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Abstract

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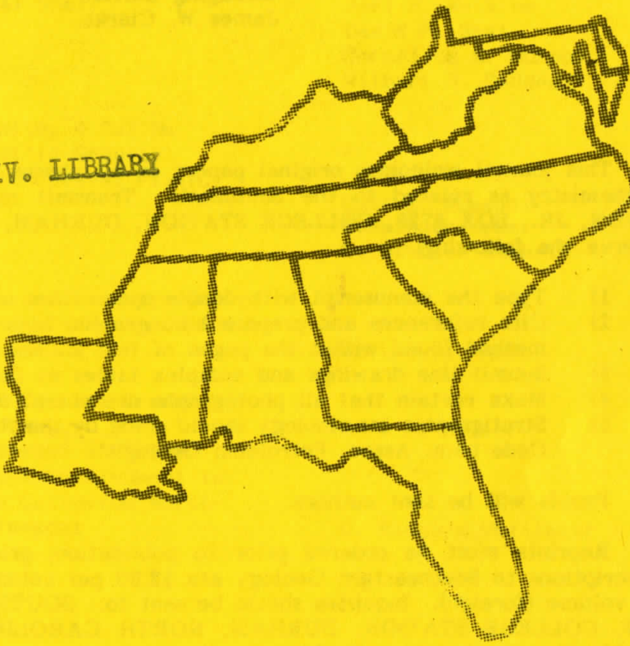

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SOUTHEASTERN GEOLOGY

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AN INVESTIGATION OF THE ALBEMARLE SOUND GRAVITY ANOMALY, NORTHEASTERN NORTH CAROLINA, SOUTHEASTERN VIRGINIA, AND ADJACENT CONTINENTAL SHELF

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ABSTRACT

The lowest Bouguer anomaly values of the eastern continental margin (-46 mgals) are in northeastern North Carolina, southeastern Virginia and the adjacent shelf. We have named this low the Albemarle Sound Gravity Anomaly. We collected sufficient gravity data to fill gaps and improve coverage in areas of low station density, then analyzed the data using maximum depth calculations and two-dimensional models. We then attempted to determine which of two possible models—a sediment-filled basin or a low-density igneous intrusion—best explains the observed data. We conclude that only the low-density intrusion model (i.e., a large batholith) is compatible with the observed gravity, basement depths and petrology from wells.

Radiometric dates from wells in the anomaly area lead us to suggest that the pluton was emplaced during the late Precambrian (585 - 610 m.y. ago). This period correlated with that of emplacement of other plutons in the Piedmont of North and South Carolina and with the period of the Pan-African orogeny (500 - 650 m.y. ago), a widespread orogenic event in western Africa. Hence, we suggest that the inferred batholith was emplaced as part of the Pan-African episode 300-400 m.y. prior to formation of the Atlantic rift.

INTRODUCTION

The lowest Bouguer anomaly (minimum value -46 mgals) on the east coast of the United States is located in northeastern North Carolina, southeastern Virginia and the adjacent continental self (Society of Exploration Geophysicists, 1982). Relief of the anomaly is slightly greater because of a zone of positive anomalies paralleling the low on its western flank. The steepest gradients outlining the gravity low coincide roughly with the -25 mgal contour and encompass an area of approximately 14,000 sq km (Figure 1). The lowest anomaly values lie near the eastern edge of Albemarle Sound, North Carolina, and adjacent areas. For this reason we have named the low the **Albemarle Sound Gravity Anomaly** (ASGA).

Elsewhere on the east coast of the United States, gravity minima often occur in conjunction with deep basins. Mattick and others (1974) presented evidence correlating the thick accumulation of sediments in the Baltimore Canyon Trough with a regional gravity low. To the north a concavity in the Bouguer anomaly over Georges Bank may result from a thick sediment accumulation (Schultz and Grover, 1974).

Other gravity minima associated with sedimentary basins are well known throughout the eastern United States. Triassic basin rocks crop out in the Piedmont and the axes of these basins are parallel to the coastline throughout much of the Atlantic region. Mann and Zablocki (1961) reported that the gravity anomaly associated with the Deep River-Wadesboro basin in

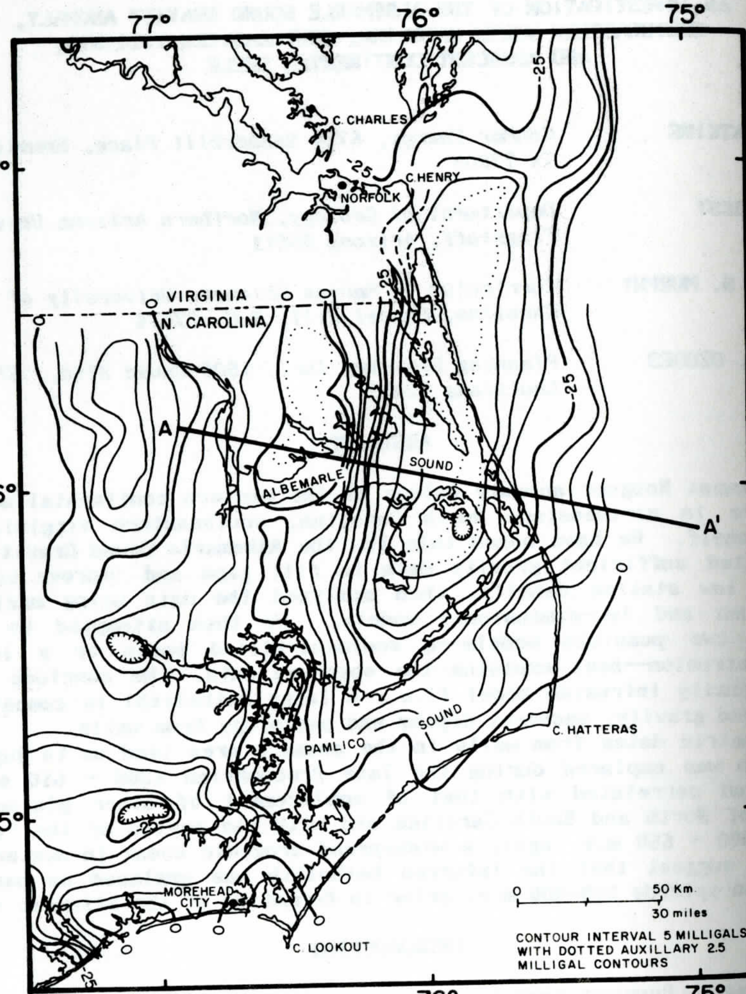

Figure 1: Simple Bouguer anomaly map of northeastern North Carolina, southeastern Virginia, and adjacent continental shelf. A - A' is the line along which the subsurface model was constructed.

North Carolina had an associated gravity low having a maximum value of 13 mgals. The Deep River-Wadesboro basin, however, has been uplifted and exposed to erosion. Restoration of eroded sediments and enclosing igneous and metamorphic rocks would probably increase the magnitude of the anomalies by significant amounts.

