrence of Magadi-type chert reported in Florida (Strom and others, 1983), all other occurrences of Magadi-type chert, magadiite, and most alkaline lakes are (or were) in volcanic terranes or are associated with air-fall tuff beds.

In the case of the McBean Formation, however, the terrane immediately surrounding the proposed closed basin in Orangeburg and Calhoun Counties during the middle Eocene was composed chiefly of clayey and sandy, calcareous to quartzose, unconsolidated to semiconsolidated Coastal Plain sediments ranging in age from Late Cretaceous to middle Eocene. Heron (1969) suggested that zeolite minerals in the Paleocene and lower Eocene Black Mingo Formation in South Carolina were formed by alteration of volcanic ash, but no air-fall tuff beds have been identified. Water derived as runoff from relatively mature, calcareous, clastic sediments such as these would probably have been too calcium- and sulfate-rich to produce a Na-CO_3-Cl-SO_4-type alkaline brine. Although two occurrences (besides this report) of Magadi-type chert in Coastal Plain sediments are known, both may be the result of somewhat uncommon conditions. Magadi-type chert occurs in the Gueydan (Catahoula) Formation in south Texas (B. B. Houser, unpublished data). There, the chert is associated with lacustrine clay, silt, and interbedded air-fall tuff, and a significant part of the sediments (and inflow water of the lake) was probably derived from the volcanic terrane of the Big Bend area of Texas and northern Mexico (McBride and others, 1968). Insufficient information regarding the other Magadi-type chert occurrence in Florida (Strom and others, 1983) precludes an evaluation of its depositional environment.

The source terrane for the water in the proposed closed basin in Orangeburg and Calhoun Counties, however, was probably the southern Appalachian Piedmont province located 40-50 km to the northwest rather than the Coastal Plain sediments surrounding the basin. The principal lithologies presently exposed in this part of the Piedmont are greenschist and amphibolite facies metamorphic rocks and less common intermediate to silicic plutons (Overstreet and Bell, 1965). Study of detrital heavy minerals in South Carolina Coastal Plain sedimentary rocks shows that the lithologic composition of the Piedmont province exposed during Tertiary time probably was not significantly different (Gohn and others, 1977).

This type of geologic terrane in conjunction with the deep chemical weathering produced by the dry (seasonal rainfall) tropical climate of the middle Eocene (Wolfe, 1978) may well have produced surface water of suitable composition to form high-silica alkaline brine upon evaporation. The total absence of detrital feldspar grains associated with detrital quartz in the Magadi-type (?) chert attests to the thoroughness of chemical weathering at this time. Alkali and silica ions leached from the abundant feldspar of the Piedmont presumably entered surface streams and were carried seaward in solution.

SUMMARY

The purpose of this report has been to record a new occurrence of Magadi-type (?) chert and to examine the implications of this occurrence for the depositional environment of the McBean Formation in Orangeburg and Calhoun Counties. The conclusions of the study are—

(1) the northeast end of the outcrop belt of the McBean Formation was subaerially exposed sometime during the middle Eocene, while to the southwest, deposition of the McBean continued in a near shore environment;

(2) the probable cause of subaerial exposure at the northeast part of the outcrop belt was a drop in sea level and (or) minor tectonic uplift of the Cape Fear arch or the northern boundary of the South Georgia rift;

(3) a closed basin was present on the newly exposed sea floor, perhaps as a result of tectonic tilting or enclosure by a barrier beach ridge;

(4) water that filled the closed basin was rich in silica and alkali ions derived from tropical weathering of the crystalline rocks of the southern Piedmont;

(5) evaporation of the lake water during dry seasons produced a high-silica alkaline brine from which magadiite precipitated;

(6) the lake was relatively short lived and the basin was either filled with fluvial sediments or the rim was breached and the lake drained.

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Rb-Sr AGES OF PALEOGENE PROVINCIAL STAGES, EASTERN GULF COASTAL PROVINCE

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ABSTRACT

We report here new radiometric dates (Rb-Sr model ages) on glauconite from formations adjacent to the Paleogene provincial stage boundaries of the eastern Gulf Coastal Province. The samples dated are biostratigraphically controlled by use of calcareous nannoplankton (NP) biozones. Ages considered to be anomalous were obtained for the Paynes Hammock Fm., the Chickasawhay Ls. and the Moodys Branch Fm. The other radiometric dates (in megannum) obtained are as follows:

Red Bluff Fm. (NP 21):	28.7 ± 5.4 Ma
	29.6 ± 1.9
	29.7 ± 0.6
	32.5 ± 0.9
Tallahatta Fm. (NP 15):	42.7 ± 0.9
(NP 14):	40.9 ± 1.1
Hatchetigbee Fm.:	
Bashi Marl Mbr. (NP 9):	54.1 ± 0.5
Nanafalia Fm.:	
Ostrea thirsae beds (NP 7):	50.6 ± 1.2
Clayton Fm. (NP 1):	58.7 ± 2.1
(NP 1):	62.7 ± 10.7
Clayton Fm./Prairie Bluff Chalk contact	
(NP 1/Cretaceous boundary:	64.7 ± 5.7

INTRODUCTION

Very few radiometric age determinations have been published for the Paleogene rocks of the Gulf Coastal Province. The most recent dates of which we are aware are three Rb-Sr isochron ages published by Harris and Fullagar (1982, 1984) on formations in Mississippi and Alabama. Ghosh (1972) earlier reported a number of K-Ar dates on selected formations in the region. A total of 20 K-Ar dates were obtained by Ghosh in Mississippi and Alabama. Samples dated were not, however, closely controlled biostratigraphically during Ghosh's investigation. The main purpose of this study is to report a number of new, biostratigraphically controlled radiometric (Rb-Sr) dates on strata adjacent to the Gulf Coast Paleogene provincial stage boundaries.

This work forms part of a larger project aimed at refining the correlation of the eastern Gulf Coast provincial Paleogene stages with the European standard stages, using Martini's (1971) calcareous nannoplankton biozones as correlation tools. All Paleogene formations in the study area have been investigated for nannoplankton, and the results are reported in Kronman (1982), Siesser (1983), Fitzgerald (1984), Siesser (1984a,b) and Siesser et al. (in press).

Fig. 1 shows where samples used in this study were collected. Fig. 2 shows the lithostratigraphic and chronostratigraphic framework of the region. Siesser (1984a) has discussed the relationship between the lithostratigraphic units and chrono- stratigraphic units, and the basis for defining provincial stages in the area.

Samples were collected for Rb-Sr analysis in order to complement the biostratigraphic and chronostratigraphic work by assignment of numerical ages to the provincial stages. In this study, we focussed on obtaining radiometric dates from the formations that lie adjacent to the provincial stage boundaries, since the boundaries in effect define the time span of each stage. In practice, many of the samples we collected were later judged unsuitable for radiometric dating because of insufficient

Figure 1. Location map showing collection localities of samples used for Rb-Sr dating. County names are underlined.

glauconite content, evidence of weathering or reworking of the glauconite, etc. We accordingly had to use some samples not immediately adjacent to the stage boundaries (viz., A2a, A19c, A46a, A24a -- see Table 1). Furthermore, the low Rb-Sr ratios of many samples caused a large uncertainty in the calculated age.

Table 1 gives the stage, formation, calcareous nannoplankton biozone, reference, and radiometric age determined for each sample. Details of the location, lithology, and stratigraphic placement of each sample may be found in the references cited in Table 1.

METHODS

"Glauconite" grains were extracted from 14 stage-boundary samples whose accurate stratigraphic placement was assured by the initial biostratigraphic zonation. The term "glauconite" is used here in the sedimentological sense, i.e., rounded, mostly sand-size, green or black pellets occurring in marine sediments.

Samples were first soaked in distilled water for 48 hours. This process disaggregates the samples without having to crush the rocks, which could damage glauconite grains. Following disaggregation, the sediment was run through two wet sieves of 30 and 250 mesh, in order to separate the smaller and larger size fractions. Glauconite was next separated from the nonmagnetic fraction by running the dried sediment through a Franz Isodynamic Magnetic Separator. Glauconite is moderately paramagnetic, and the following instrument settings were found to achieve the best separations: 0.8 amperes, 10° roll (lateral inclination), 15° pitch (longitudinal inclination), and a vibration setting just below maximum. After the sediment had been passed through the separator two or three times (to concentrate the glauconite), the magnetic fraction was sent to Dr. Paul Fullagar, University of North Carolina, Chapel Hill, for Rb-Sr analysis. The final preparation, concentration, and analysis of the glauconite samples were conducted at Chapel Hill, and the following analytical information is summarized from Dr. Fullagar's report:

Samples were examined under a binocular microscope and weighed to estimate the glauconite percentage. Each sample was again run through a magnetic separator in

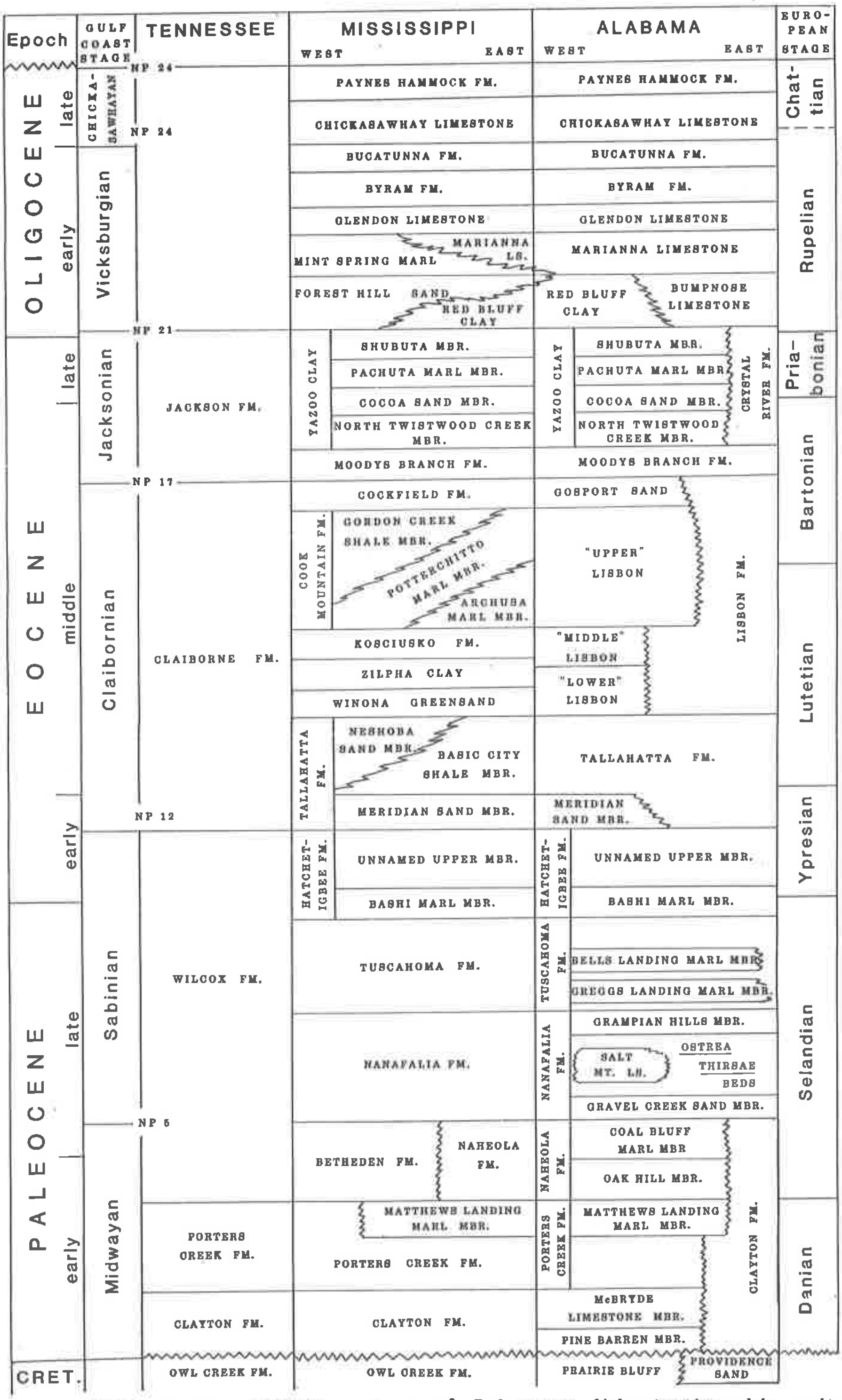

Figure 2. Generalized correlation chart of Paleogene lithostratigraphic units and provincial stages in the eastern Gulf Coastal Province. Hiatuses and facies relationships among units are not illustrated.

Table 1. Radiometric Ages and Biostratigraphic Zonation: Gulf Coast Paleogene

Stage Boundary	Formation/Member	Sample	Zone	Reference	Rb-Sr age (Ma)
Chickasawhayan-Anahuacian	Paynes Hammock Fm.	Ac1-8	NP 24	Fitzgerald (1984)	>45 (anomalous)
Vicksburgian-Chickasawhayan	Chickasawhay Ls	Aes-1	NP 24	"	>45 (anomalous)
Jacksonian-Vicksburgian	Red Bluff Fm.	A23m	NP 21	Siesser (1983)	28.7 ± 5.4
	Red Bluff Fm.	A27c	NP 21	"	29.6 ± 1.9
	Red Bluff Fm.	Mc1-1b	NP 21	Fitzgerald (1984)	29.7 ± 0.6
	Red Bluff Fm.	Mwa-15a	NP 21	"	32.5 ± 0.9
Claibornian-Jacksonian	Moodys Branch Fm.	M4a	NP 17	Siesser (1983)	51.4 ± 4.8 (anom.)
Sabinian-Claibornian	Tallahatta Fm.	A24a	NP 15	"	42.7 ± 0.9
	Tallahatta Fm.	A23a	NP 14	"	40.9 ± 1.1
	Hatchetigbee Fm. (Bashi Marl Mbr.)	A46a	NP 9	"	54.1 ± 0.5
Midwayan-Sabinian	Nanafalia Fm. (Ostrea thirsae beds)	A19c	NP 7	"	50.6 ± 1.2
Upper Cretaceous-Midwayan	Clayton Fm. (Pine Barren Mbr.)	A2a	NP 1	"	58.7 ± 2.1
	Clayton Fm. (Pine Barren Mbr.)	A10b	NP 1	"	62.7 ± 10.7
	Clayton Fm.-Prairie Bluff Fm. composite	Braggs 1(3/4)	NP 1/Cret.	Zemo (1982)	64.7 ± 5.7

order to obtain a sample containing at least 95% glauconite. This fraction was passed through a number of sieves ranging from 20 to 144 mesh. Each sample was then hand-picked to obtain 0.1 g of 100% glauconite from one of the size fractions. The hand-picked samples were then leached in 0.1 N HCl acid for three minutes in an ultrasonic cleanser and then rinsed three times in demineralized water for approximately one minute each time. The clean samples were weighed and spikes of ^{84}Sr and ^{87}Rb were added with appropriate calibrations being made. Next the samples were dissolved in hydrofluoric and nitric acids, and these solutions were run through cation exchange columns to concentrate Rb and Sr. Samples were then analyzed in a thermal-ionization-source mass spectrometer.

^{87}Rb decays to ^{87}Sr (radiogenic product) at a constant and known rate. By measuring the present concentrations of ^{87}Rb and ^{87}Sr and calculating or estimating the original amount of ^{87}Sr present, the time since the formation of the glauconite can be calculated. Age calculations were carried out using the following equation:

$$(^{87}Sr/^{86}Sr)_N = (^{87}Sr/^{86}Sr)_O + {}^{87}Rb/^{86}Sr(e^{\lambda t}-1)$$

where $(^{87}Sr/^{86}Sr)_N$ = the measured isotopic ratio
$(^{87}Sr/^{86}Sr)_O$ - the initial ratio (at the time of formation)
$(^{87}Rb/^{86}Sr)_O$ is measured
e = the natural logarithm
λ = the ^{87}Rb decay constant (1.42×10^{-11}/yr)
t = the age.

An age is obtained by solving the equation for t.

The radiometric ages obtained are model, not isochron, ages, and model ages give larger errors than isochron ages. Model ages are obtained by using the given age equation and estimating values for $(^{87}Sr/^{86}Sr)_O$, whereas isochron ages have the $(^{87}Sr/^{86}Sr)_O$ calculated from multiple samples from the same rock unit. The assumed values used in the model-age calculations were taken from analyses of Gulf Coast fossil carbonates that yielded $^{87}Sr/^{86}Sr$ values for sea-water during the Phanerozoic (Peterman et al., 1970). The Paleogene glauconites were assumed to have had an initial $^{87}Sr/^{86}Sr$ ratio comparable to seawater at the time of their formation. Rb-Sr data are summarized in Table 2. The age uncertainty of each sample (Table 2) represents an

Table 2. Rb-Sr Analytical Data: Gulf Coast Paleogene.