Several Triassic basins lie beneath the onlap of post-Triassic sediments. A basin beneath Florence, South Carolina, has been known since a 19th century well penetrated it (McCarthy, 1936). Bonini and Woollard (1960) identified two basins as a result of seismic refraction studies in the Carolinas. Marine and Siple (1974) delineated a Triassic basin buried beneath almost 350 m of Coastal Plain sediments along the Savannah River in South Carolina and Georgia. Talwani and others (1974, 1975) suggested that the source of the large -35 mgal anomaly near Georgetwon, South Carolina, was a deep buried basin in the pre-Cretaceous basement. In studies made along the Coastal Plain and offshore areas of South Carolina and Georgia, gravity lows have been attributed to both low-density igneous rocks and sedimentary basins (Popenoe and Zietz, 1977; Grim and others, 1980). Hence, a basin (either Triassic or post-Triassic in age) filled with sediments having densities less than those of surrounding basement rocks is a plausible origin of the ASGA.

The anomaly could also result from a low-density intrusion. Granitic intrusions are abundant in the North Carolina Piedmont (Butler and Ragland, 1969; Sinha and Zietz, 1982). A large granitic intrusion of batholithic dimensions trends northeast-southwest near Raleigh (Butler and Ragland, 1969). This intrusion has a -35 mgal low associated with it. Smaller granitic plutons have been gravimetrically surveyed in other parts of North Carolina (Mann, 1962). The Farrington pluton in Chatham County is about 7 km in diameter and has a -4 mgal anomaly associated with it. An unnamed pluton, about 10 km in diameter in Wayne County, has a -20 mgal low, and a pluton about 30 km in diameter in Robeson County produces a -30 mgal anomaly.

Daniels and Zietz (1978) defined a granitic zone in the eastern part of North Carolina. This is characterized by the presence of granitic and gneissic basement as seen in well data.

In a large scale study of possible sources of magnetic anomalies lying off the east coast of the United States, Taylor and others (1968) cited studies that attributed the anomaly source lying to the south of the ASGA to igneous intrusions. In addition in a profile transecting the continental shelf off North Carolina Taylor and others (1968) hypothesized an igneous source which was about 30 km wide and located 5 km below the surface (having infinite depth extent). They thought the felsic basement intrusive body was emplaced along the pre-Paleozoic continental-oceanic boundary.

In studies examining geothermal potential for the southeastern United States, Costain (1979) stated "granitic rocks...occur locally in the basement rocks beneath the sediments of the Atlantic Coastal Plain." His reported work did not encounter any such material because their wells were shallow (less than 300 m). He also suspected that the closed gravity low near Portsmouth, Virginia, was due to a granitic source of unnamed age.

We have investigated the origin of the ASGA and considered possible sources of the gravity low. We collected gravity data in areas of relatively low data density and collaborated with the Defense Mapping Agency in the collection of data from sounds and bays of North Carolina. Combining our data, DMA data, and data from the literature, we attempted to determine which of the two models--sediment-filled basin or a low-density igneous intrusion--best explains the observed data.

PREVIOUS INVESTIGATIONS

Previous investigations pertinent to our study include reports of well logs, rock density determinations, structural syntheses, seismic refraction investigations and other geophysical investigations. These are summarized below.

Well Data: Crystalline rocks which underlie the sedimentary rocks of the area range in age from Precambrian to pre-Cretaceous. At least 19 wells have reached basement (Figure 2; Table 1). Well names, depths-to-basement, rock types and ages were compiled from Drake and others (1959), Maher (1965), Denison and others (1967), Maher and Applin (1971), Daniels and Zietz (1978) and Gleason (1981). The latter two works provide excellent summaries of well data and seismic depth measurements in eastern North Carolina. The basement contour map of Gleason (1981) serves as control for gravity modeling.

Geophysical Data

Woollard (1940), in discussing the gravity low which parallels the coast of southeastern Virginia, stated that only about one-third of the low can be

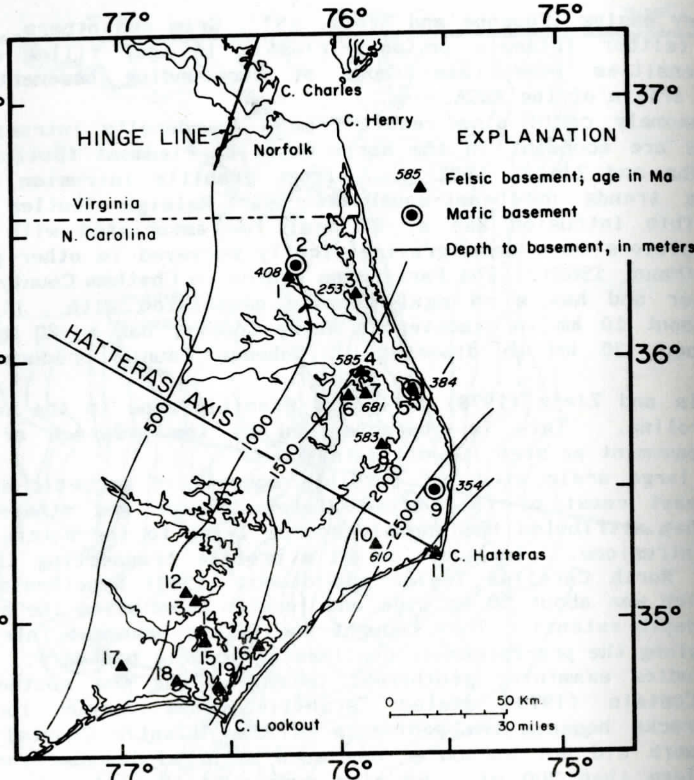

Figure 2: Location of wells to basement in eastern North Carolina (see Table 1).

accounted for by the increasing thickness of sediment. He suggested two possible explanations. The first was that the low is a result of "density transition in the crust at the continental border, traceable to changes in the relative amount of acidic and basic crustal material composing the geologic column." The second explanation was that the low may be related to a downwarping of the crust. "Changes in amounts of acidic and basic crustal material" is approximately equivalent to a low-density intrusive model.

Using combined gravity, magnetic, and seismic data on the North Carolina Coastal Plain, Skeels (1950) concluded that most gravity anomalies "must be due to variations in density in the pre-Cretaceous basement." His magnetic anomalies "show the same north-south trend as do most of the gravity features." For magnetic and gravimetric data he concluded that "the basement is of a complex nature, made up of rocks of contrasting properties, with most of the contacts striking approximately north and south. Lithologically and structurally it seems to be part of the same geological province as the Piedmont." Refraction profiles gave a depth to basement at Cape Hatteras of 2500 m (8,200 ft.), while the Hatteras well found basement at 3000 m (9,853 ft.). Skeels attributed this discrepancy to a high velocity limestone-dolomite layer which the well encountered. He felt this layer would make any such refraction measurements in the eastern part of this area erroneous.

Worzel and Shurbet (1955) presented geologic models for gravity data which extend eastward along two sections from Cape Henry, Virginia, and Cape Hatteras, North Carolina. At both of these positions the upper mantle is shown to be at a depth of greater than 30 km.

Bonini and Woollard (1960) used seismic refraction measurements to study

Table 1: Selected Wells Reaching Basement In Northeastern and Southeastern North Carolina.

Well Number (Figure 2)	Well Name	Depth to Top of Basement (m)	Oldest Rocks
1	Weyerhaeuser #1	860	408 ± 40 Ma altered rhyolite porphyry
2	Foreman #1	1494	Triassic diabase
3	Twiford #1	1380	253 ± 5 Ma biotite schist
4	Mobil #1	1575	585 ± 40 Ma granite gneiss
5	Marshall Collins #1	1911	384 ± 8 Ma carbonatized amphibolite
6	West Va. Pulp and Paper #1	1562	Quartz diorite
7	Westvaco #2A	1762	681 ± 25 Ma brecciated granite
8	Westvaco #1A	1879	583 ± 46 Ma granite
9	Mobil #2	2549	354 ± 7 Ma biotite diabase
10	Mobil #3	2201	610 ± 60 Ma gneissic granite
11	Hatteras Light (Esso #1)	3042	Pre-Mesozoic granite
12	N. C. Pulpwood #1	1115	Pre-Cretaceous granite
13	Atlas Plywood #1	1041	Pre-Cretaceous granite
14	D. H. Phillips #1	1178	Pre-Cretaceous granite
15	Guy Carraway #1	1236	Pre-Cretaceous granite
16	Bayland Corporation #1	1698	Pre-Cretaceous granite
17	Bryan #1	734	Pre-Cretaceous granite
18	Laughton #1	1230	Pre-Cretaceous granite
19	Huntley-Davis #1	1511	Pre-Cretaceous granite

Locations are shown on Figure 2. Data compiled from Denison and others (1967), Daniels and Zietz (1978), Gleason (1981), and Russell and others (1981).

the Coastal Plain of North and South Carolina. We used their data in conjunction with well data to construct the basement contours in Figure 2 (see also Gleason 1981).

The ASGA was originally delineated by Mann (1962) in the course of his gravity mapping of North Carolina. He concluded that the -45 mgal (his data) gravity trough required a mass deficiency which could be caused either by a very deep sedimentary basin or by some other feature far deeper than the underlying Cretaceous sediments.

Bassinger and others (1970) constructed a free air gravity contour map covering northeastern North Carolina, southeastern Virginia and part of the shelf area to the east. An approximate 10 nautical mile spacing was used between stations, with stations at sea being occupied using a sea-floor gravimeter. Their map shows a large, closed direction, and is roughly rectangular in shape. Their data were incorporated with other Department of Defense data in the construction of our gravity map of the region.

Investigation of the aeromagnetic and regional gravity data along the Coastal Plain of North Carolina showed that a covered Triassic basin averaging 35 km in width could exist (Won and others, 1979). Modeling resulted in configuration consisting of a series of horst and graben structures with Piedmont-type basement rock underlying the basin.

Geological Structure

Richards (1954), using samples from well no. 3 (Figure 2), suggested that a Triassic basin underlies the northeastern North Carolina Coastal Plain. His generalized log of the well placed Triassic material from 1073 to 1676 m (3,520 to 5,500 ft.) below the surface. Diabase is prominent in this well, suggesting the presence of one or more diabase dikes. At a depth of 1957 m (6,421 ft.), quartz sand and fragments of quartzite were encountered. These were taken to indicate basement. Other research in this immediate area has suggested the presence of basement rocks at 975-1097 m (3,200-3,600 ft.)

(Spangler, 1950; Skeels, 1950). Richards (1954) concluded that the discrepancy in basement depths, 1957 m (6,421 ft.) vs. 975-1097 m (3,200-3,600 ft.), can best be explained by the presence of a Triassic basin.

Ferenczi (1960) concluded that two transverse structural feature, the Great Carolina Ridge (Cape Fear Arch, outside the area of this investigation) and the Hatteras Axis (Figure 2), influenced the transgression and regression of the seas in different geologic times. The Hatteras Axis is said to represent a line where all formations are at their greatest depths. He also concluded that "the basement rock beneath the sedimentary cover has the character of a peneplained block mountain rather that that of a folded mountain chain. Structural conditions...and gravity and other anomalies...can perhaps be more easily interpreted by referring to them as blocks in the basement rock mass differing in position."

Brown and others (1962, using paleontologic, lithologic, and electric-log data from water test wells, presented evidence of the edge of a marine Cretaceous basin in northeastern North Carolina that extends into southeastern Virginia, but they did not specify the location of this basin. The top of the marine zone lies between 32 and 217 m (106 and 713 ft.) below land surface. The zone contains marine fossils and is "late Cretaceous or early late Cretaceous in age."

Rodgers (1970), from a synthesis of available literature and well log data, suggested the presence of a NNE-SSW trending line marking the western edge of a probable Triassic basin beneath the coastal plain of southeastern Virginia. This hinge line is shown in Figure 2.

Studies by the U.S. Geological Survey (Mattick and others, 1974) showed that large deep basins are located beneath the outer continental shelf of the eastern United States. The Baltimore Canyon Trough is thought to contain about 15,200 m (50,000 ft.) of post-Triassic sediments and an estimated 3,000 m (10,000 ft.) of Triassic and pre-Triassic sediments (Sawyer and others, 1982). The Georges Bank Trough is believed to contain 3,620 m (25,000 ft.) of Triassic and Post Triassic sediment (Sawyer and others, 1982) As shown in a later section of this report, thicknesses of these magnitudes are several times larger than the thicknesses required to explain the **ASGA**.

THE GRAVITY MAP

After examining the gravity data available in the literature, we selected areas of low data density and established an approximate 6-km grid of gravity stations in the region of the ASGA. Details of the survey, principal facts and base station data are available elsewhere (Murphy, 1972). Gravity data used in this investigation came from three main sources: 1) Mann's (1962) data from northeastern North Carolina, 2) the Department of Defense's data from a gravity survey of Albemarle and Pamlico Sounds, and 3) the Murphy (1972) thesis data file.

Tie loops show the DOD base stations averaged 0.6 mgal lower relative to Murphy's base station. Tie loops with Mann's (1962) base stations revealed no significant differences between his and Murphy's base station values. Tie loops were not made with Bassinger and others' (1970) bases. Bouguer reductions were based on an assumed crustal density of 2.67 g/cm^3. Bouguer anomalies were used on land and free air anomalies at sea.

DISCUSSION OF MODELS

Maximum Depth Calculation

A rough estimate of the depth to the top of a gravitating body can be obtained from a formula developed by Bott and Smith (1959). They showed that

$$h \leq \frac{0.86 g_{max}}{dg/dx_{max}}$$

where g_{max} is the maximum intensity (positive or negative) of the gravity field, dg/dx_{max} is the maximum value of the gradient (taken along A-A' of Figure 1), and h is the depth to the top of the body. Using g_{max} = -45 and dg/dx_{max} = 30 mgals/7 km yields

$$h \leq 8.83 \text{ km}.$$

The equality sign is attained when the anomaly-producing body becomes a single straight line of mass parallel to the surface of the earth. For the ASGA the depth h is probably much less than its calculated maximum value.

Sediment-filled Basins

Density contrasts of sediments in our hypothetical basins and the enclosed basement rocks were estimated as follows. The *in situ* density (ρ) of a porous, water-saturated rock is

$$\rho = \rho_0 + \phi\rho_w,$$

where ρ_0 equals the average dry density, ϕ equals the average porosity, and ρ_w is the density of salt water. Spangler's (1950) data indicated that the average dry density of selected rocks from the Cape Hatteras well is 2.02 g/cm^3. Thus

$$\rho = 2.02 + (0.27)(1.02)$$

$$\text{so} \quad \rho = 2.29 \text{ g/cm}^3.$$

This value was increased to 2.4 g/cm^3 for modeling purposes in order to compensate for some compaction (see below).