Sample	NP	$^{86}Sr/^{88}Sr$	$(^{87}Sr/^{86}Sr)_N$	Rb ppm	Sr ppm	$^{87}Rb/^{86}Sr$	Assumed $(^{87}Sr/^{86}Sr)_O$	Age (Ma)
Braggs 1 (3/4)	K/T	0.11926	0.7189	195.0	45.84	12.32	0.7077	64.2 ± 5.7
A2a	1	0.11945	0.7353	207.7	18.20	33.11	"	58.7 ± 2.1
A10b	1	0.11948	0.7136	207.4	90.55	6.63	"	62.7 ± 10.7
A19c	7	0.11944	0.7531	212.2	9.74	63.29	0.7076	50.6 ± 1.2
A46a	9	0.11936	0.8073	270.2	6.070	130.0	0.7075	54.1 ± 0.5
A23a	14	0.11954	0.7465	229.1	10.023	66.38	0.7079	40.9 ± 1.1
A24a	15	0.22946	0.7518	240.2	9.640	72.41	"	42.7 ± 0.9
M4a	17	0.12020	0.7171	187.5	37.40	12.42	0.7080	51.4 ± 4.8
A23m	21	0.11944	0.7133	184.2	40.85	13.05	"	28.7 ± 5.4
A27e	21	0.11938	0.7233	224.7	17.83	36.53	"	29.6 ± 1.9
Mc1-1b	21	0.11889	0.7550	222.4	5.81	111.36	"	29.7 ± 0.6
Mwa-15a	21	0.11955	0.7428	197.8	7.61	75.50	"	32.5 ± 0.9
Aes-1	24	0.11965	0.7260	185.5	19.53	27.53	0.7082	45.6 ± 2.6
Ac1-8a	24	0.11902	0.7292	305.8	27.10	32.72	"	45.2 ± 2.2

estimate of two-standard-deviations (2σ) analytical uncertainty. For $^{87}Sr/^{86}Sr$ measurements, the 2σ uncertainty is $\pm$ 0.0005 or less; for $^{87}Rb/^{86}Sr$ the uncertainty is $\pm$ 1% or less.

MINERALOGY

"Glauconite" pellets are known to vary considerably in their mineralogy and chemistry (e.g., see papers in Odin, 1982a). Splits from the dated samples were analyzed by X-ray diffraction to determine the mineralogy of the glauconite in the samples. We found the following types of glauconite present in the splits, based on the X-ray classification of Bentor and Kastner (1965):

1) Well-ordered glauconite; distinguished by symmetric and sharp diffraction peaks at 10.1, 4.53 and 3.3 Å. Reflections ($11\bar{2}$) and (112) are present.
2) Disordered glauconite, distinguished by asymmetric, subdued peaks, broadened at the base. Reflections ($11\bar{2}$) and ($11\bar{2}$) are absent.
3) Interlayered glauconite (d(001) > 10.15 Å.

Samples Braggs 1, A10b, A2a, A19c, A46a, A23a, A24a, Mc1-1b, A27c and A23m predominantly contain ordered to disordered glauconite; all contain some interlayered glauconite as well. The best ordered glauconites are from the Nanafalia, Hatchetigbee and Tallahatta Formations. Samples M4a, Mwa-15a, Aes-1 and Ac1-8 predominantly contain interlayered glauconite and mixed-layer clays.

RESULTS

Upper Cretaceous-Midwayan Boundary

One sample (Braggs 1) was collected at the Cretaceous-Tertiary boundary in the Braggs section in south-central Alabama (see Fig. 1 and Zemo, 1982). This sample was dated at 64.2 $\pm$ 5.7 Ma, a date which seems in good agreement with those shown on recent time scales for the Cretaceous-Tertiary boundary; e.g., 65 $\pm$ 1.5 Ma (Curry and Odin, 1982), 65 Ma (Harland et al., 1982), 66.4 Ma (Palmer, 1983; Berggren et al., in press). The original dates presented by Ghosh (1972) have been recalculated using the ^{40}K decay constant and tables in Dalrymple (1979). Ghosh's (1972) recalculated dates adjacent to the Cretaceous-Tertiary boundary in Alabama and Mississippi are 61.5 Ma, 61.5 Ma and 61.0 Ma.

A sample (A2a) we collected about two feet higher in the Braggs section gave an unexpectedly young date of 58.7 $\pm$ 2.1 Ma; a sample (A10b) collected southeast of Braggs and some 20 feet above the Cretaceous-Tertiary boundary (but still in NP 1) gave a date of 62.7 $\pm$ 10.7 Ma.

Midwayan-Sabinian Boundary

Only one sample (A19c) collected from the formations near this stage boundary contained sufficient glauconite to allow Rb-Sr analysis, and this sample is from the *Ostrea thirsae* beds (Nanafalia Formation) (Fig. 2). The *Ostrea thirsae* beds have been assigned to Zone NP 7 (Siesser, 1983), and are thus considerably younger than the Midwayan-Sabinian stage boundary, which is within Zone NP 5 (Siesser, 1984b). Sample A19c gave a date of 50.6 $\pm$ 1.2 Ma, which is somewhat younger than expected. Zone NP 7 occupies the time interval of 55.0-56.1 Ma on the Curry and Odin (1982) scale, and 59.8-60.3 Ma on the Berggren et al. (in press) scale.

Sabinian-Claibornian Boundary

Three samples collected near this boundary were analyzed (Fig. 2 and Table 1). The upper Hatchetigbee is barren of nannoplankton, but the lower Hatchetigbee (the Bashi Marl) is assigned to Zone NP 9 (Siesser, 1983). The Bashi (A46a) gave a date of 54.1 $\pm$ 0.5 Ma. Ghosh (1972) reported dates (recalculated) of 49.2, 49.8 and 53.2 Ma for NP 9, and Berggren et al. (in press) show the interval as 57.8-59.1 Ma.

A hiatus occurs at the contact between the Hatchetigbee and Tallahatta Formations in the eastern Gulf Coast region (Siesser, 1983). A sample collected from the middle part of the Tallahatta (A23a) gave a Rb-Sr date of 40.9 + 1.1 Ma. This sample is in NP 14 (Siesser, 1983). Another sample (A24a), collected from what is considered to be "upper" Tallahatta, and tentatively assigned to NP 15 (Siesser, 1983), gave a date of 42.7 $\pm$ 0.9 Ma. These apparently "older over younger" dates are more congruent when adjusted for the analytical error of each. Ghosh (1972) obtained a (recalculated) date of 48.4 Ma for the Tallahatta.

Curry and Odin (1982) show NP 14 as the 42.5-45.5 Ma interval and NP 15 as 40.0-42.5 Ma. Berggren et al. (in press) show NP 14 as 50.0-52.6 Ma and NP 15 as 45.2-50.0 Ma. Our dates for NP 14 appear to be in closer agreement with the Curry and Odin scale.

Claibornian-Jacksonian Boundary

Only one of our samples from this stage boundary contained enough glauconite to merit analysis. This sample (M4a) is from the Moodys Branch Formation and has been assigned to Zone NP 17 (Siesser, 1983). This sample gave a date of 51.4 $\pm$ 4.8 Ma. Ghosh (1972) obtained a (recalculated) date of 38.3 Ma and Harris and Fullagar (1982) a date of 39.2 $\pm$ 3.2 Ma for the Moodys Branch in this area. Our date appears to be anomalously old, even after taking the large analytical uncertainty into consideration. The X-ray data for this sample suggest the anomalous age may be caused by insufficient mineral glauconite in the pellets.

Jacksonian-Vicksburgian Boundary

All the datable samples collected from this boundary are from the Red Bluff Formation, and all are assigned to NP 21. Table 1 shows the dates ranging from 28.7 $\pm$ 5.4 to 32.4 $\pm$ 0.9 Ma. The average age is about 30 Ma.

The Curry and Odin (1982) scale shows Zone NP 21 as the 33-34 Ma interval, whereas Berggren et al. (in press) show it as 35.1-37.0 Ma. Only sample Mwa-15a, and possibly A23m, on Table 1 seem reasonable dates in comparison with the other time scales (the Curry and Odin (1982) scale in particular); the other samples (A27c and Mc1-1b) are anomalously young.

Vicksburgian-Chickasawhayan Boundary

Few samples collected from the formations adjacent to this boundary contained suitable glauconite. The single sample finally chosen for analysis gave an age in excess of 45 Ma, which is clearly anomalous. Either older glauconite has been reworked into the Chickasawhay Limestone, or, as suggested by the X-ray data, the pellets contain insufficient mineral glauconite.

Chickasawhayan-Anahuacian Boundary

As with the last stage boundary, samples collected here were generally unsuitable for Rb-Sr analysis. Another date of greater than 45 Ma was obtained on the single sample analyzed. Again, based on the X-ray data, we suspect insufficient glauconitization resulted in an erroneous age.

CONCLUSIONS

The radiometric dates on the Gulf Coast Paleogene formations reported here are reconnaissance in nature, and were obtained to complement the biostratigraphic age assignments.

A survey of the relevant literature shows that radiometric dating of glauconite, either by the K-Ar or Rb-Sr method, is fraught with problems and contention -- the interested reader is referred to the numerous articles in Odin (1982a) for an exhaustive discussion of sources of error. The literature is, moreover, replete with critiques supporting or refuting various radiometric dates obtained. Despite the contentious nature of the subject, we feel that the Rb-Sr dates obtained during our investigation should be reported because: 1) there are currently so few radiometric dates available on the Gulf Coast Paleogene, and 2) our dates are closely tied to a refined biostratigraphic framework.

As mentioned earlier, Ghosh (1972) reported the most recent radiometric dates, but without biostratigraphic control. Hardenbol and Berggren (1978), recognizing the lack of this element in Ghosh's study, investigated the foraminifera in the stratigraphic levels examined by Ghosh and assigned the levels to standard planktic foraminiferal biozones. Odin (1982b) and Odin and Worsley (1982) provide detailed critiques of Ghosh's (1972) analytical data and ages. Curry and Odin (1982) have also strongly criticized Ghosh's results on analytical grounds, and state that "... of the 21 data selected in the work by Ghosh (1972), 16 must be considered as *very tentative* if not definitely unreliable."

We are not qualified to comment on the various arguments presented concerning the analytical accuracy of Ghosh's dates, or any other radiometric dates, including our own. We simply note that, where we obtained dates (by Rb-Sr) from the same formations as Ghosh (by K-Ar), our dates are significantly different. Nor is there a consistent difference: our single date on the Bashi Marl is older than all three of Ghosh's on the Bashi; our two dates on the Tallahatta are both younger than Ghosh's. Our date on the Moodys Branch is considered to be anomalous. We also merely note that our dates more closely correspond to the younger dates of the Curry and Odin (1982) scale than to the Berggren et al. (in press) scale for all formations dated.

Establishment of a "true" time scale is an elusive, perhaps unobtainable, goal, but one which we feel is, nevertheless, worthy of pursuit. In this study, we have obtained a number (admittedly few) of new radiometric dates for the provincial chronostratigraphic units of the Gulf Coast Paleogene. It is difficult to say if our dates are correct, too old or too young. We hope, however, that these new age determinations will stimulate additional radiometric studies in the region.

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ON A SINGLE ELEPHANT TOOTH: MOST PRIMITIVE MAMMOTH FROM COASTAL PLAIN OF SOUTHESTERN UNITED STATES

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ABSTRACT

A deciduous premolar from near New Bern, North Carolina, Atlantic Coastal Plain, represents the most primitive species of mammoth from the Southeastern United States. Probabilities of sampling its low plate number and very short length from middle and late Pleistocene *Mammuthus imperator* (Imperial Mammoths) are one chance in 100 and eight chances in 10,000. Its primitive morphology suggests that the premolar is from deposits of early Pleistocene or latest Pliocene age even though its precise locality is unknown.

INTRODUCTION

McCarten and others (1983) have indicated recently the presence of deposits older than middle Pleistocene (greater than 1.0 my) along the Neuse River, near New Bern, North Carolina, on the Atlantic Coastal Plain of the Southeastern United States. They indicated that the James City Formation of Du Bar and Solliday (1963) is of early Pleistocene age. Such an age determination is not inconsistent with Du Bar and Solliday's (questionable) Pliocene age for their formation. In this short paper I report upon a single fossil elephant tooth from the New Bern area. Although its precise locality is unknown, its morphology supports the hypotheses of McCarten and others (1983) and Du Bar and Solliday (1963) of the presence of deposits of early Pleistocene or Pliocene age in the region. The tooth, an isolated, lower "second milk molar" or third deciduous premolar (dP/3), was collected from along the Neuse River, near New Bern, some years ago by Druid Wilson of the U. S. Geological Survey (USGS), Washington, D. C.. It is housed in the Department of Paleobiology, National Museum of Natural History, the Smithsonian Institution (USNM), and curated as number 299779.

The time scale used here follows that in Madden (1981) with the early Pleistocene--medial-third of the Irvingtonian Land Mammal Age (LMA)--beginning about 1.5 mya; the middle Pleistocene--last-third or late Irvingtonian--beginning about 1.0 mya; and the late Pleistocene--nearly the entire Rancholabrean lma--beginning about 0.5 mya and ending .01 mya. "Geologic" or biochronologic ranges for mammoth species discussed are: *Mammuthus imperator* (Imperial Mammoths) ranges from the middle Pleistocene to earlier late Pleistocene, late Irvingtonian through earliest Rancholabrean lma's, or middle Kansan to early Illinoian glacials, and *M. columbi* (Columbian Mammoths) ranges from earlier late Pleistocene to early Holocene (Madden, 1981, tables 38,42).

Statistical and character abbreviations used are: $\underline{a}$, Alpha or Type I error; $\underline{CV}$ coefficient of variation; ET, average enamel thickness; L, length; $\underline{N}$, size of samples; σ, standard deviation of population; $\underline{OR}$, observed range of samples; P, number of plates; p, probability; PR, plate ratio (number of plates per 100 mm of tooth length); $\underline{s}$, standard deviation of samples; $\underline{U}$, mean of population; w, width; $\underline{X}$, individual observation or parameter; $\underline{\overline{X}}$, mean of samples and $\underline{z}$, statistic computed in $\underline{z}$ score statistical test.

Institutional abbreviations are: DMNH, Denver Museum of Natural History; ISM, Illinois State Museum, Springfield; ISUM, Idaho State University Museum, Pocatello, Idaho; LACM, Los Angles County Natural History Museum; MNHNP, Museum Nacional d'Histoire Naturelle, Paris; PM GVP, Putnam Museum, Davenport, Iowa; SAHSM, San Antonio Historical Society Museum, Texas; UADA, University of Alberta Department of Anthropology, Edmonton; UF, Florida State Museum, University of Florida, Gainesville; UNSM, State Museum, University of Nebraska, Lincoln; UUVP, University of Utah vertebrate paleontology collection, Salt Lake City; and UW, Geological Museum, University of Wyoming, Laramie. All parameters or measurements are maximum and

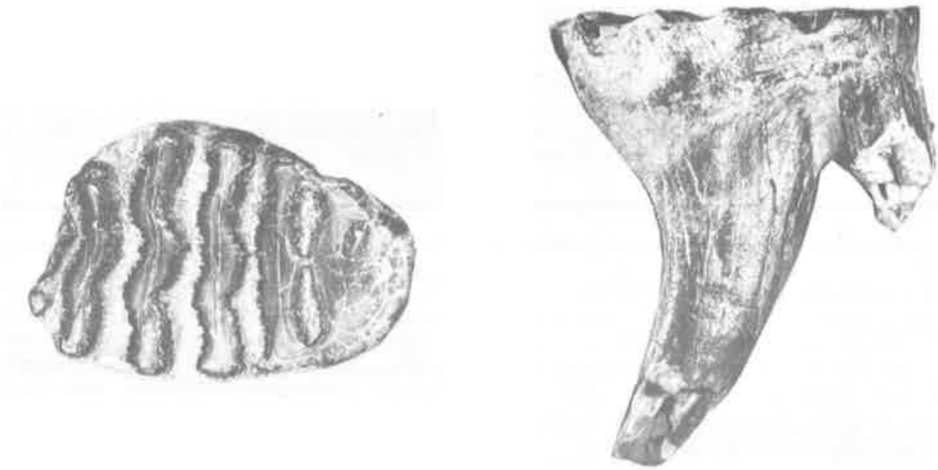

Figure 1. Top and internal views of USNM 299779, a left lower deciduous premolar from probable early Pleistocene deposits near New Bern, North Carolina. Length 48 mm. Courtesy of U. S. Geological Survey, Washington, D. C.

in millimeters (mm).

UNSM 299779 (Figure 1) is extremely primitive, the most primitive dP/3 from North America encountered to date. It has a low number of plates, extremely short length (and therefore great relative width), and thick enamel. The tooth has only six plates, is only 40 mm long, 34 mm wide, and has an average enamel thickness of 1.0 mm (Table 1). As shown by z score statistical tests, the probabilities for sampling its low plate number and extremely short length from populations similar or identical to those for middle Pleistocene and earlier late Pleistocene *Mammuthus imperator* are one chance in 100 and eight chances in 10,000 (Table 2). Additionally, its very short length shows that USNM 299779 does not represent late Pleistocene to early Holocene *M. columbi*. Its length falls below even the observed range of variation for that Rancholabrean species (Table 1). Although its plate number is identical to that for late Pliocene to early Pleistocene *M. meridionalis* (Table 1), the tooth cannot represent that primitive species because it is unknown from North America (see Madden, 1980, 1981, 1983). Contrary to popular opinion (Osborn, 1932; Maglio, 1973), *M. meridionalis* is restricted to Eurasia (Madden, 1981). The most primitive species of mammoth known from North America is *Mammuthus hayi* (Barbour, 1915), from the latest Pliocene (greater than 1.5 my) of California, Colorado, Nebraska, and Texas (Madden, 1981, 1983). However, USNM 299779 is not identified as *M. hayi* because it might equally well represent a new species of mammoth being described from the early Pleistocene of Colorado (Madden, unpub. data). Sample statistics calculated in Table 1 for characters of dP/3 in closely related *M. imperator* and *M. columbi* are based upon individual parameters for those two species given in Tables 3 and 4.

The single elephant tooth described here represents the most primitive mammoth known from the Atlantic Coastal Plain of the Southeastern United States. Although its precise locality is unknown, its primitive morphology supports McCarten and others (1983) and Du Bar and Solliday's (1963) hypotheses of the presence of deposits of early Pleistocene or Pliocene age near New Bern, North Carolina. Its low plate number and very short length indicate that USNM 299779 has a unique morphology, one decidedly less advanced than that known for dP/3 in closely related, middle Pleistocene and earlier late Pleistocene *Mammuthus imperator* (Imperial Mammoths).

Hitherto, the most primitive mammoth known from the Atlantic Coastal Plain of the United States was a lower third molar or M/3 from the Santee Canal of South Carolina (Madden, 1981; Hay, 1923, in part). This molar was described by Harlan (1823). It has at least 15 plates, is 368 mm long, and has a plate ratio of 4.1 (Harlan, 1923, p. 66; pl. 5, fig. 2). The Santee Canal sample is the easternmost occurrence of *M. imperator* (Madden, 1981).

Table 1. Parameters for USNM 299779 and sample statistics for characters of dP/3 in closely related North American **Mammuthus** *species and Eurasian* **M. meridionalis**[a].

Character	USNM 299779	M. imperator	M. columbi	M. meridionalis
Plates				
X or $\bar{X}$	6	7	7	6
s	-	0.45	0.49	0.55
CV	-	6.4	7.0	9.2
N	1	5	20	5
OR	-	6-7	6-8	5-6
Length (mm)				
X or $\bar{X}$	48	70	67	60
s	-	6.93	6.07	2.55
CV	-	9.9	9.1	4.2
N	1	3	20	5
OR	-	62-74	57-80	57-63
Width (mm)				
X or $\bar{X}$	34	35	38	34
s	-	5.0	5.59	4.56
CV	-	14.3	14.7	13.4
N	-	4	20	6
OR	-	30-42	31-52	25-38
Enamel thickness (mm)				
X or $\bar{X}$	1.0	1.3	0.8	1.3
s	-	-	0.26	0.26
CV	-	-	32.5	20.0
N	1	2	14	3
OR	-	-	0.5-1.3	1.0-1.5
Plate ratio				
X or $\bar{X}$	12.5	10.4	10.2	10.5
s	-	0.88	1.0	0.89
CV	-	8.5	9.8	8.5
N	1	5	21	6
OR	-	9.5-11.3	8.7-12.3	9.5-11.9

[a]Calulated from data given by Maglio (1973), Stehlin (1909) and Adams (1881).

Table 2. Z Score tests for plates and length (in mm) of USNM 299779 and dP/3 in Mammuthus imperator.

	USNM 299779		M. imperator
Plates			
X, U	6		7
N	1		5
σ	-		0.45
a		0.05	
z		-2.22	
p		0.01***	
Length			
X, U	48		70
N	1		3
σ	-		6.93
a		0.05	
z		-3.17	
p		0.0008***	

Table 3. *Individual parameters for characters of dP/3 in Mammuthus imperator.*

Specimen number	P	L	W	ET	PR	Reference
UNSM 2050	7	74	42	1.3	9.5	This Report
UNSM 2090	7	74	36	-	9.5	This Report
UNSM 39290	7	62	-	-	11.3	This Report
UNSM 4004-38	-	-	30	1.3	10.8	This Report
DMNH 1146	-	-	34	-	11.1	This Report
Private collection	7	-	-	-	-	Pontier (1933)
Private collection	6	-	-	-	-	Pontier (1933)

Table 4. *Individual parameters for characters of dP/3 in* Mammuthus columbi *(in mm).*

Specimen number	P	L	W	ET	PR	Reference
UF 18950	7	79	39	-	8.9	This report
UF 12238	7	67	44	-	10.4	This report
Private collection	7	71	-	-	9.9	Osborn (1942)
DMNH 1897	6	68	40	0.7	8.8	This report
USNM 24033	7	77	52	0.7	9.1	This report
UNSM 49244	6	61	32	1.2	9.8	This report
ISM DD-38-3, 40-1	7	66	34	-	10.6	This report
USNM 9230	8	73	37	-	11.0	This report
USNM 9232	-	-	34	0.6	-	This report
LACM 16433	7	67	36	0.6	10.4	This report
SAHSM no number	7	62	32	-	11.3	Hay (1924)
ISUM 81009	7	63	-	-	11.1	This report
ISUM Display	7	64	31	1.1	10.9	This report
ISUM 27756	7	65	40	0.8	10.8	This report
ISUM 27555	7	64	32	0.7	10.9	This report
ISUM 23427	7	66	39	0.9	10.6	This report
UADA Old Crow 14N	7	62	35	0.5	11.3	This report
ISUM 28039	7	57	33	0.7	12.3	This report
UUVP 8063	6	64	37	0.9	9.4	This report
UW 4286	6	68	42	1.2	8.8	This report
PM GVP 253	7	80	47	1.3	8.7	This report
MNHNP 77	-	-	43	-	10.1	Falconer (1863)

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CRETACEOUS PALEOSOLS FROM THE EASTERN GULF COASTAL PLAIN

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ABSTRACT

Paleosols on the pre-Upper Cretaceous crystalline rock surface and within the Upper Cretaceous (Cenomanian) Tuscaloosa Formation have contrasting fabrics and degrees of horizon differentiation that result from differences in parent material and post-depositional history. A well-developed lateritic soil on the crystalline rock surface formed in a warm, humid climate during the Early to Mid-Cretaceous. This paleosol is contrasted with a weakly developed alluvial soil that formed during Tuscaloosa deposition, and resulted largely from dewatering, minimal weathering, and compaction of mineralogically mature clays during a comparatively short time interval. Pedogenesis in both settings indicates that ancient subaerially exposed surfaces are well preserved in the Coastal Plain of the southeastern United States and are suitable for stratigraphic, paleoclimatological, and paleoenvironmental analyses.

INTRODUCTION

Paleosols or fossil soils have formed in a number of different weathering environments throughout the geologic record. In general, soil formation (pedogenesis) requires subaerial exposure and a relatively stable ground surface for an uncertain time interval, probably hundreds to thousands of years (Fitzpatrick, 1971). Consequently, paleosols represent a depositional hiatus and may be the only record of certain time intervals. Fossil soils provide paleoclimate data and, where consistent recognition is possible, can also be used for local and long distance stratigraphic correlation.

Paleosols are recognized by the same criteria used to identify modern soils--principally evidence of physical and pedochemical horizon zonation (see Morrison, 1967; Hunt, 1972). Both modern and ancient soils are differentiated by color, biogenic features, soil structure and texture, clay mineralogy, oxide distribution, and sometimes concretions (Fitzpatrick, 1971; Soil Survey Staff, 1975). Geochemical, micromorphological, and geomorphic criteria, presented by Loughnan (1969), Brewer (1964), and Daniels and others (1971), respectively, are also useful for the interpretation of fossil soils, but discussion of these criteria is beyond the scope of this study.

This paper documents the physical and stratigraphic characteristics of two Mesozoic paleosols in the Chattahoochee River Valley and provides an interpretation of their environmental settings. The older paleosol, on the pre-Upper Cretaceous crystalline rock surface along the Fall Line near Columbus, Ga., is described and compared with an alluvial flood-plain soil formed within the Upper Cretaceous Tuscaloosa Formation in the vicinity of Phenix City, Ala. (fig. 1). These soils are contrasting examples that formed in different geologic settings; nevertheless, both profiles show the kinds of pedogenic processes that occurred in the southeastern Coastal Plain throughout the late Mesozoic and Cenozoic.

PREVIOUS WORK

The small, but growing body of knowledge of Mesozoic soils includes some early studies from the Gulf Coastal Plain. Mellon (1937), Pryor and Ross (1962), following Sutton (1931) and Russell (1889) respectively, recognized different types of paleosols on Paleozoic rocks at several localities in the Mississippi Embayment. These workers concluded that a relatively long period of subaerial exposure and weathering preceeded deposition of Cretaceous sediments along the inner edge of the Coastal Plain.

Descriptions from elsewhere in the United States and the world have documented

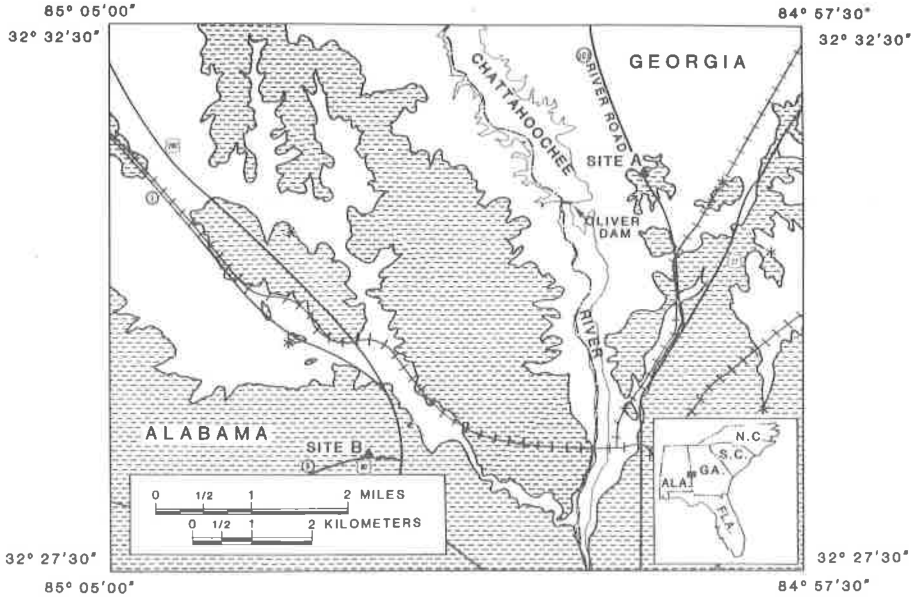

Figure 1. Map of the study area showing the distribution of the Tuscaloosa Formation (Cretaceous; shaded), Piedmont rocks (unpatterned), and paleosol localities described in this paper.

deep weathering and/or paleosol development on crystalline rocks during the late Mesozoic: for example, Blank (1978) in New York; Pavich (1974) in South Carolina; Reinhardt and Cleaves (1978) in Maryland; Singer (1974) in Israel; and Abbott and others (1976) in Mexico. R. F. Freeman (1981, unpublished manuscript) recognized a lateritic soil, the "Columbus laterite", at several localities in the Columbus, Ga., area and provided a brief physical and chemical description of several profiles. Our study of the paleosol at Site 1 extends and refines Freeman's findings. Description of Tuscaloosa paleosols is restricted to our analysis of a single paleosol at Site 2. To our knowledge, comparable treatment has not been given to paleosols elsewhere in the Tuscaloosa Formation, but lateritic paleosols are widely recognized in Tertiary sedimentary sequences in the Gulf Coast, because they were important in the formation of economic bauxite deposits from Georgia to Arkansas (e.g., Gordon and others, 1958).

STRATIGRAPHY

The petrology and structure of Piedmont rocks in the Chattahoochee River Valley near the Fall Line have been briefly described by Schamel and others (1980). Rocks of the Uchee Belt consist mainly of layered, migmatitic hornblende-biotite gneiss and amphibolite of intermediate mafic composition. The contact between the Tuscaloosa Formation[1] and the crystalline rock is generally irregular and one of low relief, probably less than 20 m in the Columbus area. In deeply weathered sections, the contact is difficult to define particularly where massive saprolite is overlain by poorly bedded Tuscaloosa. In addition, saprolite thickness varies considerably and the variable preservation of either residuum or lateritic soil at the crystalline rock surface makes precise boundary definition difficult (see Drennen, 1950; Eargle, 1955).

In the Chattahoochee River Valley, Tuscaloosa sediments are composed of poorly sorted arkosic sand, probably equivalent to the Gordo Formation (Tuscaloosa Group) of western Alabama, and red to red-green mottled kaolinitic clay similar to the Coker Formation of the Tuscaloosa Group. The Tuscaloosa Formation in this area consists of

[1] The usage of Tuscaloosa Formation in the Chattahoochee River Valley follows the usage of Reinhardt and Gibson (1981), regardless of location of exposures in Georgia or Alabama.

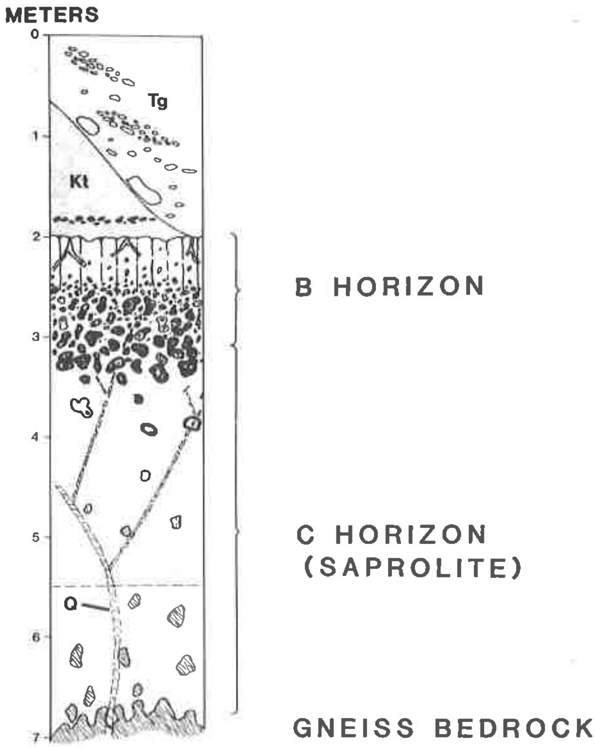

Figure 2. Measured section and description of the paleosol at Site A. Tg=Pliocene (?) terrace gravel; Kt=Tuscaloosa; Q=quartz vein cutting saprolite and parent gneiss.

continental deposits that accumulated in a variety of fluvial environments; marine equivalents occur considerably downdip and to the west (Applin, 1964; Sohl, 1964). Tuscaloosa deposits are typically organized into fining-upward sequences as much as 5 m thick, a sedimentary motif characteristic of meandering stream deposits.

LATERITIC PALEOSOL

The lateritic paleosol on the crystalline basement north of Columbus, Ga., (fig. 1, Site A; fig. 2) is a remnant of the pre-Tuscaloosa landscape. The parent material of the "Columbus laterite" at Site A consists of thinly banded gneiss, locally associated with coarse quartzose rocks containing potassium feldspar lenses and augen (T. B. Hanley, written commun., 1981).

The upper 100 cm of the Site A profile, interpreted as the B horizon, is a red to yellowish-red, silty clay with prismatic or blocky peds. Downward-branching channels or tubules, interpreted as root traces, are truncated and lie below the sharp upper contact with the overlying Tuscaloosa sand (fig. 3). Argillans or clay skins are well developed, and the soil matrix is slightly mottled. Dark, angular quartz granules are distributed throughout the profile. Few pisolites are found in the upper part of the soil; they increase in size and abundance with depth.

Between 55 and 125 cm, pisolites are concentrated in a complex zone, which contains a dense interlocking network of mottles (fig. 4). Pisolites are rounded and show a concentric structure in section that has definite boundaries, skins, or layers. Mottles are dark red to brown, and are more or less closely packed in a white soil matrix. The mottles are larger, as much as 20 cm in diameter near the base of the horizon, and are less densely packed. Between 100-125 cm, there is a gradual transition to the C horizon or saprolite. The soil matrix is predominantly pinkish-white and structureless; pisolites are generally smaller, and those having definite boundaries

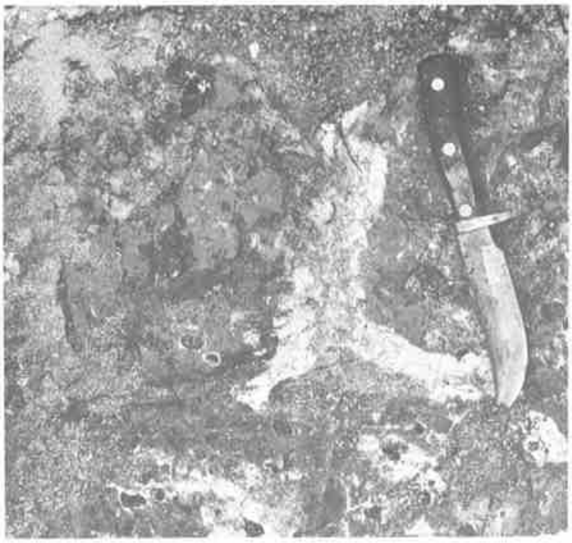

Figure 3. Downward-branching, sand-filled structure, interpreted as the trace of a root, in the upper part of the lateritic soil profile (Site A).

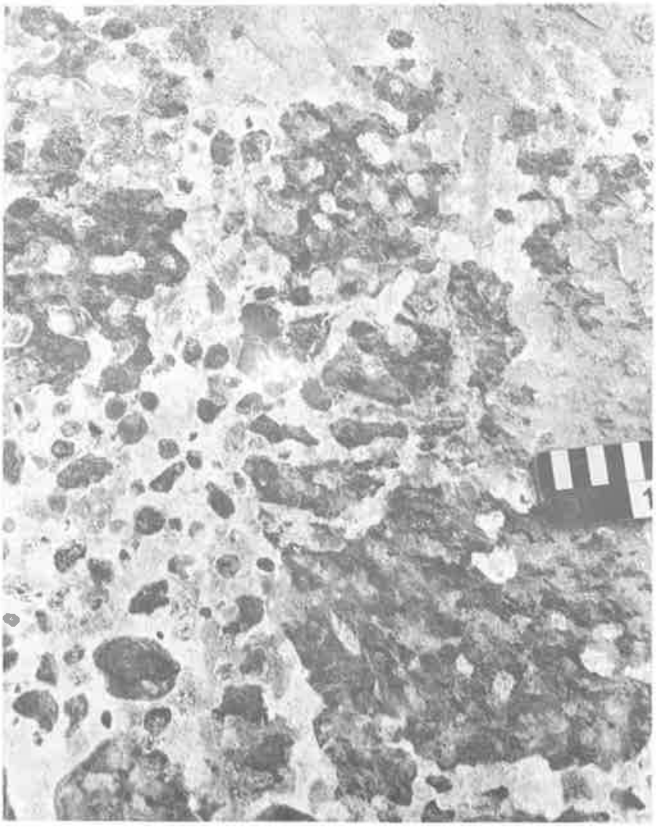

Figure 4. Mottled and pisolitic zone of the paleosol at Site A interpreted as the lower part of a B soil horizon. The dark iron-rich mottled interval is delineated by light (pallid) matrix.

are less abundant than those higher in the profile. Small patchy mottles are found throughout the matrix as are quartz granules.