As previously mentioned considerable evidence indicates that sediment-filled basins showing gravity minima underlie part of the Atlantic Coastal Plain. Therefore, we considered two different types of sedimentary models as sources of gravity low. The first is a sediment-filled basin (Figure 3a). Basement rocks consist of igneous and metamorphic rocks whose depths have been determined from interpolations of published well log and seismic data. The sediment-filled polygon in Figure 3a explains the observed gravity data rather well. The density contrast of $\Delta\rho$ = -0.4 g/cm^3 represents the difference in densities between compacted sediments (ρ = 2.4 g/cm^3) and basement rocks (ρ = 2.8 g/cm^3) under the northeastern North Carolina Coastal Plain. This polygon has a maximum depth to its top of three km along the easternmost edge, well within the maximum depth of 9 km.

An additional sedimentary model (Figure 3b) represents a salt basin. This model was proposed because of evaporites encountered by the Hatteras Light Esso #1 well (Table 1, well #11). The density contrast of $\Delta\rho$ = -0.6 g/mc^3 assumes the salt has a density of 2.2 g/cm^3. The computed gravity for the salt model also shows close agreement with the observed gravity data.

In order to be geologically viable, the basin models must be consistent with well log and geophysical data. On Figure 2 are contours of basement depths based on Table 1 data, Bonini and Woollard's (1960) seismic data, and data compiled by Gleason (1981). Eight of the well and seismic data points lie within the -25 mgal contour. The sedimentary basin models (Figure 3a and 3b) require basement to rise eastward in order to account for the increasing values of gravity on the eastern edge of the low. Sufficient well log and seismic data are available to enable us to state that basement continues to increase in depth in an easterly direction (Gleason, 1981).

Low-density Igneous Intrusion

The gravity low could be caused by density differences in basement rocks. Evidence that several types of basement rocks of different densities exist is found in the analysis of cores (Denison and others, 1967; Russell and others, 1981) and in the magnetic work of Zietz (1970) (see also Woolard,

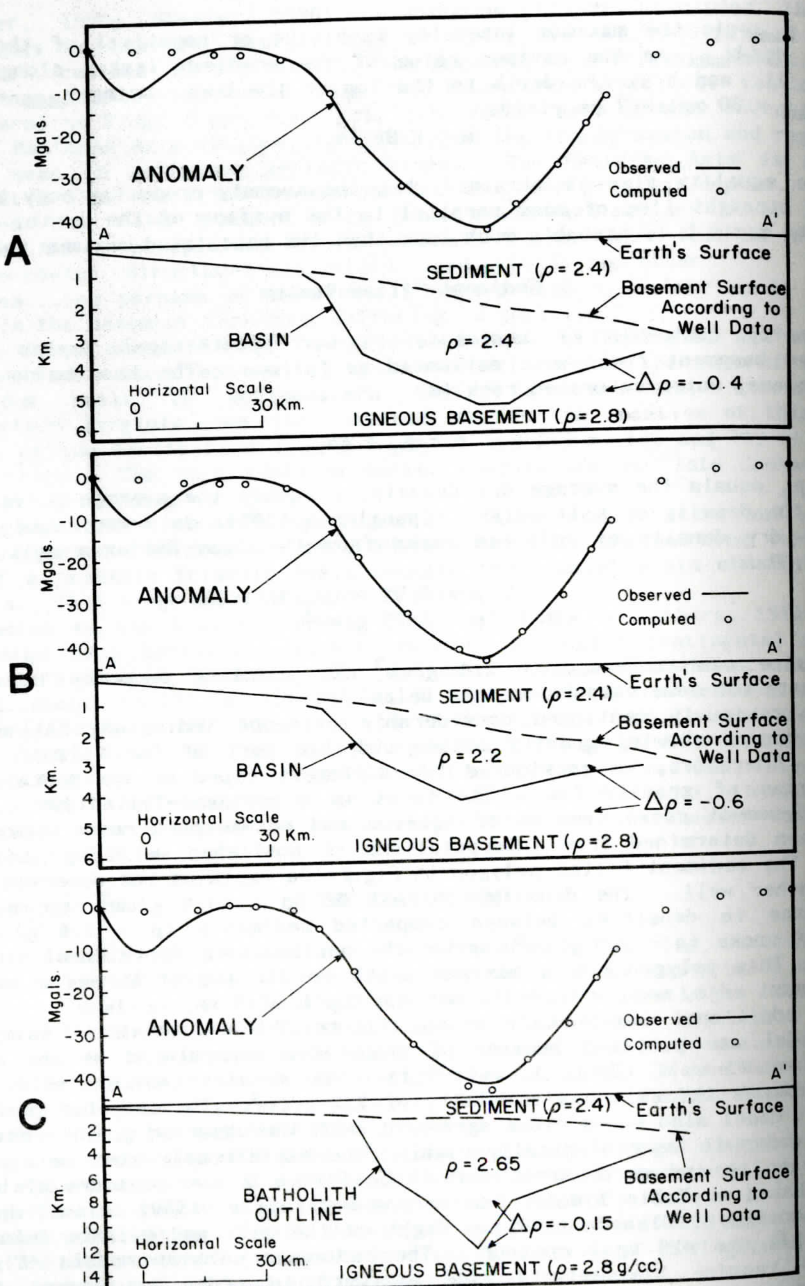

Figure 3: Models generated for cross section A - A'. a. Sediment-filled basin; b. salt (evaporite) basin; c. igneous intrusion. Densities and

1940; Skeels, 1950; Ferenczi, 1960; Mann, 1962; Taylor and others, 1968; Popenoe and Zietz, 1977).

The third model (Figure 3c) is based on the assumption that the gravity low results from a low-density intrusion into the upper crust. This model assumes an intrusion with a density of 2.65 g/cm^3 which is enclosed by upper crustal material having a density of 2.80 g/cm^3. This density contrast of -0.15 g/cm^3 seems to be the one likely without extensive injection of basalt or other dense rock into the crust. Smaller density contrasts would require larger and much deeper intrusions.

We considered the possibility that the gravity low is due to a Triassic basin largely covered by basalt. If this were true, drillers striking the basalt might stop drilling and report depth to basalt as depth to basement. This situation seems unlikely in the area of the ASGA. Although wells 2, 3, 5 and 9 bottomed in basaltic or similar rocks, dates from all wells are older than early Triassic. The pre-Triassic ages suggest that the rocks are basement rocks and not younger flows hiding deeper basins. Our best fitting intrusive model represents a zone of low-density basement rocks running paralled to the north-south axis of the low. The density contrast of -0.15 g/cm^3 represents a reasonable difference in density between the basement rocks in the area. This zone would have to be flanked on either side by higher density rocks in order to give the increasing values of gravity shown in Figure 1.

As noted in Table 1, wells 1, 4, and 10 cored gneissic granite or rhyolite porphyry, a lower density material than was cored in the remaining wells. Rocks from wells 3, 5, and 9 are also younger than rocks from wells 1, 4, and 10. This implies at least two different origins for the basement rocks in this area. The highest density rocks (wells 5 and 9) are so located (Figure 2) that they may be the source of the increase in the Bouguer anomaly values on the east side of the low. The low-density rocks appear to be the cause of the gravity low.