Below about 350 cm, saprolite with preserved rock texture grades into relatively unweathered gneiss. The base of the profile is covered at this locality, but the pedogenic relationship between the paleosol, saprolite, and bedrock can be seen along fresh road cuts some 200 m north of Site A.

The contact between the paleosol and sediments of the Tuscaloosa Formation is sharp and irregular. Well-rounded fragments of the paleosol are contained in the lower 30 cm of the low-angle crossbedded, medium-to-coarse sand and grit. Subangular quartz, relict potassium feldspar, and muscovite constitute the Tuscaloosa sand, a composition readily derived from both the saprolite and crystalline rock. No major vertical changes in sediment texture or bedform are seen in the 3-m thick section of the Tuscaloosa, but iron-staining of the top 50 cm indicates a subsequent weathering, probably during the Quaternary.

The contact between the Tuscaloosa and overlying Pliocene (?) terrace gravel is sharp. The younger deposit fills a broad, ±4-m-deep, west-trending channel, that cuts through the entire Cretaceous section and into the saprolite immediately south of the measured section (fig. 2). Mineralogically, the terrace deposits are similar to the Tuscaloosa, but texturally they are dominated by weakly imbricated, discoidal pebbles 5 to 10 cm across. Subrounded to subangular, hematite-cemented sandstone boulders as much as 1.5 m in diameter are a conspicuous element of the terrace deposits at this locality. These boulders indicate an episode of weathering and surface cementation between the deposition of the Tuscaloosa and the terrace deposits.

Locally, remnants of the lateritic paleosol vary considerably in profile development. Most profiles, including the one described at Site A, are clay-rich, but also contain varying amounts of coarse quartz sand and granules. Organic A horizons either were not developed or were eroded; however, the tops of the same profiles are bioturbated and contain roots and possibly burrows. These soils have red, ferruginous B horizons that grade downward into mottled and gleyed C horizons and into unaltered bedrock at depth.

B horizons range from weakly indurated zones of iron oxide in some profiles to well-cemented, coarsely vesicular horizons with abundant red, iron-rich mottles and pisolite concretions (fig. 4). Preliminary x-ray analysis of the B horizon of these soils indicates that the mottles contain mainly kaolinite with varying amounts of quartz and small amounts of halloysite. The light-colored or pallied matrix between the red mottles is composed almost entirely of kaolinite with only trace amounts of halloysite. Detailed x-ray and chemical analysis of the profile at Site A is in progress.

The pisolites vary in size and shape, and are well developed in the lower part of the mottled zone. Subsequent hardening of pisolites and mottles seems to take place late during profile development (Daniels and others, 1978). Some authors suggest water-table lowering and consequent dehydration of the laterite as a possible mechanism of formation and induration (McFarlane, 1976).

These lateritic paleosols formed on a pre-Upper Cretaceous surface that appears to have coincided with the present Fall Line. Subsequent erosion has removed or modified the original low relief surface, and profiles everywhere are truncated. The irregular distribution of the "Columbus laterite" suggests that most profiles were (1) eroded prior to deposition of the Tuscaloosa Formation; (2) eroded during downcutting of the Chattahoochee River Valley; and/or (3) obscured by Quaternary weathering.

ALLUVIAL PALEOSOL

The fossil soil described from study Site B exhibits sedimentary and pedogenic features that indicate formation on a floodplain during and after alluvial aggradation. Figure 5 shows characteristic bedding features and lithofacies of the Tuscaloosa Formation in the Chattahoochee River Valley. The discussion of soil features, fabrics, and inferred pedogenic processes relates specifically to unit 2, near the base of the measured section.

Unit 2 consists primarily of clayey silt to silty clay that grades upward from a thin bed of indurated grit and well-sorted fine sand. Sand-filled fissures cut through the unit; fine sand has infilled the fissures to form a dense network of downward-

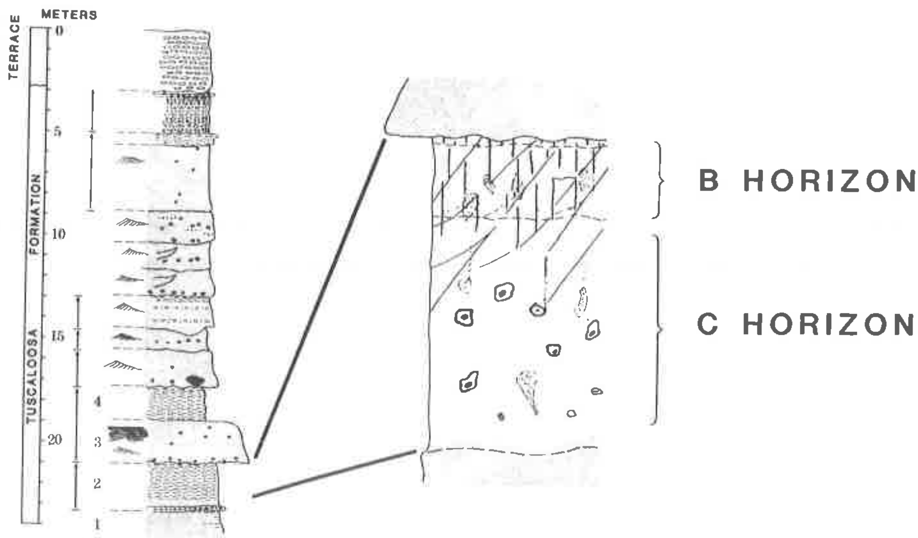

Figure 5. Measured section at Site B with detailed description of soil profile in unit 2 (Section modified from Reinhardt and Gibson, 1981).

branching stringers and local pods from the overlying crossbedded fine to medium sandstone. The unit is interpreted as an overbank or floodplain deposit associated with an aggrading, sandy alluvial system with low sinuosity. This conclusion is based on the fine-grained texture of the unit, its lack of internal bedding, and the lateral equivalence to small-scale crossbedded sand beds near the western margin of the exposure.

Unit 2 shows weak pedologic organization including color, structural, and textural differentiation. The upper 70 cm, interpreted as the B horizon of the soil profile, has a blocky structure with argillans on peds. Slickensides are on many peds as well as on fracture planes. Mottles are present throughout, and are comparatively better formed and of redder hue in the lower 30 cm of this horizon. Indistinct root traces and/or burrows are confined to the upper part of the profile. Fracture surfaces or craze planes (Brewer, 1964) in the upper boundary of unit 2 are spaced 10-15 cm apart (fig. 6). These are inclined at angles of 50° to 60° at the top and merge with a master horizontal shear surface, which delineates the base of the upper 65 cm of the profile.

The lower 205 cm, interpreted as the C horizon of the profile, is somewhat lighter in color than the upper part and is characterized by much coarser soil structure. Argillans are thin and discontinuous; slickensides are weakly formed and concentrated along major fracture surfaces. Mottles show a patchy distribution and constitute less than 25 percent of the groundmass. Most mottles are oval to elongate or tubular, and commonly harden to form irregular ferruginous concretions (plinthite) as much as 3 cm in diameter. Unit 2 becomes progressively sandier with depth, and the abundance of mottles and concretions decreases near the base.

Specific indications of *in situ* Cretaceous pedogenic development include: argillans, the pattern of color mottling, position of ironstone concretions, and the vertical pattern of ped organization in the unit. Although mottling is associated with roots, burrows, and groundwater fluctuations, some mottles may result from modern vadose weathering. Iron nodules (plinthites) indicate downward translocation of iron, and the nodules are closely associated with the network of sand-filled fissures.

Unit 2 was extremely cohesive before deposition of the overlying sand. The fissure network was rapidly infilled as the sand was deposited. Fracture planes which are typical of clayey soils, formed as unit 2 was compacted; connections within the fissure network were disrupted as dislocation along craze places occurred.

Since the Cretaceous Period, vadose weathering has tended to obscure the pedogenic features preserved within this section of the Tuscaloosa. Reduction bands along sediment interfaces result from lateral migration of groundwater, a process that

Figure 6. Upper part of the soil profile at Site B showing contact relationship with the overlying channel sand. Cross-cutting relationships between craze planes (open fractures) and discontinuous sand-filled fissures are well displayed.

may still be operating. More pervasive is the oxidation of relict geomorphic surfaces which have been exposed during erosion of the Chattahoochee River Valley. Iron staining and banding to hematite cementation, especially in sand-dominated sequences, constitute the latest stage of diagenesis. In addition, a Quaternary weathering front can be traced from the top of this exposure and down the adjacent slopes.

DISCUSSION

The two Cretaceous paleosols studied in the Chattahoochee River Valley result from fundamental differences in soil forming factors (see Jenny, 1941). Both soils appear to have developed on low-relief land surfaces, but differences in parent material and intensity of weathering produced different types of profiles. Duration of weathering and paleoclimate are also factors in explaining the development of these paleosols.

The lateritic paleosol on crystalline basement is a remnant of an ancient landscape that predates deposition of Tuscaloosa alluvium. This pre-Cenomanian soil is truncated by the unconformity that separates the crystalline rock surface from the Tuscaloosa. Facies of the paleosol are preserved locally, and we suggest that variations in soil development may be related to the local topography of the basement surface and to the amount of truncation.

Modern lateritic soils are found primarily in warm, humid climates and have been studied on several continents including South America, Africa, and Australia; excellent reviews are given in Alexander and Cady (1962), McFarlane (1976), Sirarajasingham and others (1962), and Stephens (1971). There is considerable disagreement about processes and exact climatic conditions of laterite formation; however, most authors believe that these soils result from intense weathering in tropical or sub-tropical climates. Lateritic soils have B horizons characteristically rich in sesquioxides of iron and aluminum. They are usually devoid of organic matter, bases, and primary silicates, and may contain abundant quartz and kaolinite.

Laterites can form on most rock types. Environmental conditions that seem to favor laterite formation include: 1) a relatively flat land surface; 2) a climate with alternating wet and dry seasons; and 3) decomposing organic matter in the profile (Alexander and Cady, 1962). Precise temperature and precipitation requirements for "laterization" are controversial, but high temperature and humidity favor desilicification, and frequent wetting and drying enhance mobilization and subsequent precipitation of sesquioxides (Mohr, and others, 1972; Ollier, 1969).

Sesquioxide enrichment in lateritic soils results from a complex interaction between pedogenic and groundwater processes (Duchaufour, 1977). The distribution of pisolites and mottles in a profile can provide a basic clue to identify the dominant process involved in the genesis of certain lateritic soils. For example, laterites, that are predominantly pedogenic in origin, commonly contain pisolites which increase in frequency towards the base of the profile (Ollier, 1959). In contrast, groundwater laterites generally show the reverse distribution, and pisolites and mottles tend to be less frequent and well defined with depth. A sheet of closely packed pisolites usually occurs at the contact between the soil profile and the underlying saprolite (see McFarlane, 1976, p. 74).

The pisolites and mottles in the lateritic profile at Columbus are concentrated in the well indurated, mottled zone at the base of the B horizon, a condition that suggests groundwater control in development. However, isolated nodules are found throughout the soil profile as well as in the underlying saprolite. This distribution suggests that downward migration of soil solutions as well as groundwater enrichment of sesquioxides were responsible for profile development. A polygenetic origin for the soil is also supported by oxidation, soil structure, and argillans in the upper part of the B horizon. Sesquioxide enrichment by capillary rise may have played some role in profile development (see Holmes, 1914; Woolnough, 1927), but this process is considered to be of secondary importance.

Our evaluation of paleoclimate is based largely on analogy with modern tropical to subtropical climates in which lateritic soils form. Several other lines of geologic data from the Cretaceous support the inference that the Columbus paleosol formed in a warm, humid environment. For example, during the Early Cretaceous, the study area was 5 to 10 degrees latitude closer to the equator than today (Hospers and van Audel, 1968), and mean annual temperatures may have been as much as 15 degrees C warmer. The support for higher temperatures is based on isotopic studies of oceanic Foraminifera and other marine invertebrates, as well as paleoclimatic interpretations of terrestrial bauxite deposits (see reviews by Frakes, 1979; Savin, 1982). In addition, paleobotanical data from numerous locations, including the southeastern United States, indicate that a warm, humid climate prevailed throughout the Early Cretaceous with floras dominated by tropical ferns, gymnosperms, and primitive angiosperms (Berry, 1919; Darrah, 1960; Smiley, 1967).

The genesis of the "Columbus laterite" and its relation to the Coastal Plain unconformity suggest a period of surface stability and weathering prior to truncation by the Tuscaloosa fluvial sediments. The paleosol has a minimum pre-Cenomanian age; its maximum age cannot be determined. It is usually impossible to assign absolute rates of development to pre-Quaternary soils; however, numerous studies of Tertiary laterites and denudation chronologies from Africa and Australia indicate that these soils evolve over thousands, if not millions of years, particularly when associated with landforms of regional extent (e.g., Ruhe, 1956; Hays, 1967). Similarly, the Columbus paleosol most likely represents a prolonged period of soil formation before erosion and deposition of the Tuscaloosa Formation.

In contrast, the alluvial soil profile that we have described within the Tuscaloosa Formation almost certainly developed in a much shorter time. The fine-grained parent material of this profile is previous weathered regolith that was eroded and transported in a vertically aggrading flood plain. The more stable mineral composition and the fine grain size of the alluvium precluded the development of a strongly differentiated profile. Following deposition, the surface of the flood plain stabilized and soil development began. The sand-filled fissures resulted from the infilling of mudcracks that formed during alternating wet and dry conditions. The fissures channelized vadose water and consequently controlled the concentration of plant roots and distribution of mottles and ferruginous nodules. Dominant pedogenic processes of this soil include minimum illuviation of clay and reduction of iron compounds in a poorly drained environment. The argillans may in part post-date the soil, but the mottled profile indicates alternating oxidizing and reducing conditions due to fluctuating water table. The open fracture network is superimposed on the original structure of the profile, and it reflects adjustment of the alluvium to sedimentary loading under the laterally accreting channel deposits. The upper bounding surface of the soil and the fractures now act as pathways for groundwater as weathering continues on the modern land surface.

CONCLUSIONS

Paleosols are a means to interprete paleoclimate and correlate stratigraphic successions. The lateritic soils at Columbus appear to be regional facies of a pre-Cenomanian soil-stratigraphic unit that is widespread in the eastern Gulf Coastal Plain. Possible equivalents occur in northern Mississippi, western Kentucky, southern Illinois, and Arkansas. These well-developed paleosols occupy the same stratigraphic position relative to the regional unconformity and are truncated and buried by later Cretaceous and/or Tertiary sedimentary rocks. Pedogenic variations between localities are controlled by differences in parent material and possibly regional climate, as well as truncation geometry. These paleosols seem to have formed on a well-vegetated hinterland in a humid and tropical to subtropical climate before Late Cretaceous time. The apparent synchroniety of soil development appears to be of intercontinental proportions because similar pre-Late Cretaceous paleosols are reported from several continents.

In contrast, the thin alluvial soil in the Tuscaloosa Formation formed in an actively aggrading fluvial system where flood plain sediments were deposited intermittently and weathered in a somewhat less humid environment. Similar profiles are found in fine-grained facies of the Tuscaloosa and are usually of local extent. These soils represent temporary stability of the flood plain and are difficult to correlate outside the immediate drainage basin. This type of paleosol is important for local stratigraphic and paleoenvironmental studies, but is of limited value in regional basin analysis.

In conclusion, paleosols indicate land surface stability and exposure, and provide valuable evidence to interpret paleoclimate and paleogeography. From pedogenic characteristics, we can make inferences about the nature of tectonic and vegetational stability, and suggest paleoclimatic conditions during soil formation. Because well preserved paleosols are present in much of the Coastal Plain in the southeastern United States, further study will enhance our knowledge about the local Mesozoic and Cenozoic climates and paleoenvironments.

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STRATIGRAPHY OF THE LYNCHBURG GROUP AND SWIFT RUN FORMATION, LATE PROTEROZOIC (730-570 Ma), CENTRAL VIRGINIA

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ABSTRACT

The Lynchburg Group and Swift Run Formation comprise a belt of upper Proterozoic (730-570 Ma) metasedimentary rocks in the eastern Blue Ridge of Virginia. Mapping in two separate areas has clarified stratigraphic relationships within this belt. In the Culpeper area in northern Virginia, the Lynchburg Group is divisible into the Bunker Hill, Monumental Mills (new), Thorofare Mountain (new), Ball Mountain (new) and Charlottesville formations. The lower three formations represent a transition from alluvial outwash plain through delta front and slope into submarine fan deposition, whereas the remaining units were deposited by sediment gravity flows in deep water. In the Rockfish River area 90 km to the southwest, a similar sequence is present, but the Bunker Hill and Monumental Mills formations are absent. In their position is a lenticular unit of glaciogenic pebbly sandstone called the Rockfish Conglomerate. In both areas, the Lynchburg Group is overlain by the Swift Run Formation, a predominantly metasedimentary sequence distinguished by scattered occurrences of volcanic rock, quartzite and marble. Petrographic data from coarse-grained Lynchburg Group sandstones indicate an ensialic source similar to the present-day Blue Ridge basement terrane.

The transition from non-marine to deep marine sedimentation from north to south in the Lynchburg belt may reflect an oblique section through a late Proterozoic rifted margin which trended more east-west than the present Blue Ridge anticlinorium. This basin margin appears to have controlled sedimentation patterns into the early Paleozoic.

INTRODUCTION

Metasedimentary and metavolcanic rocks of late Proterozoic age occur in thick sequences within the Blue Ridge province of the central and southern Appalachians. These rocks record rifting which preceded development of an early Paleozoic passive margin, and they are the oldest evidence that the Appalachian region was beginning to differ tectonically from cratonic North America (Rodgers, 1970, p. 212). Because of polyphase deformation, metamorphism and a lack of age control, stratigraphic relationships within the upper Proterozoic of the Blue Ridge are poorly understood.

The Lynchburg Group is part of the upper Proterozoic sequence in northern and central Virginia. It crops out on the eastern limb of the Blue Ridge anticlinorium, comprising a belt up to 12 km wide of metamorphosed siliciclastic and both mafic and ultramafic rocks. The Lynchburg Group nonconformably overlies both Grenville-aged basement and the younger Robertson River pluton. It is overlain by the Swift Run Formation, a unit originally defined on the western limb of the Blue Ridge anticlinorium and consisting of metasedimentary and metavolcanic rocks. The Swift Run Formation grades upward into metabasalts of the Catoctin Formation. The entire succession has been metamorphosed in the upper greenschist facies and penetratively deformed at least twice. Nevertheless, primary sedimentary structures are remarkably well preserved, and a coherent stratigraphy persists throughout the study area.

Most geologists have mapped the Lynchburg as an undivided formation. The purpose of this report is to present a lithostratigraphic scheme for the Lynchburg Group based on mapping in two areas*: one in the vicinity of Culpeper, Virginia and the other about 80 km to the southwest along the Rockfish River (Fig. 1). Reconnaissance in the intervening areas and as far south as the Tye River suggests that this lithostratigraphy

[1]present address: Exxon Production Company, P. O. Box 2189, Houston, Texas 77001
*Geologic maps of these areas at 1:62,500 are available upon request from the author.

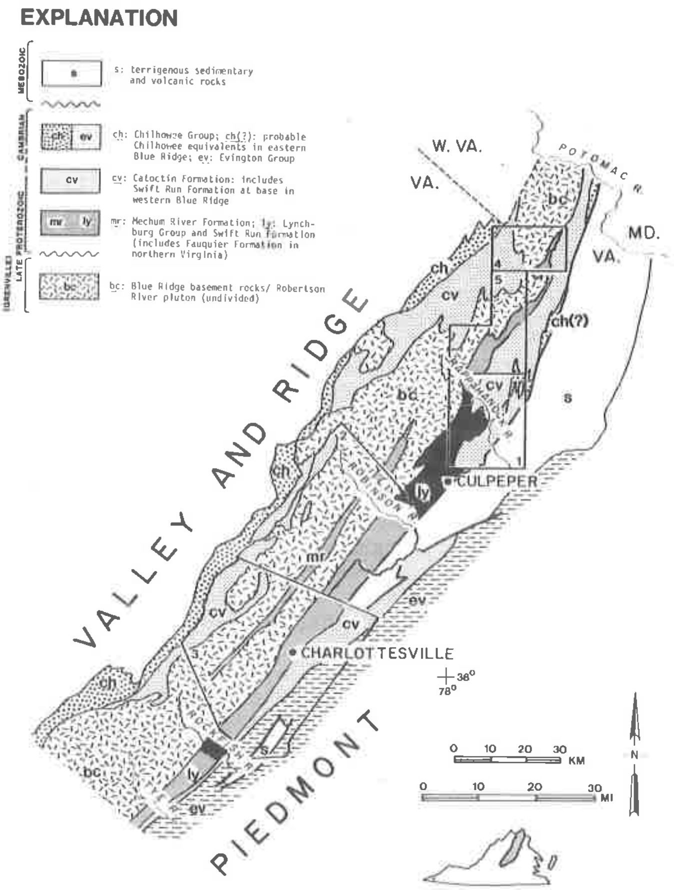

Figure 1: Geologic map of the Blue Ridge anticlinorium in northern Virginia showing study areas and areas of previous work: (1) Furcron, 1939; (2) Allen, 1963; (3) Nelson, 1962; (4) Parker, 1968; (5) Espenshade and Clarke, 1976. Dark stipple in Lynchburg belt indicates study areas.

persists throughout much of northern and central Virginia. Further southwest, towards Lynchburg and beyond, metamorphic grade reaches amphibolite facies and sedimentary fabrics are poorly preserved.

AGE CONSTRAINTS

Age constraints on the Lynchburg Group are poor because of metamorphic overprint and a lack of fossils. However, radiometric and biostratigraphic data suggest an age range of 730-570 Ma for deposition.

In northern and central Virginia, Blue Ridge basement rocks are intruded by a hornblende granitoid called the Robertson River pluton. This pluton yielded a discordant U-Pb zircon age of 730 Ma (Lukert and Banks, 1984). Cobbles and boulders of Robertson River granitoid have been identified in both the Mechums River and Fauquier formations (the northern equivalent of the Lynchburg Group), indicating a maximum age of 730 Ma for deposition. A minimum age of about 570 Ma is inferred from stratigraphic relations on the northwest limb of the Blue Ridge, where the Chilhowee Group overlies the Swift Run-Catoctin sequence. The upper part of the Chilowee contains Lower Cambrian fossils *(Olenellus)*, and probable Chilowee Group equivalents overlie the Catoctin Formation on the eastern limb of the Blue Ridge in northern Virginia.

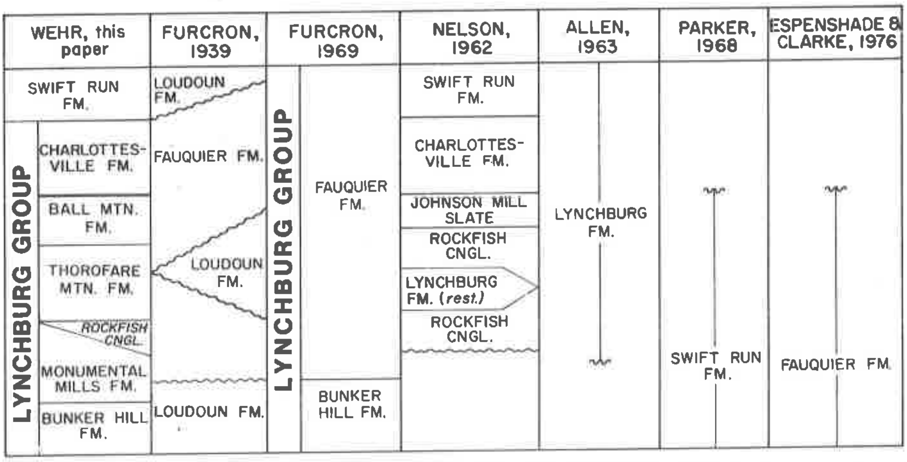

Figure 2: Proposed stratigraphic correlation for pre-Catoctin sedimentary rocks on the eastern limb of the northern Virginia Blue Ridge. Lateral equivalence of units is implied, based upon stratigraphic position, reconnaissance mapping and unit descriptions in the literature. Ranges indicated for Allen (1963), Parker (1968) and Espenshade and Clarke (1976) are based upon regional map patterns.

PREVIOUS WORK

The Lynchburg Formation was named by Jonas (1927) for exposures in the vicinity of Lynchburg, Virginia and included rocks previously called the Lynchburg Gneiss and the Loudoun Formation (Virginia Geological Survey, 1928). It extends southwest to the Virginia-North Carolina line (Virginia Division of Mineral Resources, 1963), where it probably interfingers with the Ashe Formation of Rankin (1970). To the northeast, the Lynchburg trends into the Fauquier Formation of northern Virginia, a probable nonmarine equivalent of the Lynchburg (Espenshade and Clarke, 1976).

Most geologists have mapped the Lynchburg - Fauquier belt as a single formation, although different formation names have been used in different areas along strike (Fig. 2). The first attempt at subdivision was in the Warrenton 15' quadrangle, where Furcron (1939) distinguished the Loudoun and Fauquier formations. At that time, Furcron interpreted most of the coarse-grained sandstones in the Lynchburg belt as Cambrian in age, lying nonconformably upon both the fine-grained Fauquier Formation and metabasalts of the Catoctin Formation. Thus, he correlated them with the Loudoun Formation, the basal sandstone of the Lower Cambrian Chilhowee Group to the north and west. Much later, Furcron (1969) revised his stratigraphy and assigned the coarse-grained basal sandstones of the Lynchburg Group to the Bunker Hill Formation, while retaining the name Fauquier Formation for the remainder.

Southwest along strike in Albemarle County, Nelson (1962) divided the belt into five formations, reserving Lynchburg Formation (restricted) for "fine-grained silty sediments" containing "varved-like layers of graphitic and sericitic schist". He mapped much of the coarse-grained lower Lynchburg as Rockfish Conglomerate, and he correlated the uppermost part of the Lynchburg with the Swift Run Formation, a thin, tuffaceous unit which occurs at the base of the Catoctin Formation on the northwestern limb of the Blue Ridge anticlinorium. Conley (1978, p. 128) supported Nelson's Swift Run correlation, describing distinctive felsic volcanic rocks and arkose present at the top of the Lynchburg in the Charlottesville area.

In northern Virginia, Parker (1968) mapped all of the metasedimentary rocks beneath the Catoctin Formation on both limbs of the Blue Ridge as Swift Run Formation. However, Espenshade and Clarke (1976) disputed Parker's correlation because of the greater thickness and lack of felsic volcanic rocks in "Swift Run" in the eastern Blue Ridge. Moreover, Espenshade and Clark (1976) were unable to find evidence for Furcron's (1969) division of the Lynchburg into formations, and they were not convinced that the rocks in Fauquier County were equivalent to the Lynchburg farther south. Thus, they mapped the entire belt as Fauquier Formation.

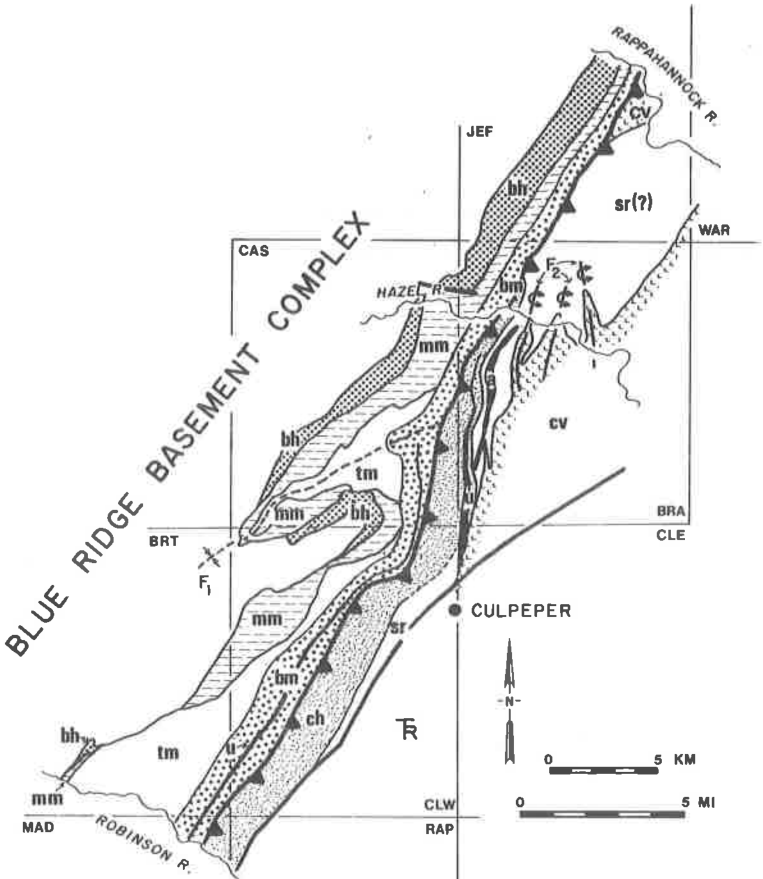

Figure 3: Geologic map of the Culpeper area. bh: Bunker Hill Formation; mm: Monumental Mills Formation; tm: Thorofare Mountain Formation; bm: Ball Mountain Formation; ch: Charlottesville Formation; sr: Swift Run Formation; sr(?): poorly exposed belt of probable Swift Run Formation north of Hazel River; a: amphibolite; u: ultramafic rock; cv: Catoctin Formation greenstone; Tr: Culpeper Triassic basin. Note pinch-out of Bunker Hill and Monumental Mills formations to the south and of Thorofare Mountain Formation to the north, reflecting transition from non-marine (Fauquier Formation of Espenshade and Clarke, 1976) to deep-marine (Lynchburg Formation of Allen, 1963) deposits.

Mapping of the Lynchburg belt through Culpeper County (Wehr, 1983 and this report) indicates that (1) the Lynchburg includes lithostratigraphic units mappable at 1:24,000, and (2) the Lynchburg Formation of Allen (1963) and the Fauquier Formation of Espenshade and Clark (1976) are laterally equivalent, related through a sedimentary facies transition. Therefore, the Lynchburg Group of Furcron (1969) was resurrected and divided into five formations, three of which are named in this report. The name Fauquier Formation was not used within the Lynchburg Group because of conflict with Espenshade and Clarke (1976). Finally, the Swift Run Formation is described in this report but not included within the Lynchburg Group because of the uncertain correlation with Swift Run Formation at its type locality (M. J. Bartholomew, 1971 and written comm., 1983).

STRATIGRAPHY

Lithostratigraphy proposed in this report is based on mapping of the Lynchburg Group and Swift Run Formation in the Culpeper (Fig. 3) and Rockfish River (Fig. 4)

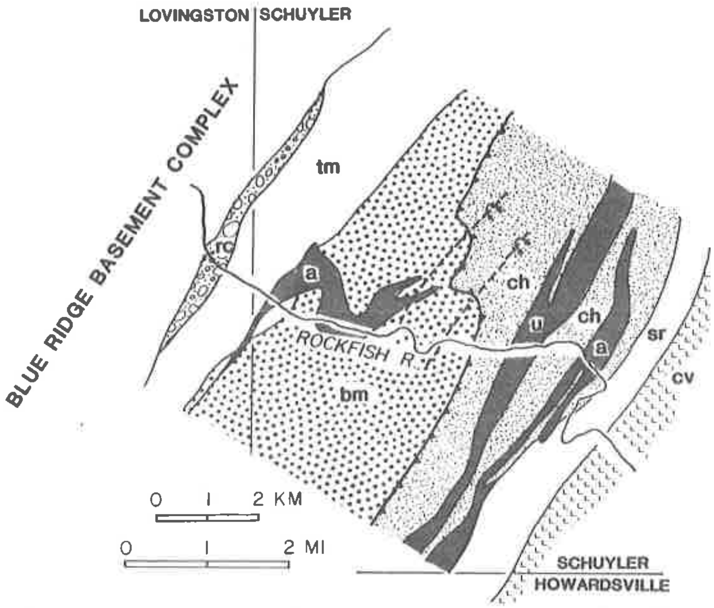

Figure 4: Geologic map of the Rockfish River area. rc: Rockfish Conglomerate; tm: Thorofare Mountain Formation; bm: Ball Mountain Formation; ch: Charlottesville Formation; sr: Swift Run Formation; cv: Catoctin Formation greenstone; a: amphibolite; u: ultramafic rocks.

areas, and on comparison of this data with published accounts of the Lynchburg. Representative localities are listed in the appendix. The stratigraphic scheme is based primarily upon sandstone characteristics, because finer grained lithologies crop out poorly. Sedimentary rock terminology is used in unit descriptions although all of the rocks contain metamorphic fabrics and mineral assemblages.

Stratigraphic relations in the Lynchburg Group have been affected by tectonism, chiefly through juxtaposition of units by folding and faulting. Contacts between formations are not exposed, and original stratigraphic thicknesses are unknown. Abrupt changes in formation thickness along strike probably reflect unmapped faults. Despite these limitations, a coherent stratigraphy exists in both areas and structural data do not indicate large scale transposition of bedding: facing criteria indicate younging consistently to the southeast. The relationship between the lower Lynchburg Group (Bunker Hill-Monumental Mills-Thorofare Mountain-Ball Mountain formations) and the overlying units (Charlottesville-Swift Run formations) is unknown due to probable faulting.

In the Culpeper area, the Lynchburg is divisible into the Bunker Hill, Monumental Mills (new) Thorofare Mountain (new), Ball Mountain (new) and Charlottesville formations (Fig. 5). In the Rockfish River area, the basal Lynchburg Group is the Rockfish Conglomerate, a lenticular unit of pebbly sandstone and conglomerate which grades up into the lower Thorofare Mountain Formation. Both the Bunker Hill and Monumental Mills formations are absent in the Rockfish River area. In both areas, the Lynchburg is overlain by the Swift Run Formation, similar to underlying units but locally containing volcanic and volcaniclastic rocks as well as rare quartzites and marbles.

Bunker Hill Formation

The Bunker Hill Formation was named by Furcron (1969) for exposures near Marshall, Virginia, 20 km north of Culpeper. It consists of light gray, poorly sorted, medium-grained to granule feldspathic arenite with minor pale green siltstone and mudstone. The Bunker Hill Formation is partially equivalent to rocks previously called Loudoun Formation in the Warrenton 15' Quadrangle (Furcron, 1939) and to meta-arkose

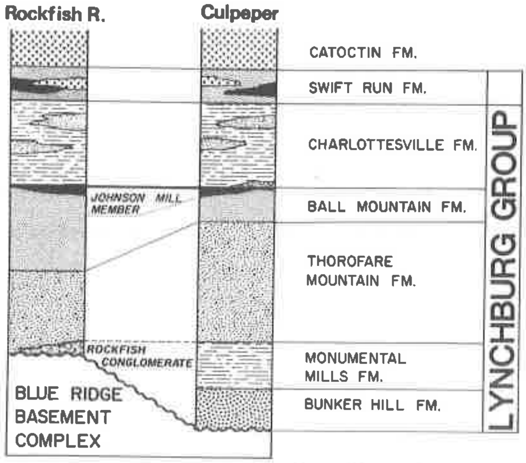

Figure 5: Schematic geologic column of the Lynchburg Group in the Rockfish River and Culpeper areas, showing homotaxis and correlation of Bunker Hill-Monumental Mills formations and Rockfish Conglomerate. Original formation thicknesses are unknown.

described by Espenshade and Clarke (1976, p. 6) and Conley (1978, p. 125).