Further evidence for the intrusive origin of the source of the gravity low is that Russell and others (1981) reported drill samples near Stumpy Point, NC (about 35 km south of line A-A') which were granitic in composition. Their Rb-Sr isochron age was 583 ± 46 Ma. In addition relatively high heat flow values exist in northeastern North Carolina and southeastern Virginia (Kron and Heiken, 1980). Contours show the highest geothermal gradients (about 35 to 45° C/km) lie just to the east of the axis of the igneous mass in Figure 3c.

TWO-DIMENSIONAL MODELS

Using gravity anomaly values shown in Figure 1, we attempted to discriminate between the two plausible causes of the anomaly discussed previously. As shown below, it was possible to eliminate a crustal downwarp on the basis of maximum depth calculations. Comparison of two dimensional model data with well data and seismic data provided the basis for discrimination between sedimentary basin and intrusion models.

To expedite our model calculations, we first estimated the planar mass deficiency along profile A-A' (Figure 1). For convenience, the "plane" was taken to have constant thickness of 1 cm. Following Grant and West (1965)

$$\int\int \Delta g(x,y)\, dx\, dy = 2\pi GM$$

where g(x,y) is in gals, G is the universal constant of gravitation (6.67 x 10^{-8} dynes-cm^2/gm), and M is the mass. Taking x to be along the traverse A-A', dy to be 1 cm, and using a Simpson's rule integration, we obtained

$$M = -6.6 \times 10^{11} \text{ gm.}$$

This value has no significance except to provide a basis for the estimation of the size of the body necessary to cause the anomaly.

In order to be valid, two-dimensional modeling of gravity anomalies requires that the length of the anomaly-producing body is much greater than the width. A length-width ratio of 10 is usually considered adequate. The length-width ratio of the ASGA is significantly less than 10. It is also true, however, that a thin near-surface body can be satisfactorily modeled if the length-depth ratio is sufficiently great. If we assume that a length-depth ratio of 10 is adequate (i.e., the distance from the cross-section to the end of the anomaly is five or more times greater than the bottom of the anomaly-producing body), profile A-A' is suitable for two-dimensional modeling of bodies at shallow to moderate depths. The distance from A-A' to the -25 mgal contour south of A-A' is about 60 km. The maximum depth to which modeled two-dimensional bodies can extend without introducing significant errors is about one-fifth of this distance or 12 km. The maximum depth of our modeled bodies is slightly less than 12 km. Therefore, errors due to the assumed two-dimensional shape of the anomaly-producing body are probably not significant.

AGE OF THE INFERRED BATHOLITH

Fullagar (1971) noted that Piedmont plutonic rock ages are in three age groups: (1) 595-520 Ma, (2) 415-385 Ma and (3) approximately 300 Ma. Representatives of the intermediate age group have not been found in the eastern Piedmont. In the eight wells in the anomaly region with dated basement rocks (Table 1), three dates (585, 583 and 610 Ma) probably correlate with Fullagar's 595-520 Ma episode. The 610 Ma age seems to old, but it does carry a large probable error (±60 Ma). The one date of 681 Ma, is undocumented except as cited by Daniels and Zietz (1978). Two dates, 384 and 408 Ma, could coincide with Fullagar's 415-385 Ma group, but the dates could be meaningless due to sample alteration. The two remaining dates, 253 Ma and 354 Ma, may reflect post-crystallization alteration of the rocks (Denison and others, 1967; P.D. Fullagar, pers. comm., 1983).

Which, if any, of these age groups might represent the age of the inferred batholith? The most likely age of the batholith may be late Precambrian-early Paleozoic as indicated by dates from wells 4 (585 Ma), 8 (583 Ma) and 10 (610 Ma). Wells 4, 7, 8, and 10 lie near the axis of the gravity low. All yielded granitic basement rocks required to explain the nomaly. The three wells from which radiometric dates were obtained yielded ages which were essentially consistent with each other and with a period of orogenic activity as outlined below. The ages obtained from these wells are also thought to be formation ages rather that metamorphic ages. Thus, we interpret the anomaly as deriving from batholith emplaced during the late Precambrian or early Paleozoic.

IMPLICATIONS

The possibility of a late Precambrian-early Paleozoic batholith close to the continental margin has many implications. During the time of emplacement of the batholith, the North American Piedmont and Africa were probably joined (Watkins, 1971). To the west and the northwest (in terms of present day directions) the Piedmont was tectonically active, but northwest of the Piedmont and Appalachian orogeny 100 m.y. later (see Rodgers, 1970, p. 216). To the east and southeast, however, the African continent was in the midst of the Pan-African orogeny.

Radiometric ages (Hurley and Rand, 1968) show that west Africa experienced three major orogenies within the past three billion years. The youngest of these, the Pan-African orogenic episode (Kennedy, 1965), lasted from at least 650 to 500 Ma ago. Today, Pan-African age rocks extend as a continouous belt around African cratonic blocks (Hurley and Rand, 1968).

Along the west coast of Africa where Cape Hatteras probably lay before rifting, Pan-African age rocks crop out on the eastern margin of the Tertiary onlap in Mauritania, Mali, and Senegal. Farther to the south in Guinea and Sierra Leone, Pan-African age rocks crop out in abundance. Rodgers (1970) pointed out stratigraphic similarities between Paleozoic rocks of the Fonta-Djallon Plateau in Guinea and Paleozoic rocks in southern Georgia and northern Florida. Rodgers noted that although the African sequence is

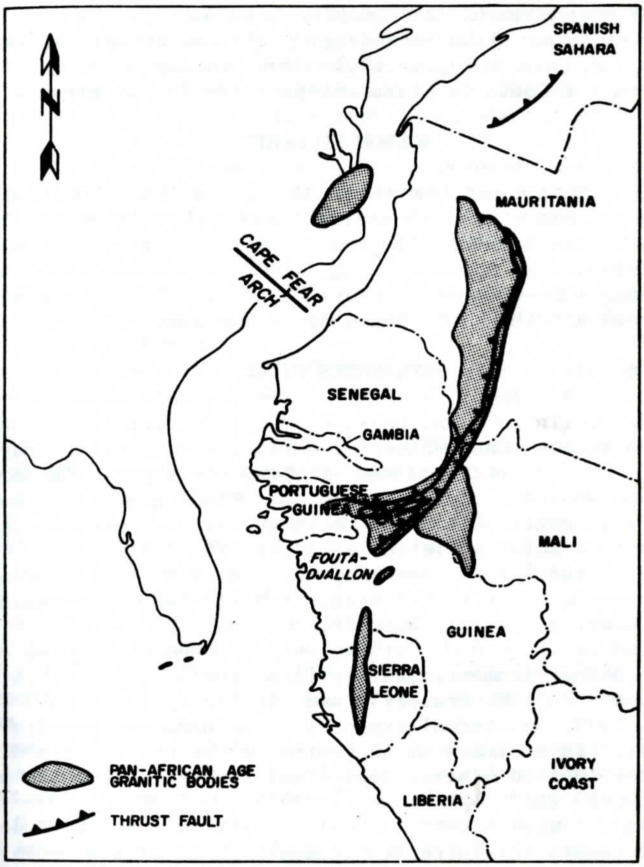

Figure 4: Relationship of the granitic basement in eastern North Carolina with rocks of similar composition and age in western Africa. Figure is not to exact scale.

thinner, both sequences consist of lower Paleozoic sandstones overlain by a relatively thick sequence of black shales which he thought to originally be part of a single unit.