Within the Culpeper area, the Bunker Hill Formation ranges in apparent thickness from 0-1000 m and is thickest to the north. South of Culpeper, it is present only in thin, isolated lenses at the base of the Lynchburg Group and is absent in the Rockfish River area. Bunker Hill Formation arenite rests directly on crystalline basement rock, generally Robertson River granite, but the contact is obscured by extensive shearing combined with compositional similarities between arkose and granitic basement. The Bunker Hill Formation fines upward from granule sandstone with sparse pebbly lenses near the base to coarse-grained sandstone in the upper portion. Cobble and boulder conglomerate occurs to the north (Parker, 1968; Espenshade and Clarke, 1976) but is absent in the Culpeper area.

The Bunker Hill Formation is distinguished from other coarse-grained sandstones of the Lynchburg Group by its abundant, large-scale cross-stratification and arenitic texture. Typical arenite is composed of 1-3 mm angular grains of clear to smoky quartz and perthitic potassium feldspar in a very fine-grained feldspathic sandstone matrix. Biotite, muscovite, epidote, magnetite and titanite are common as interstitial metamorphic phases.

Facies analysis of the Bunker Hill Formation indicates deposition in a braided alluvial environment, probably an outwash plain adajacent to glaciated plutonic highlands (Wehr, 1983).

Monumental Mills Formation (new)

The Monumental Mills Formation is herein named for exposures in the area of Monumental Mills, Castleton 7.5' quadrangle, Virginia (representative localities in appendix). It is informally divided into two members: (1) a lower sandstone member, dominated by light to medium gray, fine- to medium-grained, well-sorted sandstone and siltstone, and (2) an upper siltstone member, containing dark gray to greenish, thinly laminated siltstone and mudstone.

Although not formally named until this report, part of the Monumental Mills Formation were noted by previous workers. Furcron (1939), in the Warrenton 15' quadrangle described thinly laminated quartz-mica schist in the Fauquier Formation and traced it south into the area of Monumental Mills. Thiesmeyer (1939) interpreted a belt of laminated slates in Fauquier County, Virginia as varved lacustrine deposits. Most recently, Espenshade and Clark (1976) mentioned similar slates near the top of the Fauquier Formation in Fauquier County.

The outcrop belt of the Monumental Mills Formation is 0-1500 meters wide and thins to the south. Contacts are not exposed, but the basal Monumental Mills

Formation appears to coarsen downwards over several hundred meters where it grades into the Bunker Hill Formation. At the top, coarse-grained feldspathic wackes of the Thorofare Mountain Formation sharply overlie the upper siltstone member.

Sandstone member: The sandstone member of the Monumental Mills Formation makes up the bulk of the formation. It consists mainly of thin (1-4 cm) planar sandstone beds separated by biotite-rich, silty partings. Rippled beds and thick beds of massive or faintly stratifed sandstone are common. Synsedimentary folds and liquefaction structures are well developed and may be difficult to distinguish from tectonic folds. Sandstones are dominantly light gray, very fine- to fine-grained and well sorted. They have been entirely recrystallized during metamorphism and are composed of equant quartz and albite, with interstitial muscovite, biotite, epidote, calcite, chlorite, titanite and magnetite. Porphyroblastic garnet and biotite are common, as are euhedral pyrite cubes or limonite pseudomorphs.

The sandstone member has been divided into six facies and interpreted as a delta front developed seaward of the Bunker Hill outwash plain (Wehr, 1983).

Siltstone member: The siltstone member of the Monumental Mills Formation is best exposed in the southern Castleton 7.5' quadrangle where it crops out as several hundred meters of dark, laminated siltstone and mudstone containing rare, graded sandstone beds. To the north it is absent, probably faulted out beneath the Thorofare Mountain Formation. In outcrop, the siltstone member is coarsely laminated, with abundant synsedimentary deformation features, including folds, faults, convolute bedding and erosional-depositional discordances. Petrographically, it consists of completely recrystallized, dark gray to green siltstone composed principally of quartz, albite, biotite, muscovite and chlorite. Accessory phases present include garnet, magnetite, epidote and titanite. Garnet, biotite and magnetite may occur in euhedral porphyroblasts up to 4 mm.

Thiesmeyer (1939) interpreted laminated siltstone along strike to the north as varved lacustrine deposits. However, the suite of synsedimentary deformation structures and their presence in a stratigraphic sequence sandwiched between delta front and submarine fan deposits suggest a slope environment (Wehr, 1983).

Rockfish Conglomerate

The Rockfish Conglomerate is a pebbly, feldspathic sandstone with conglomerate lenses which makes up the lowermost Lynchburg Group in the Rockfish River area. It is absent in the Culpeper area. The Rockfish Conglomerate was named by Nelson (1932), who later included the coarse-grained quartz sandstones of the Ball Mountain Formation as Rockfish Conglomerate (Nelson, 1962). In this study, Rockfish Conglomerate is restricted to coarse-grained feldspathic sandstone, pebbly sandstone and cobble conglomerate up to 500 m thick at the base of the Lynchburg Group. Along the Rockfish River, cobble conglomerate is mostly confined to the lower 30 m, with the bulk of the unit consisting of coarse-grained to pebbly sandstone. The upper 20 m is thin-bedded sandstone and siltstone locally containing outsized clasts. North of the Rockfish River in the Fan Mountains, Rockfish Conglomerate occurs below the Ball Mountain Formation in a belt up to 500 m wide of cross-stratified, coarse-grained to pebbly feldspathic sandstone.

The contact between the Rockfish Conglomerate and underlying basement rocks is a mylonitic zone several tens of meters wide. This zone may have led Bloomer and Werner (1955) to propose deposition on a deeply weathered basement surface. Its upper contact along the Rockfish River is gradational into the lower Thorofare Mountain Formation, marked by the uppermost occurrence of outsized clasts.

Most of the larger clasts in the Rockfish Conglomerate are very coarse-grained leucocratic basement gneiss, but it also contains fragments of granite, biotite gneiss, fine-grained aplite (?) and dark siltstone. In conglomerate outcrops in the Fan Mountains north of the Rockfish area, pebbles and cobbles of bluish quartzite were found. In thin section, Rockfish Conglomerate sandstones consist of detrital grains of feldspar and quartz in a schistose matrix of quartz, plagioclase, mica and magnetite. Magnetic susceptibility is sufficient to produce aeromagnetic highs of amplitudes reaching 800 gammas (Virginia Division of Mineral Resources, open-file rept.)

A glacial origin for the Rockfish Conglomerate was suggested by Cooke (1952), whereas later workers (Brown, 1970; Armstrong, 1977) preferred a deep-water, resedimented interpretation. Facies analysis of outcrops along the Rockfish River has shown that the outsized, extrabasinal clasts are ice-rafted dropstones and indicates that the Rockfish Conglomerate was deposited as subaqueous glacial outwash (Wehr, 1983). Depositional environments inferred from the Rockfish Conglomerate, evidence for glacial influence, and its stratigraphic position beneath Thorofare Mountain Formation turbidites are all consistent with equivalence to the Bunker Hill-Monumental Mills sequence in the Culpeper area.

Quartzite clasts in the Rockfish Conglomerate are unlike any rocks described from the adjacent basement complex. They may be the only evidence of pre-Lynchburg supracrustal rocks in central Virginia.

Thorofare Mountain Formation (new)

The Thorofare Mountain Formation is herein named for exposures on and immediately east of Thorofare Mountain, Brightwood 7.5' quadrangle, Virginia. It consists of medium-grained to pebbly, poorly-sorted feldspathic sandstone with minor conglomerate, siltstone and graphitic mudstone. In the Culpeper area, the outcrop belt ranges from 1-7 km wide. To the southwest, Thorofare Mountain lithologies continue along strike for over 100 km and have been variously described as "Loudoun Formation" (Furcron, 1935), "Rockfish Conglomerate" and "Lynchburg Formation, restricted" (Nelson, 1962) and "gneissic facies" or "metamorphosed graywacke" (Allen, 1963). Sandstones of the Thorofare Mountain Formation are resistant to erosion and form monadnocks and ridges up to 300 m high from Culpeper to the Tye River.

Typically, Thorofare Mountain sandstones are in massive to faintly stratified beds from a few cm to over 8 m thick, averaging around a meter. Grading and scoured bases are abundant, and loading features are well developed. Cross-stratification is very rare. Coarsely laminated siltstone and graphitic mudstone occur interbedded with sandstone and are common as angular rip-up clasts in intraformational conglomerates. Along the Rockfish River, the basal few hundred meters consist of thin-bedded, fine-grained turbidites conformably overlying the Rockfish Conglomerate, whereas in the southern part of the Culpeper area, thick-bedded, fine- to medium-grained sandstone sharply overlies the siltstone member of the Monumental Mills Formation. In general, the Thorofare Mountain Formation becomes coarser and more thick-bedded upwards.

Typical Thorofare Mountain sandstones are very poorly sorted, containing subangular to subrounded 1-4 mm grains of quartz and perthitic potassium feldspar in a recrystallized matrix of quartz, albite, muscovite, biotite and epidote. Titanite, calcite, zircon, apatite and opaque minerals are present as accessories. Quartz-feldspar ratios of the coarse (> 1 mm) fraction average about 3:2.

The Thorofare Mountain Formation in the Culpeper area is divided into four facies and interpreted as coarse-grained submarine fan deposits (Wehr, 1983).

Ball Mountain Formation (new)

The Ball Mountain Formation is here named for exposures on Ball Mountain south of the Rockfish River, Schuyler 7.5' quadrangle, Virginia. It is equivalent to the upper part of the Rockfish Conglomerate and the Johnson Mill Formation of Nelson (1962). The Ball Mountain Formation consists of coarse-grained to pebbly quartz wackes and quartzites interbedded with laminated siltstone and graphitic mudstone. In the Culpeper area, it occupies a belt 1-2 km wide, but south of the Rockfish River where it is thickened by folding, it reaches 4 km in width (Fig. 4). The quartz-rich Ball Mountain Formation is a ridge-former in both areas, and it is unlike other Lynchburg formations in its lithologic constancy and persistance along strike. The upper 100 m is locally a graphitic schist named the Johnson Mill Member (after Nelson, 1962).

In much of the Culpeper area as well as in southern Albemarle County, the Ball Mountain Formation truncates underlying units and is apparently either unconformable upon or in fault contact with them. Elsewhere, it seems to conformably overlie the Thorofare Mountain Formation with a gradational contact, marked by an increase in coarse detrital blue quartz and an increase in grain size. The top of the Ball Mountain Formation is a persistent stratigraphic horizon marked by thick beds of coarse-grained to pebbly, blue quartz wacke which die out over a few meters into a belt of graphitic schist up to 100 m thick (Johnson Mill Member, mapped as Johnson Mill Formation by Nelson, 1962). The Johnson Mill Member is locally strongly sheared and recrystallized, suggesting that it may have acted as a decollement during faulting (Wehr, 1983, p. 33).

Sandstones of the Ball Mountain Formation occur in beds up to 4 m thick interbedded with thin beds of graphitic mudstone. Grading, deep scouring and rip-up clasts of mudstone are characteristic. Distinctive quartz wacke containing angular grains of blue quartz in a gray, pyrite- and biotite-rich sand matrix is common. Locally, the Ball Mountain Formation contains light-colored quartz-muscovite schist, and in the Culpeper area, white, poorly-sorted quartzite is present.

Sedimentary characteristics of the Ball Mountain Formation are similar to the underlying Thorofare Mountain Formation and indicate continued deposition by sediment gravity flows in deep water. The increase in quartz content upward between the Thorofare Mountain and Ball Mountain sandstones may reflect changes in weathering conditions or changes in the composition of the source terrane. The sharp transition upward to graphitic mudstone of the Johnson Mill Member suggests abrupt cessation of clastic influx and basin-wide starvation following Ball Mountain deposition.

Charlottesville Formation

The Charlottesville Formation was named by Nelson (1962), who described it as a fine-grained, massive quartz-biotite gneiss, calcareous in places, with a few beds of sericitized and graphitic schist. It is partly equivalent to rocks described by Furcron (1969) as the Fauquier Formation and the "upper sequence" of Conley (1978, p. 125). In both the Rockfish River and Culpeper areas, fine-grained Charlottesville Formation sandstone are in fault (?) contact with either Johnson Mill graphitic schist or with coarse-grained sandstone of the Ball Mountain Formation.

The Charlottesville Formation weathers deeply to a dark red soil, and exposures are poor. It typically consists of schistose siltstone and mudstone, with isolated outcrops of medium- to coarse-grained, commonly amalgamated sandstone beds. Sandstone beds range from a few centimeters to about a meter thick and are typically massive, although grading, horizontal stratification and complete T(a-e) Bouma sequences are locally preserved. A few beds of pebbly sandstone and one cobble conglomerate outcrop is present in the Culpeper area. The lower 1000 meters in the Rockfish River area is composed of coarsely laminated to very thin-bedded, fine-grained sandstone and siltstone containing prominent biotite porphyroblasts. Similar rocks are found locally near the base of the Charlottesville Formation around Culpeper. Concordant tabular bodies of mafic and ultramafic rock are abundant in the Charlottesville Formation and include serpentinites and talc-tremolite schists.

Typical Charlottesville Formation sandstones are fine- to medium-grained feldspathic wackes, petrographically similar to sandstones of the Monumental Mills Formation and to the finer sandstones of the Thorofare Mountain Formation.

Detailed facies analysis of the Charlottesville Formation was not done due to poor exposure. However, primary textures and structures preserved in the sandstones suggest deposition from turbidity currents in deep water.

Swift Run Formation

The Swift Run Formation was originally defined on the northwestern limb of the Blue Ridge anticlinorium east of Swift Run Gap (Stose and Stose, 1946), where it occurs in lenses as much as 400 m thick of meta-arkose, phyllite and greenstone at the base of the Catoctin Formation (Fig. 1). At its type locality the Swift Run rests directly on basement gneiss. Because of its stratigraphic position between Catoctin Formation and basement, the Swift Run Formation has been interpreted as the thinned western equivalent of the Lynchburg (Stose and Stose, 1946; Bloomer, 1950; Brown, 1970). More recently, the Swift Run Formation was correlated with rocks on the southeastern limb of the Blue Ridge, overlying the Lynchburg Group (Nelson, 1962; Conley, 1978). This correlation is based on lithologic similarities (particularly the presence of arkose and felsic volcanic rocks) as well as its stratigraphic position immediately beneath the Catoctin Formation.

The Swift Run Formation is present in both the Culpeper and Rockfish River areas. In the Rockfish River area, it is divisible into three units: a lower unit 80-360 m wide of coarse-grained feldspathic sandstone, a middle unit 30-240 m wide of greenstone, fine-grained sandstone, graphitic mudstone and rare felsic volcanic rock, and an upper unit 80-460 m wide of coarse-grained, blue quartz sandstone and arkose interbedded with pale green mudstone and a few thin greenstone beds. The total width of the belt is 530-850 m. In the Culpeper area to the south of Culpeper, the Swift Run Formation is a thin belt (0 to 400 m wide) of coarse-grained, feldspathic and blue quartz arenite and slate which is truncated to the east by the Triassic border fault of the Culpeper Basin (Fig. 3). North of Culpeper, the Swift Run interfingers with a broad, complexly folded belt of coarse-grained sandstone, greenstone, siltstone, graphitic schist and pale green muscovite schist. No felsic volcanic rocks were found in the Culpeper area, although greenstone and intrusive rocks are common.

North of Culpeper, exposure is very poor in the Swift Run Formation and structure is complex, so it is mapped as Swift Run Formation (?). The Swift Run Formation (?) contains many of the same rock types as the upper Ball Mountain Formation but occupies a much broader belt. Both feldspathic and quartz sandstone are present, including distinctive quartz arenites composed of rounded grains of blue quartz in a fine-grained quartz matrix. One sample of quartz sandstone from near the top of the Swift Run (?) contains greenstone fragments. Interbedded with coarse-grained sandstone are laminated to very thin-bedded, fine-grained sandstone and mudstone; both black (graphitic) and pale green (muscovitic) mudstones are present.

Many Swift Run (?) sandstones are calcareous, and along the Hazel River, a conglomerate is exposed which contains white to pale blue, tabular marble clasts up to 45 cm long in a coarse-grained sandstone matrix (see appendix). To the north of the study area, thin lenses of marble are present near the base of the Catoctin Formation (Furcron, 1939; Parker, 1968) and may be correlative with this conglomerate.

Thin-bedded, fine-grained greenish sandstone and mudstone containing well-preserved turbidite structures occur in a strike-parallel belt south of the Rappahannock River. Porphyroblasts of biotite and magnetite are well developed in this lithology.

The contact between the Swift Run (?) and Catoctin formations is sharp, but feldspathic sandstones persist in lenses throughout the Catoctin (mapped as Cambrian "Loudoun Formation" by Furcron, 1939).

Where the Swift Run Formation is well exposed (e.g., along the Hazel River), it consists of thick, massive or graded sandstone beds interbedded with coarsely laminated siltstone and mudstone. No cross-stratification or other evidence for shallow-water conditions is present, so it is tentatively interpreted as the deposits of sediment gravity flows in deep water. In contrast, at its type locality near Swift Run Gap, Swift Run sandstone is typically cross-stratified and interpreted as an alluvial deposit (Gathright, 1976, Schwab, in press).