Although North America-Europe and South America-Africa pre-drift fits are well established, details of North America-Africa fits remain controversial. In our interpretation, placing the southern Georgia-northern Florida Paleozoic rocks adjacent to the Fouta-Djallon Plateau rocks, and Cape Hatteras opposite the Senegal basin and western Mauritania appears most consistent the the gravity data (Figure 4).

We have examined the literature in an attempt to find evidence in Africa of a northern extension of either the inferred batholith or the gravity anomaly, but data are too sparse along the continental margin of west Africa. Resolution of the question of the northern extent, if any, of the anomaly and batholith must wait.

SUMMARY

Analysis of gravity data from northeastern North Carolina, southeastern Virginia and the adjacent shelf indicates that a low-density batholith is responsible for the large gravity low in the area. Radiometric ages from wells in the region indicate that the granitic intrusion was probably emplaced in late Precambrian or early Paleozoic time. We suggest that since at the time of emplacement the anomaly area was probably contiguous with parts of the African mainland experiencing the Pan-African orogenic activity, the emplacement of this and possibly other plutons of similar ages in the Piedmont of North and South Carolina was part of the Pan-African orogeny.

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THE PORTSMOUTH GRANITE, A 263 Ma POSTMETAMORPHIC BIOTITE GRANITE BENEATH THE ATLANTIC COASTAL PLAIN, SUFFOLK, VIRGINIA

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ABSTRACT

A drillhole, centered in a −40 mgal Bouguer gravity anomaly near the mouth of the Nasemond River, Virginia, at 36°51.01'N, 76°29.83'W, penetrated crystalline basement rocks beneath Atlantic Coastal Plain sediments at 557 meters depth. Fifty-four meters of recovered core of basement rock is in a postmetamorphic, biotite monzogranite. The granitic body, which is probably associated with the gravity anomaly, is termed the Portsmouth pluton. A Rb-Sr whole-rock isochron age of 263 ± 25 Ma (2σ MSRS = 0.8186) has been determined for the core. Biotite/whole rock ages are indistinguishable from the isochron age, indicating rapid cooling of the pluton. The Portsmouth is among the youngest and easternmost members of the group of late Paleozoic plutons in the southern Appalachians and was emplaced in a metamorphic terrain which was cooling as well. A moderately high $^{87}Sr/^{86}Sr$ initial ratio of 0.7076 ± 0.0012 indicates the influence of older crustal material.

INTRODUCTION

The late Paleozoic granitoid rocks of the southern Appalachians have been used as benchmarks for determining the nature and timing of metamorphic and deformational events in the southeastern U. S. (for example, Glover and others, 1983) and have been interpreted as products of a former magnatic arc (Sinha and Zietz, 1982). These plutons are increasingly abundant toward the Atlantic Coastal Plain (Figure 1), but of those believed to be buried beneath coastal plain sediments, only the Dort pluton, North Carolina (Becker, 1981) and the Springfield pluton of South Carolina (Speer, 1982) have been described.

An approximately 50 mgal Bouguer gravity anomaly, centered at the mouth of the Nansemond River, Virginia (Figure 2), suggested the presence of an additional granitic body beneath the Atlantic Coastal Plain. A drill-site (CP25A) was located near the center of the circular anomaly at 36°51.01'N, 76°29.83'W to obtain heat flow measurements and petrographic samples. A continuous, 1-1/2 inch diameter core was obtained from 557 to 611 m (1828-2005 ft.). The basement rock obtained from drillcore CP25A is a nonfoliated granitoid which is here in named the Portsmouth pluton. The sample numbers in the following discussions indicate the depth in feet.

Samples of basement rock from beneath the Atlantic Coastal Plain are rare and are generally limited to cuttings or short pieces of core. Thus, study of the petrography, geochemistry and geochronology of this subsurface granitoid is in greater detail than is usually possible. In particular, the Rb-Sr whole-rock isochron technique has been used to determine the age of the pluton. Previous reliable geochronologic information about the subsurface Appalachians in the southeast has been limited to mineral cooling ages (Dennison and others, 1967).

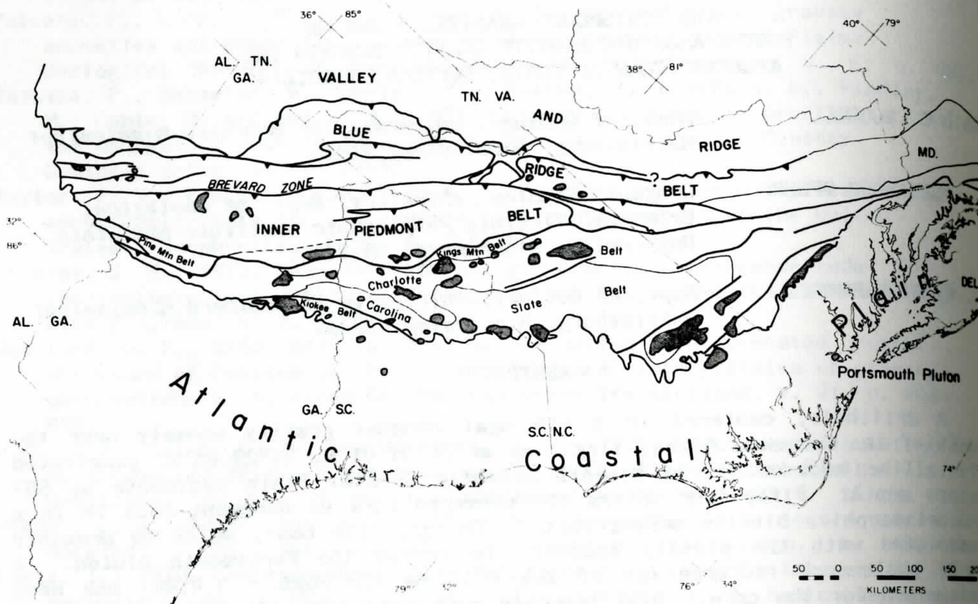

Figure 1. Occurrence of postmetamorphic, granitoid plutons (black) in the southern Appalachians, and the location of the Portsmouth pluton drillhole site.

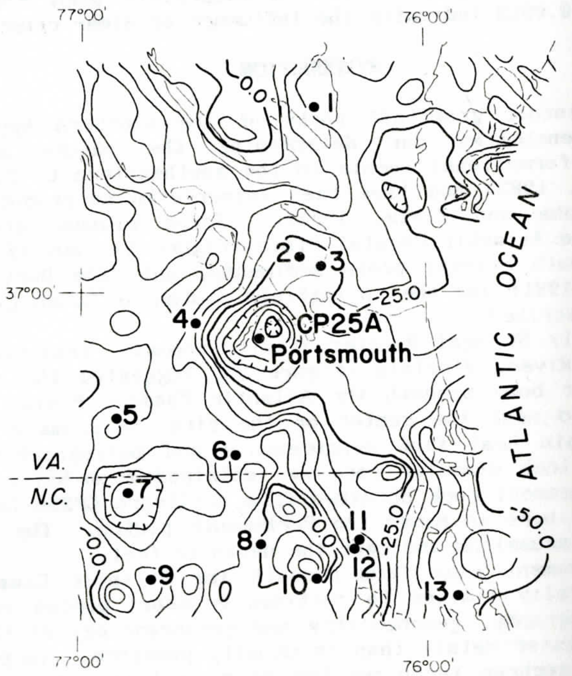

Figure 2. Location of drillhole CP25A, Suffolk, Virginia, superimposed on the Bouguer gravity map constructed from data in the Department of Defense Gravity Library (DOD, 1976). Numbers refer to drillholes encountering basement rocks, (Table 8). Contour interval is 5 mgal.

PETROGRAPHY

The CP25A basement core is composed of a massive, orange-pink, coarse-grained, inequigranular granitoid rock. A point count of a slab stained for plagioclase and alkali feldspar shows that the rock is a monzogranite, with 29% quartz, 30% K feldspar, 38% plagioclase, and 3% biotite + accessory minerals. The color index of the monzogranite, based on visual estimates, varies from less than 1 to 5 along the length of the core. Average density of the granite 2.63 gm cm $^{-3}$. The alkali feldspar occurs as anhedral to subhedral orange crystals up to 15 mm long. Plagioclase is present as orange or white, euhedral to subhedral crystals up to 7 mm long. Quartz is grey and anhedral, with grains less than 6 mm in diameter. Biotite is flakes 5 mm or less in length is the most abundant mafic minieral.

Accessory minerals in the granite include allanite, apatite, calcite, chlorite, epidote, fluorite, ilmenite, magnetite, thorite, and zircon. Inclusion of euhedral to subhedral allanite, apatite, and zircon in the major minerals suggests that these accessory minerals are igneous minerals. The occurrence of titanite rims on ilmenite and magnetite suggests that, in part, the titanite is a later reaction product. Most calcic cores of the plagioclases contain abundant muscovite, calcite, and epidote which are interpreted as subsolidus saussuritization products. The biotites are widely altered to chlorite, which replaces either the entire grain, just the rim, or patches. Accompanying the cholorite are fine needles of rutile and lenses of alkali feldspar, epidote, muscovite, and fluorite. By contrast, biotites which occur as isolated inclusions in the major minerals show none of these feature, suggesting that these chloritized biotites are alteration products.

The granite is cut by numerous veins up to 7 cm thick, which from their attitude in the presumably vertical core, must be gently dipping. The granite adjacent to these veins is extensively altered, with increased modal abundance of albite, calcite, chlorite, epidote, and muscovite. Within certain parts of the core, as much as 30 cm lengths are cut by numerous 2-3 mm thick veins and these zones are referred to as veined granites. The veins are predominately either quartz, carbonate minerals, or muscovite. The muscovite dominant-bearing veins and veined granite consist of optically zoned muscovite up to 2 cm across with lesser amounts of albite and quartz and variable amounts of fluorite, pyrite, pyrophanite, chalcopyrite, and thorite. Carbonate dominant-bearing veins in veined granite intervals contain Mn+Fe+Mg-bearing calcite, Ca+Mg+Mn-bearing siderite, barite, nickeliferrous pyrite, and uraninite.

MINERAL CHEMISTRY

Mineral compositions were determined in polished thin sections using the procedures described by Solberg and Speer (1982). Mineral formulas were calculated using a modification of the program SUPERECAL (Rucklidge, 1971).

Biotite

Pleochroic tan-to-brown biotite is widely and extensively altered in the granite. Biotites occurring as inclusions in the major minerals are the best preserved. Microprobe analyses (Table 1, Figure 3) of these included biotites show them to be aluminous, intermediate phlogopite-annite solid solutions with $Fe_{total}/(Fe_{total} + Mg)$ between 0.48 and 0.58. The biotites are fluorian with F/(F + Cl + OH) between 0.17 and 0.27.

Feldspar

The alkali feldspar in the Portsmouth granite is microcline with locally developed microperthite. The average composition is Or_{94} with a range of

Table 1. Average probe analyses of biotites, Portsmouth pluton, VA.

CP25A	1831	1856.5	1956	1959.3	1981	1991
SiO_2	37.06	37.18	36.30	37.34	37.40	37.36
TiO_2	2.40	2.09	2.60	2.34	2.57	2.41
Al_2O_3	15.44	15.28	18.16	15.07	15.54	15.28
FeO*	20.43	18.18	19.13	19.97	19.15	19.76
MnO	0.55	0.76	0.52	0.43	0.57	0.47
MgO	9.78	11.57	9.54	10.28	10.46	10.33
CaO	0.04	0.0	0.21	0.03	0.08	0.07
BaO	0.27	0.07	0.31	0.20	0.25	0.17
Na_2O	0.07	0.05	0.12	0.08	0.08	0.07
K_2O	9.78	9.79	9.34	10.37	9.91	9.72
F	1.58	2.27	1.73	1.59	1.69	1.60
Cl	0.05	0.08	0.07	0.04	0.04	0.08
H_2O**	3.16	2.84	3.10	3.16	3.14	3.15
Total	100.61	100.16	101.13	100.90	100.88	100.47
-O=F+Cl	0.68	0.97	0.74	0.68	0.72	0.69
Total	99.93	99.19	100.39	100.22	100.16	99.78
number of cations on the basis of 24(O,OH,Cl,F)						
Si	5.676	5.681	5.485	5.703	5.681	5.704
Al^{iv}	2.324	2.319	2.515	2.297	2.319	2.296
	8.000	8.000	8.000	8.000	8.000	8.000
Al^{vi}	0.463	0.432	0.718	0.416	0.463	0.453
Ti	0.276	0.240	0.295	0.269	0.294	0.277
Fe^{+2}	2.617	2.323	2.417	2.551	2.433	2.523
Mn	0.071	0.098	0.067	0.056	0.073	0.061
Mg	2.233	2.635	2.149	2.340	2.368	2.351
	5.660	5.729	5.646	5.632	5.631	5.664
Ca	0.007	0.0	0.034	0.005	0.013	0.011
Na	0.021	0.015	0.035	0.024	0.024	0.021
K	1.911	1.908	1.800	2.020	1.920	1.893
Ba	0.016	0.004	0.018	0.012	0.015	0.010
	1.954	1.927	1.888	2.061	1.972	1.935
Cl	0.013	0.021	0.018	0.010	0.010	0.021
F	0.765	1.097	0.827	0.768	0.812	0.773
H	3.229	2.895	3.125	3.220	3.182	3.208
	4.007	4.013	3.969	3.998	4.004	4.001
Fe/(Fe+Mg)	0.546	0.479	0.536	0.525	0.514	0.524
Number of analyses	3	4	2	6	4	4

*Total iron as FeO.
**Calculated to give 4(OH+F+Cl).

Or_{96} to Or_{92} (Table 2). Calcium contents are less than $An_{0.2}$ and [illegible] contents range up to $Cn_{1.18}$. Plagioclase has weak oscillatory, normal zoning from An_{21} to An_7 with local, discontinuous An_0 rims; potassium content corresponds to less than $Or_{1.3}$ (Table 2). In many of the plagioclases, the originally calcic cores have been saussuritized and now comprises albite with abundant inclusions of muscovite, calcite, and epidote. Plagioclase in the veins and veined granites (CP25A 1837.0 and 1837.8, Table 2) is albite.

Muscovite

White mica has a variety of occurrences in the granite: in the matrix, as replacement of plagiclase, and in the veins and veined granite. Microprobe analyses (Table 3) show that micas are phengitic with $Fe/(Fe+Al^{VI}+Mg+Ti)$ up to 0.17 and $Mg/(Fe+Al^{VI}+Mg+Ti)$ of 0.13. They contain only a minor paragonite component with Na/(Na+K) less than 0.05. The muscovites' low totals, high silica, low alumina, and low alkali contents indicate that they

Table 2. Selected probe analyses of feldspars, Portsmouth pluton, VA.