PETROGRAPHY

Detrital framework mineralogy of sandstones has been shown to depend largely upon plate tectonic setting (Dickinson and Suzcek, 1979; Dickinson and Valloni, 1980). In crystalline terranes such as the Blue Ridge, however, metamorphic overprint may radically alter sandstone mineralogy; therefore, petrographic data must be gathered so as to minimize the effects of metamorphism. The degree of metamorphic recrystallization in the Lynchburg Group is largely a function of grain size: finer grained sandstones and mudstones are entirely recrystallized, whereas the coarse detrital fraction of coarse-grained to pebbly sandstones has survived. For this provenance study, therefore, the constituents of the coarse fraction were considered as the most reliable measure of the source area composition.

A total of 52 slabs of very coarse-grained to granule sandstone were point counted from the Culpeper area. All samples were stained for potassium feldspar. Sampling was limited to the Bunker Hill, Thorofare Mountain, Ball Mountain and Swift Run formations because of the lack of coarse sandstone in the Charlottesville and Monumental Mills formations. 500 points were counted per sample on a 0.67 mm grid, and any point falling on a grain less than 1 mm in diameter was tallied as "matrix". Framework constituents present were quartz (mostly monocrystalline), perthite,

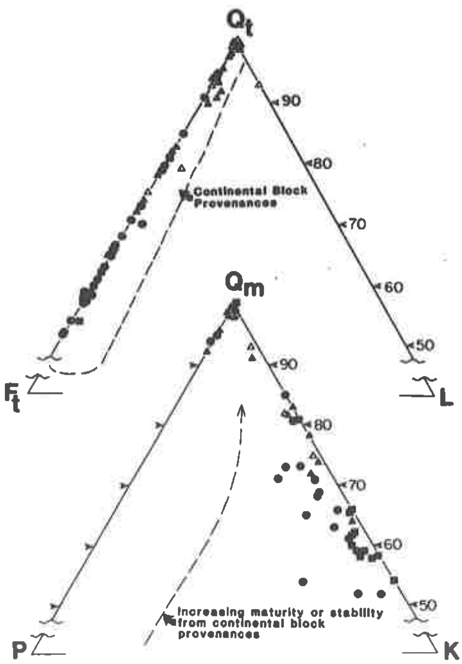

Figure 6: Petrography of Lynchburg Group sandstones. Plots of 52 modes on QFL and QKP diagrams of Dickinson and Suzcek (1979). Closed squares are Bunker Hill Formation, closed circles are Thorofare Mountain Formation, closed triangles are Ball Mountain Formation, and open triangles are Swift Run Formation.

plagioclase (albite), polycrystalline granitoid rock fragments (counted as Qt: stable quartzose rock fragments) and lithic fragments. Most of the lithics were graphitic mudstone or siltstone clasts of intraformational origin, so they were excluded from QFL computations.

The matrix composition of sandstones varies among units: wackes of the Thorofare Mountain and Ball Mountain formations are dark and micaceous, with a matrix of quartz, albite, biotite, muscovite and epidote, and accessory amounts of titanite, apatite, chlorite, calcite, tourmaline and zircon. In contrast, arenites of the Bunker Hill and Swift Run formations have a very fine-grained sandstone matrix, composed primarily of quartz and albite with minor muscovite. Certain samples of Swift Run Formation are calcareous and contain up to 8% modal calcite.

Petrographic data are summarized in Figure 6. Bunker Hill and Thorofare Mountain sandstones are arkosic, containing quartz and perthite in a ratio averaging about 2:1. Plagioclase is generally less than 10%. Ball Mountain and Swift Run formation samples are on average more quartz-rich but include both arkose and quartzite.

Despite metamorphic overprint, Lynchburg Group samples nearly all fall into the continental block provenances field of Dickinson and Suzcek (1979) and suggest a source area similar to the presently exposed Blue Ridge basement terrane. Rutilated blue quartz and perthite are the primary constituents and are abundant in the basement. Lack of detrital plagioclase in samples which contain substantial metamorphic albite and epidote in the matrix suggests that either (1) plagioclase was completely recrystallized during metamorphism, or that (2) detrital plagioclase was concentrated in the fine fraction of the original sediment. There is an almost complete lack of extrabasinal lithics in these rocks, which could reflect both a granitic source terrane as well as the susceptibility of lithic fragments to post-depositional incorporation into the matrix (Shannon, 1978). In a metamorphic sequence such as the Lynchburg, the preserved detrital mineralogy is probably as dependent upon framework grain stability during diagenesis and metamorphism as it is upon original composition.

DISCUSSION

Mapping in the Culpeper area has clarified the relationship between the non-marine Fauquier Formation of Espenshade and Clark (1976) and the deep marine Lynchburg Formation of Allen (1963). It shows that non-marine, delta front and slope facies are confined to the lower two formations of the Lynchburg Group (Bunker Hill and Monumental Mills formations). They make up the bulk of the Lynchburg Group at the northern edge of the study area and are probably equivalent to the Fauquier Formation of Espenshade and Clarke (1976). South of Culpeper, non-marine facies are absent or in thin, discontinuous lenses at the base of the Lynchburg.

This along-strike transition from alluvial to deep-water deposition can be explained in several ways. First, faulting along the basement-Lynchburg contact may be responsible for absence of the Bunker Hill-Monumental Mills sequence to the south; faulting is indicated by both pervasive ductile deformation along the contact and by truncation of Lynchburg stratigraphy. Another possibility is a facies change, by which the periglacial alluvial-delta front system preserved in the Culpeper area passed southward into a more extensive glaciated terrane, where conglomerate and feldspathic sandstone of the Rockfish Conglomerate were deposited directly from ice in a marine environment. A third possibility is the effect of a slight discordance between depositional and structural trends. If the Lynchburg basin trended slightly more east-west than the present strike of the Blue Ridge, then the stratigraphy preserved in the Lynchburg belt represents an oblique section through the basin margin.

A similar change from shallow to deep-water facies is present between the Chilhowee and Evington groups, parts of an early Paleozoic clastic sequence which overlie the Catoctin Formation (Fig. 1). In northern Virginia north of the Culpeper Mesozoic Basin, the Catoctin is overlain by quartz and feldspathic arenites which are correlative with the Chilhowee Group. Where the Blue Ridge stratigraphy emerges from beneath the Culpeper Basin to the south, Chilhowee arenites are replaced by Candler Formation graywackes and pelites of the Evington Group (Brown, 1970, p. 345). Although the transition from Chilhowee to Evington Group is covered, it is in the same area as the non-marine to marine transition in the Lynchburg Group, suggesting that a basin margin which developed in Lynchburg time may have controlled gross facies distribution into the early Paleozoic.

SUMMARY

A coherent stratigraphy exists in the Lynchburg Group through much of northern and central Virginia. In the Culpeper area, the Lynchburg is divided into five formations (Bunker Hill-Monumental Mills-Thorofare Mountain-Ball Mountain-Charlottesville formations). The basal Bunker Hill Formation is an alluvial outwash deposit which passes upwards into the delta front-slope deposits of the Monumental Mills Formation, which themselves pass up into deep-water turbidites which make up the rest of the Lynchburg Group and the overlying Swift Run Formation. To the southwest along the Rockfish River, the Bunker Hill and Monumental Mills formations are absent and the Lynchburg Group is made up almost entirely of deep-water deposits; however, the basal Rockfish Conglomerate is a subaqueous outwash deposit possibly correlative with the alluvial-delta front succession to the north. Coarse fraction petrography indicates a source area similar to the present Blue Ridge basement complex.

Finally, stratigraphic and regional relations suggest that the transition from shallow-water Lynchburg of northern Virginia ("Fauquier Formation" of Espenshade and Clarke, 1976) and deep-water Lynchburg of central Virginia may reflect an oblique section through a late Proterozoic basin margin which trended more east-west than the present Blue Ridge anticlinorium. This basin margin controlled regional facies patterns from late Proterozoic into early Paleozoic time.

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APPENDIX

Locations of Representative Outcrops

Numbers given in parentheses are Universal Transverse Mercator grid coordinates; quardrangle maps are 1:24,000 scale.

BUNKER HILL FORMATION

Jeffersonton, Va Quad: coarse-grained, cross-stratified feldspathic arenite along the Rappahannock River 2.25 km N30°W of the junction of U.S. Route 211 and Va. 229 (244.80 * 4286.72).

Castleton, Va Quad: coarse-grained, cross-stratified feldspathic arenite containing thin beds of green siltstone, in a pasture 350 m N30°W of the junction of State Roads 629 and 632 (758.75 * 4273.55).

MONUMENTAL MILLS FORMATION

Sandstone Member:

Castleton, Va Quad: thin-bedded and rippled sandstone with dewatering structures, in a sandpit near the Hazel River 400 m N50°W of Monumental Mills; also along the south bank of the river across from the sandpit (760.80 * 4276.32).

Castleton, Va Quad: thin-bedded to massive sandstone and siltstone with dewatering structures, in a pasture 800 m S75°E of the junction of State Roads 629 and 632 (759.69 * 4273.09).

Siltstone Member:

Castleton, Va Quad: coarsely laminated siltstone and mudstone, in a pasture 700 m N82°W of the junction of State Road 629 and Muddy Run (755.07 * 4268).

Castleton, Va, Quad: laminated siltstone and mudstone with erosional-depositional discordances, along an unnamed creek 1.2 km N75°E of the junction of State Roads 634 and 716 (752.47 * 4265.65).

ROCKFISH CONGLOMERATE

Lovingston, Va Quad: stratifed cobble conglomerate, under bridge off State Road 617 800 m west of Rockfish Station (697.3 * 4186.7); also, thin-bedded sandstone containing dropstones, in small outcrop on south bank of river 400 m downstream from bridge (697.4 * 4186.3); also thin-bedded, coarse-grained sandstone near house on north bank of Rockfish River directly across river from dropstone-bearing outcrop.

Covesville, Va Quad: cross-stratification pebbly sandstone in draw in Fan Mountains 2.65 km S15°W of junction of Harris Branch and Hammer Branch (706.3 * 4195.8).

THOROFARE MOUNTAIN FORMATION

Brightwood, Va Quad: medium- to coarse-grained feldspathic sandstone, along the Robinson River immediately downstream from a dam 800 m N45°E of the junction of State Roads 634 and 702 (744.45 * 4251.41).

Lovingston, Va Quad: fine- to very coarse-grained feldspathic sandstone and mudstone, in Dutch Creek along State Road 639, 300 m SW of bridge over Rockfish River (698.1 * 4185.1).

Brightwood, Va Quad: coarse-grained to pebbly, massive sandstone, on Thorofare Mountain 700 m N33°E of the junction of State Roads 630 and 632 (748.3 * 4254.6).

BALL MOUNTAIN FORMATION

Schuyler, Va Quad: very coarse-grained quartz sandstone and graphitic mudstone, along unnamed creek immediately downstream of spillway 500 m S10°W of Harris Bridge dam along Rockfish River (700.5 * 4184.35).
Madison Mills, Va Quad: coarse-grained to pebbly, feldspathic and blue quartz sandstone and graphitic mudstone, along south bank of Robinson River 100 m downstream of private residence off State Road 634 (749.8 * 4249.8).
Brandy Station, Va Quad: quartzite, near top of slope on south bank of Hazel River 1500 m N 13°W of Rixeyville (240.34 * 4275.3).

CHARLOTTESVILLE FORMATION

Schuyler, Va Quad: fine- to medium-grained, thin-bedded sandstone, along Rockfish River off State Road 617 600 m E of junction with State Road 800 (703.6 * 4184.7).
Culpeper West, Va Quad: coarse-grained sandstone and conglomerate, in pasture immediately west of abandoned farm 1.1 km S65°E of junction of U.S. Route 29 and State Road 643 (757.7 * 4260.0).

SWIFT RUN FORMATION

Brandy Station, Va Quad: coarse-grained quartz sandstone, mudstone and carbonate clast conglomerate, on south bank of Hazel River 1.1 km S60°E of State Road 229 bridge (242.3 * 4274.7).
Brandy Station, Va Quad: massive quartzite, along State Road 640 immediately W of junction with pipeline (241.35 * 4273.1).
Schuyler, Va Quad: schist and one bed of felsic volcanic rock, along State Road 650 m N15°W of junction with State Road 722 (704.1 * 4183.5).

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THE SOAPSTONE HILL ULTRAMAFIC BODY,

OCONEE COUNTY, SOUTH CAROLINA

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ABSTRACT

The Soapstone Hill ultramafic body is located in the Blue Ridge of northwestern Oconee County, South Carolina. The ultramafite is subovate in plan, tabular or lenticular in cross section, and is exposed over an approximately 8000 square meter area. The ultramafic assemblage includes three rock types designated hornblende amphibolite, tremolite-chlorite schist and soapstone. Progressive, non-pervasive hydrothermal alteration which followed the last deformation/metamorphism locally proceeded to almost complete steatitization of the rock, producing soapstone. The progressive removal of hornblende and replacement by tremolite-actinolite, chlorite and talc was accompanied by systematic increases in MgO, Ni, and Pb and systematic decreases in Al_2O_3, CaO, Na_2O, TiO_2, Cr, Zr and V. Abundant tourmaline (dravite), chlorite (as surface coatings), and euhedral pyrite are conspicuous alteration minerals.

We believe that the ultramafic body was originally pyroxenite which was derived from oceanic crust offboard the North American continent and was emplaced during pre- or synmetamorphic thrusting. The hydrothermally altered Soapstone Hill ultramafic body appears to be similar to the altered ultramafic rocks at Schuyler, Virginia described by Hess (1933).

INTRODUCTION

In recent years, Appalachian ultramafic bodies have received increasing attention, not only because of their unique mineralogy and petrology, but also because of their potential tectonic significance. Misra and Keller (1978) reviewed the distribution, petrology, origin and emplacement of ultramafic bodies in the southern Appalachians in an effort to establish guidelines to explain the occurrence of the ultramafic bodies and to facilitate their placement in tectonic models. The present study consists of the description of another Appalachian ultramafic body briefly described by Sloan (1908) and an explanation of its mineralogy and alteration.

GEOLOGIC SETTING AND PREVIOUS WORK

The Soapstone Hill ultramafic body lies within the Tamassee 7.5-minute quadrangle and is located 8.29 km N 54°W of the junction of Oconee County Road 172 and South Carolina Highway 11 at Tamassee in northwestern Oconee County, South Carolina (Figure 1). Soapstone Hill is in the Blue Ridge geologic belt and should be included in the well-defined but discontinuous western chain of ultramafic bodies described by Misra and Keller (1978).

Roper and Dunn (1970) produced a geologic map of the Tamassee quadrangle and, although a few small ultramafic bodies were located, the Soapstone Hill deposit was not recognized. Its exposure area was included within a unit designated fine-grained, gray mica gneiss, and it is near the contact of the fine-grained, gray mica gneiss with another unit which was designated mica schist and gneiss. The country rock surrounding the Soapstone Hill ultramafite has been metamorphosed to the amphibolite grade (Roper and Dunn, 1970).

Reconaissance studies of Oconee County geology and mineral deposits have been done by Tuomey (1848), Lieber (1859), Sloan (1908) and Cazeau (1967). Sloan (1908) and Gazdik (1981) wrote very brief descriptions of the Soapstone Hill ultramafic body. The only detailed examination of an ultramafic body in the Blue Ridge of South Carolina was done by Hatcher (1970), who mapped and described the Long Creek soapstone

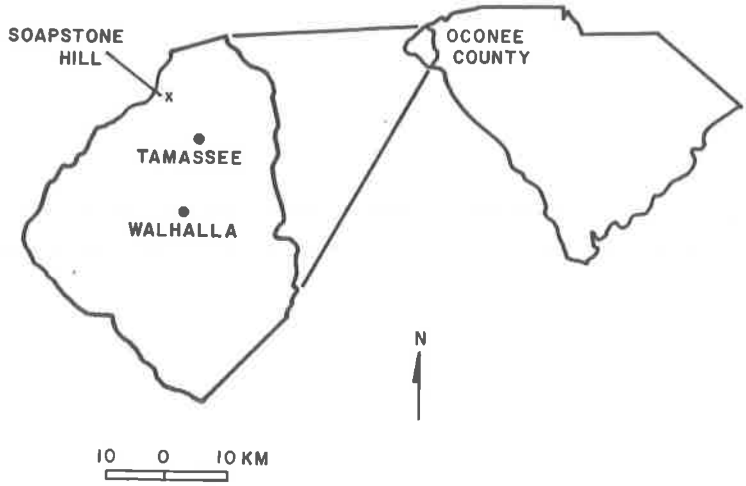

Figure 1. Location map of Soapstone Hill ultramafic body.

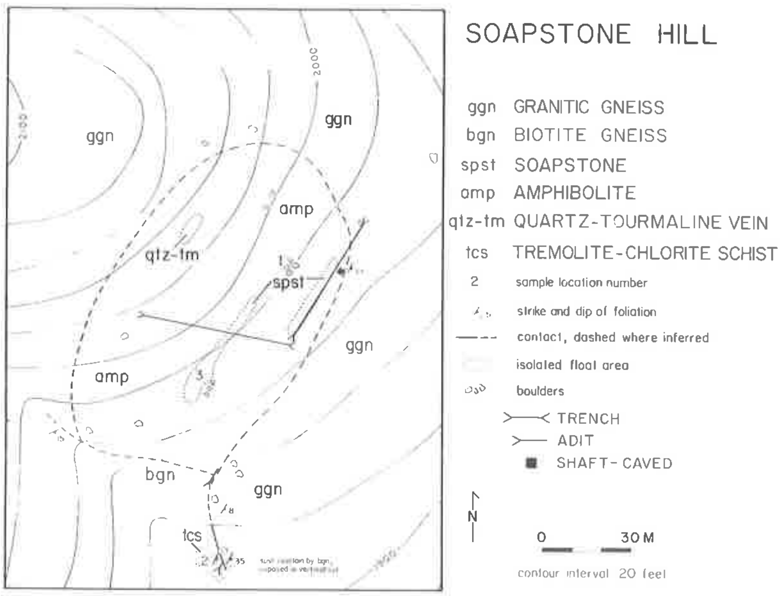

Figure 2. Geologic map of Soapstone Hill ultramafic body.

deposit along Camp Branch, approximately 4 km west of the Long Creek community in western Oconee County.

DESCRIPTION OF THE SOAPSTONE HILL ULTRAMAFIC BODY

Field Relations

The Soapstone Hill ultramafic body is tabular or lenticular in cross section and subovate in plan, with maximum plan dimensions of 125 m and 70 m (Figures 2 and 3). The body is elongated NE-SW, parallel to the local and regional structural trend. The enclosing rocks dip approximately 15-35^{o} SE. The body was mapped on the basis of float, boulder exposures and a very limited number of outcrops. Virtually all of the outcrops are located in the drainage gully on the southern end of the ultramafic body.

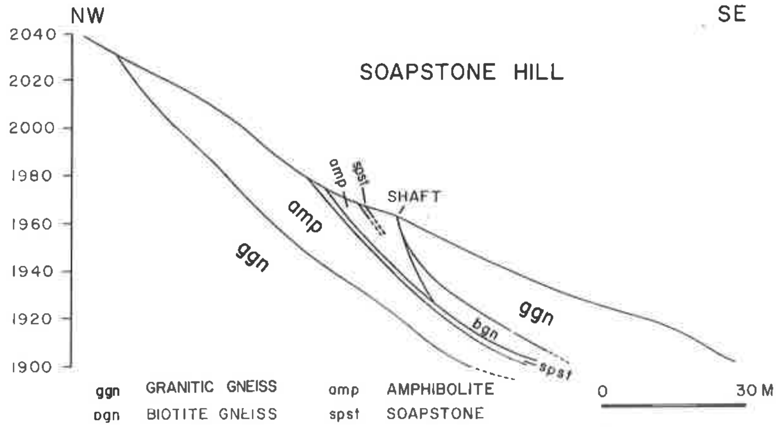

Figure 3. Idealized cross section of Soapstone Hill ultramafic body.

Weathering is deep and has most affected those ultramafic rocks which are the least hydrothermally altered.

Three rock types comprise the Soapstone Hill ultramafic assemblage. From the least to the most altered they are hornblende amphibolite, tremolite-chlorite schist and soapstone. The hornblende amphibolite (not amphibolite *sensu strictu* because it contains no plagioclase) is by far the most common of these rock types. The tremolite-chlorite schist is the least common of the three and has been identified only in float blocks from the gully on the southern end of the ultramafic body (Figure 2). Soapstone crops out in this same gully and is common in the central part of the deposit, particularly along the NE-SW trending trench (Figure 2).

The ultramafic body is surrounded on all sides by granitic and biotite gneiss (Figure 2). Minor quartzite and quartz-mica schist are also country rocks. Where observable, the contact between the ultramafic body and the country rock is very sharp.

As with many of the altered ultramafic bodies in the area, local residents "mined" the Soapstone Hill body for soapstone. There is uncertainty as to whether soapstone was the only commodity removed from the site. Although sawed slabs of soapstone are found at the deposit, the extensiveness of the workings preclude the possibility that soapstone was the only mineral resource obtained from Soapstone Hill. In addition, no soapstone was observed in the adit walls. There are no production records for this deposit. The major workings at the site apparently post-date Sloan's visit, because Sloan (1908) described only an old pit from which soapstone apparently was mined.

Petrology and Geochemistry

Hornblende amphibolite, the main rock type of the Soapstone Hill body, is moderately to weakly foliated, yet in weathered boulders, it appears to be massive. Olive to dark green when fresh, it weathers to a greenish-brown or orange-brown color. The rock is composed of coarse (2-3 mm) relict green hornblende, smaller euhedral crystals of tremolite-actinolite, randomly distributed knots of chlorite, minor orthopyroxene (enstatite), anthophyllite and abundant opaques (Table 1). Many of the opaques are concentrated within the relict hornblende crystals (Figure 4) and are believed to be chromian magnetite on the basis of the high chromium content of the rock and its weak magnetism.

The tremolite-chlorite schist is dark greenish-gray and has moderate foliation. Thin-section examination shows that the schist has nematoblastic to lepidoblastic texture. It is composed of chlorite, tremolite, orthopyroxene (enstatite), talc, opaques, carbonate, clinozoisite, corundum and trace apatite (Table 1). Euhedral, fresh pyrite cubes, up to 3.5 mm across, are common in this rock (Figure 5).

The soapstone (talc schist) is off-white to light gray in color, has moderate foliation and a felted, lepidoblastic texture. The rock is 90 percent talc and 10 percent

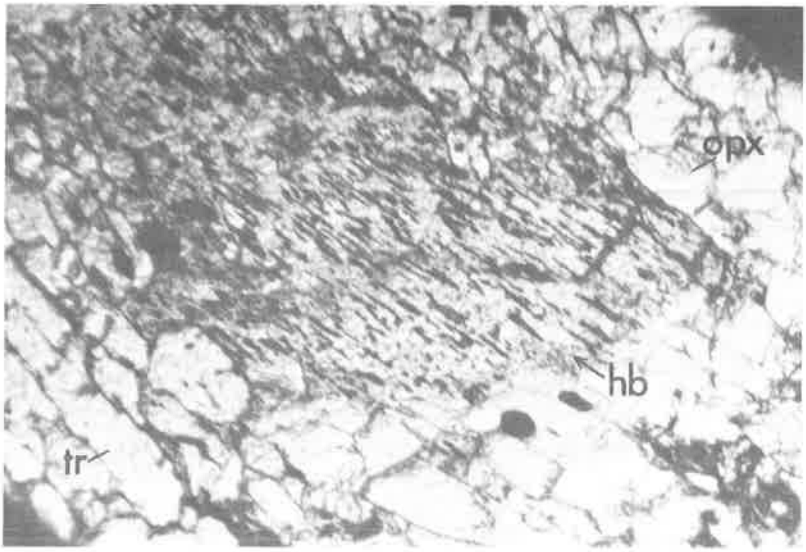

Figure 4. Photomicrograph of hornblende amphibolite (plane-polarized light, width of field is approximately 1.5 mm). hb--hornblende, tr--tremolite-actinolite, opx--enstatite. Note in hornblende the abundant opaque exsolution lamellae, probably chromian magnetite.

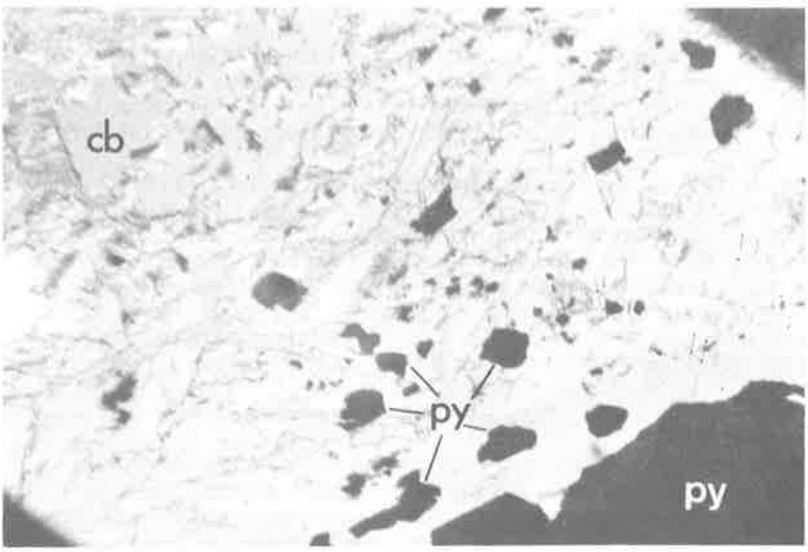

Figure 5. Photomicrograph of tremolite-chlorite schist (plane-polarized light, width of field is approximately 1.5 mm). Minerals include carbonate (cb), pyrite (py) and intergrown chlorite and tremolite (unlabeled).

chlorite, opaques and serpentine (Table 1). Most of the opaques are fresh pyrite cubes and goethite pseudomorphs after pyrite.

The hornblende amphibolite to tremolite-chlorite schist to soapstone alteration process is characterized by a not unexpected systematic increase in MgO and by systematic decreases in Al_2O_3, CaO, Na_2O and TiO_2 (Table 2). Note that the schist to soapstone alteration is accompanied by a marked increase in SiO_2 as chlorite and tremolite are converted to talc (Deer and others, 1966). Low Al_2O_3 values require us to assume that the chlorite in these rocks is alumina-poor, probably due to substitution

Table 1. Estimated visual modes of Soapstone Hill ultramafic rocks (in percent).

	Hornblende Amphibolite	Tremolite-Chlorite Schist	Soapstone
hornblende	40-50	-	-
tremolite-actinolite	30-45	35	-
chlorite	5-10	35	7
orthopyroxene	5-10	7	-
opaques/pyrite	10-15	4	2
talc	-	6	90
carbonate	-	3	-
clinozoisite	-	2	-
corundum	-	2	-
serpentine	-	-	1
apatite	-	tr	-
anthophyllite	0-5	-	-

Table 2. Major-oxide analyses of Soapstone Hill ultramafic rocks (in weight percent).

	1	2	3
SiO_2	49.39	44.74	58.98
Al_2O_3	6.67	5.86	1.05
Fe_2O_3*	10.92	12.70	5.33
CaO	11.10	8.17	0.05
MgO	17.79	21.33	26.28
Na_2O	0.53	0.15	0.06
K_2O	0.27	0.08	0.10
TiO_2	0.68	0.40	0.03
MnO	0.19	0.23	0.05
P_2O_5	0.05	0.03	0.10
BaO	0.01	0.00	0.01
SrO	0.00	0.00	0.00
ZrO_2	0.00	0.00	0.00
LOI	1.31	4.59	1.98
Total	98.91	98.28	94.02

1 - hornblende amphibolite
2 - tremolite-chlorite schist
3 - soapstone

* total iron as Fe_2O_3

Sample numbers correspond to location numbers (Figure 2).

Analyst - A. Debnam, Technical Service Laboratories, Mississauga, Ontario

in the Al site by Si and further, Mg and/or Fe for Al. The systematic reduction in CaO and Na_2O is due to the removal of amphibole and subsequent replacement by chlorite and finally talc. Some trace metals also show systematic increases and decreases; Ni and Pb were progressively enriched in the alteration process and Cr, Zr and V were progressively depleted (Table 3).

In part, the geochemical data from the Soapstone Hill ultramafite do not correspond with the data of Scotford and Williams (1983). They found that metasomatism of ultramafic rocks in North Carolina and Virginia caused MgO and Ni depletion and Al_2O_3 and CaO addition. We have found, however, that the progressive

Table 3. Trace-metal analyses of Soapstone Hill ultramafic rocks (in parts per million).

	1	2	3
Cr	4190	2780	2070
Zr	34	22	11
Cu	85	265	3
Ni	610	1080	1790
Pb	23	27	29
Zn	30	105	52
V	300	188	3
Sr	<1	2	<1
Co	80	99	80
Mo	13	<1	10
Cd	6	10	7
Be	2	2	<1

1 - hornblende amphibolite
2 - tremolite-chlorite schist
3 - soapstone

Sample numbers correspond to location numbers (Figure 2).

Analyst - A. Debnam, Technical Service Laboratories, Mississauga, Ontario

Table 4. Analysis of tourmaline specimen from Soapstone Hill.

Major oxides (in weight percent)		Number of ions on the basis of 31 (O, OH, F)	
SiO_2	35.96	Si	6.0813
Al_2O_3	31.41	B	2.5325
Fe_2O_3	5.89*	Al	6.2622
MgO	9.54	Fe	.7499
Na_2O	1.41	Mg	2.4045
CaO	1.09	Ti	.0244
MnO	0.03	Mn	.0411
TiO_2	0.19	Na	.4614
K_2O	0.18	Ca	.1972
BaO	0.01	K	.0386
SrO	0.03	Sr	.0030
ZrO_2	0.03	Zr	.0021
B_2O_3	8.86	OH	3.2154
LOI	2.85 (assumed to be H_2O+)		
Total	97.48		

* total iron as Fe_2O_3

Trace-metal analyses (in parts per million): Cr, 2610; Zr, 270; Cu, 160; Ni, 280; Pb, 130; Zn, 20; V, 76; Sr, 270; Co, 36; Mo, <1, Cd, <1; Be, 2.

Analysts - A. Debnam and P.E. Burgener, Technical Service Laboratories, Mississauga, Ontario

steatitization at Soapstone Hill produced MgO and Ni enrichment and Al_2O_3 and CaO depletion. This suggests that any chemical exchange between the Soapstone Hill ultramafite and the country rock was minimal; therefore, the alteration resulted from the introduction of hydrothermal fluids rather than from metasomatism.

Other Alteration Minerals

The hydrothermal activity at Soapstone Hill also produced tourmaline, euhedral pyrite cubes and chlorite (as surface coatings). Hot, CO_2-rich fluids probably transported boron and metals into the Soapstone Hill area and converted both tremolite and chlorite to talc. Tourmaline is exceptionally abundant in altered talc-rich rocks along the NE-SW trending trench (Figure 2) and is common as a surface coating. A single tourmaline specimen from the altered ultramafite was analyzed and determined to be dravite, with Na Ca and Mg Fe (Table 4). Compositionally similar tourmaline was described by Holgate (1977) from Radnorshire, Wales and it was associated with amphibolized gabbro. Holgate's tourmaline had SiO_2/Al_2O_3 = 1.136 and MgO/total iron = 1.436 as compared to Soapstone Hill tourmaline which has SiO_2/Al_2O_3 = 1.145 and MgO/total iron = 1.620. Only the CaO content of the Radnorshire tourmaline was markedly different from the Soapstone Hill specimen and Holgate (1977) admitted that the calcium content was unusually high. The tourmaline, in both instances, is part of the alteration mineral suite in altered mafic-ultramafic rock.

Very low strontium in the ultramafite (Table 3) and appreciable strontium in the tourmaline (Table 4) indicate an external source for the strontium—probably hydrothermal fluids derived from the upper continental crust. Conversely, the high chromium and nickel in the tourmaline (Table 4) were probably derived locally from the ultramafite.

Power's (1968) research on chemical variations in tourmalines from southwest England also supports a hydrothermal origin for the tourmaline present in the Soapstone Hill body. Power (1968) reported that hydrothermal tourmaline has a distinctively high strontium content (mean value 173 ppm) as compared to tourmaline from other environments (mean value 34 ppm).

Although it would be expected that tourmaline associated with such a magnesium-rich assemblage would be dravite, it should be pointed out that dravite is commonly associated with hydrothermal, ore-bearing solutions. Dravite is found worldwide in close proximity to major mining districts and is believed to be genetically related to many base- and precious-metal deposits (Slack, 1982).

GEOLOGIC HISTORY AND DISCUSSION

Misra and Keller (1978) divided "alpine-type" ultramafic complexes into two sub-types on the basis of their geologic setting. The first sub-type, ophiolites (or ophiolitic bodies), may occur as allochthonous sheet-like bodies or as chaotic blocks in melange terranes, both of which are related to thrusting. The second sub-type includes tectonic and diapiric intrusives. We believe that the Soapstone Hill ultramafic body should be included in the first sub-type and that it represents a slice of dismembered ophiolite. The small size of the body precludes the possibility of a complete ophiolite stratigraphy being present.

The ultramafic body exposed at Soapstone Hill was probably derived from oceanic crust offboard the North American continent and was tectonically emplaced by thrusting during the early or middle Paleozoic before or during the last deformation/metamorphic event. The very sharp contacts and the petrological nature of the ultramafic body as compared to the country rock suggest that the body is exotic. Hatcher and others (1981) believed that several nearby ultramafic bodies in the southern Blue Ridge which have a similar geologic setting as the Soapstone Hill body were emplaced by pre- to synmetamorphic thrusts and were later deformed and metamorphosed.

We believe that the Soapstone Hill ultramafic body was pyroxenite at the time of emplacement. Through regional metamorphism, foliation was imparted to the ultramafic mass. Most of the primary pyroxenes were converted to amphibole, present in the previously descibed hornblende amphibolite as relict green hornblende. Chlorite and tremolite-actinolite formed in response to post-metamorphic hydrothermal alteration which locally proceeded to an almost complete steatitization of the ultramafic body. Quartz-tourmaline veins (Figure 2), characterized by euhedral black tourmaline crystals in coarse-grained crystalline quartz, and tabular granitic pegmatite veins, are believed to represent the alteration fluids at Soapstone Hill. The alteration system which acted on the ultramafic body at Soapstone Hill effectually lowered the metamorphic grade of the rock.

The rock types and progressive, non-pervasive alteration observed at Soapstone Hill appear to be similar to those described by Hess (1933). Hess (1933) termed it "hydrothermal metamorphism of an ultrabasic intrusion" at Schuyler, Virginia, wherein rocks of his "picrite suite" were progressively altered from actinolite amphibolite to chlorite greenstone and finally to steatite. We conclude, as did Hess (1933), that aqueous solutions facilitated alteration and resulted in the formation of "a successive series of facies in response to continuously decreasing temperature conditions". At Schuyler, Virginia and at Soapstone Hill, the alteration was non-pervasive, "so that various stages of the alteration remain" (Hess, 1933).

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