CP25A	1831	1831	1831.8	1831.8	1837.0	1837.8	1856.5	1856.5	1898	1898	1856	1901.5	1901.5	1920	1920	1959.3	1981	1981	1991	1991
SiO_2	67.87	65.34	65.45	66.34	66.31	68.36	64.07	67.99	65.73	63.93	62.55	68.06	69.65	63.62	64.86	63.31	65.02	65.30	65.89	64.24
Al_2O_3	20.83	18.54	21.51	18.72	18.75	19.29	18.56	20.48	18.59	22.43	20.18	19.51	19.92	18.01	21.63	18.08	18.89	22.56	19.12	22.23
CaO	1.46	0.01	2.62	0.0	0.20	0.10	0.0	0.71	0.01	4.13	0.04	0.09	0.53	0.0	2.94	0.0	0.02	3.25	0.02	3.20
BaO	0.0	0.26	0.0	0.0	0.0	0.0	0.0	0.0	0.08	0.00	0.57	0.0	0.0	0.0	0.0	0.21	0.21	0.02	0.14	0.11
Na_2O	9.84	0.63	8.76	0.65	10.44	11.31	0.59	9.45	0.68	8.39	0.72	9.80	10.29	0.52	9.49	0.63	0.46	8.49	0.78	8.21
K_2O	0.04	15.81	0.14	16.54	0.07	0.04	16.64	0.22	16.46	0.15	15.47	0.04	0.03	17.28	0.15	16.19	16.27	0.08	15.99	0.11
Total	100.04	100.59	98.48	102.25	95.77	99.10	99.86	98.85	101.55	99.03	99.53	97.50	100.42	99.43	99.07	98.42	100.87	99.70	101.94	98.10
number of cations on the basis of 8 (0)																				
Si	2.956	2.999	2.904	2.999	3.011	3.006	2.977	2.984	2.996	2.839	2.916	3.021	3.007	2.983	2.876	2.986	2.983	2.866	2.984	2.866
Al	1.069	1.003	1.124	0.997	1.003	1.000	1.016	1.059	0.998	1.174	1.108	1.020	1.015	0.995	1.130	1.005	1.021	1.167	1.020	1.169
Ca	0.068	0.0	0.121	0.0	0.010	0.005	0.0	0.033	0.0	0.196	0.002	0.004	0.025	0.0	0.140	0.0	0.001	0.153	0.001	0.153
Na	0.831	0.056	0.753	0.057	0.919	0.964	0.053	0.804	0.060	0.722	0.065	0.843	0.862	0.047	0.816	0.057	0.041	0.723	0.068	0.710
K	0.002	0.926	0.008	0.954	0.004	0.003	0.986	0.012	0.957	0.008	0.920	0.002	0.002	1.033	0.008	0.974	0.952	0.004	0.924	0.006
Ba	0.0	0.005	0.0	0.0	0.0	0.0	0.0	0.0	0.001	0.0	0.010	0.0	0.0	0.0	0.0	0.004	0.004	0.000	0.002	0.002
An	7.56	0.05	14.04	0.0	1.04	0.48	0.0	3.93	0.05	21.19	0.20	0.50	2.76	0.0	14.48	0.0	0.10	17.36	0.10	17.56
Ab	92.19	5.68	85.06	5.64	98.52	99.26	5.11	94.62	5.90	77.89	6.53	99.23	97.05	4.37	84.64	5.54	4.10	82.09	6.88	81.51
Or	0.25	93.79	0.89	94.36	0.43	0.26	94.89	1.45	93.91	0.92	92.23	0.27	0.19	95.63	0.88	94.08	95.42	0.51	92.77	0.72
Cn	0.0	0.47	0.0	0.0	0.0	0.0	0.0	0.0	0.14	0.0	1.04	0.0	0.0	0.0	0.0	0.38	0.38	0.04	0.25	0.22

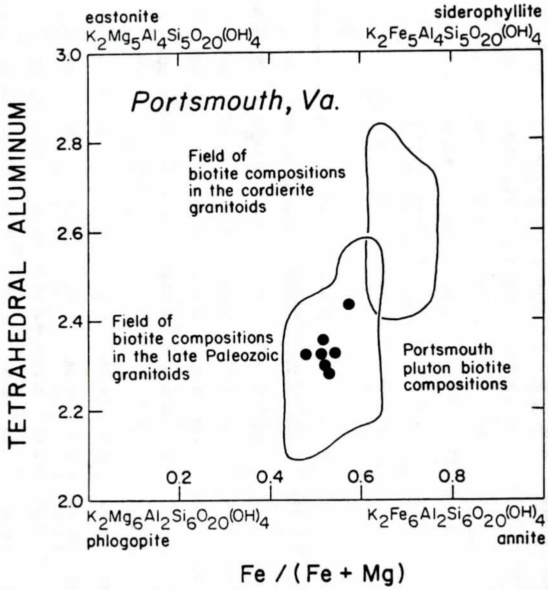

Figure 3. Average biotite compositions for the Portsmouth monzogranite projected onto the phlogopite-annite-eastonite-siderophyllite quadrilateral. Compositional fields for the biotites of the late Paleozoic and cordierite-bearing granitoids are from Speer and others (1980) and Speer (1981).

are hydromicas in part, and more likely, Li-muscovites, as suggested by their optical zoning. The muscovites contain variable amounts of F, ranging from 0.2 to 1.48 wt %.

Accessory And Secondary Minerals

The apatites are fluorapatites. The chlorites are pycnochlorites and ripidolites (Hey, 1954), with Fe/(Fe+Mg) of 0.44 to 0.50 (Table 4). An analysis of an epidote from CP25A 1959.3 shows a pistacite content of 31.6 mol % (Table 5). Ilmenites in the granites have nearly ideal end-member compositions; but the oxide phase from a veined granite is a pyrophanite with 62 mol % $MnTiO_3$.

MAJOR ELEMENT CHEMISTRY

Chemical analyses presented in Table 6 were obtained using a Phillips P. W. 1450 automatic sequential X-ray fluorescence spectrometer. Major elements, except for sodium, were analyzed from glasses prepared by fusing the sample with a lithium borate flux (Spectroflux 104). Procedures modified from Harvey and others (1972) and Norrish and Hutton (1969) were used in preparing glasses and in data reduction. Sodium analyses were obtained from pressed powders using techniques similar to Norrish and Chappell (1977). USGS standards PCC-1, BCR-1, AGV-1, GSP-1 and G-2 were used in calibration. Estimated standard deviations associated with each element measured, based on seven sets of duplicates over a period of nine months, are as follows: 0.14% for SiO_2, o.65% for TiO_2, 0.30% for Al_2O_3, 0.65% for Fe_2O_3, 6.3% for MgO, 1.23% for CaO, 0.5% for K_2O. Average differences in P_2O_5 and MnO were 0.01 wt %. C.I.P.W. normative minerals were calculated using a computer program written by Bowen (1971). The average U and Th contents of eight samples of Portsmouth granite reported by Perry and others (1980) are 8.9 and 21.5 ppm respectively. The Portsmouth granite is comparable in chemistry to the other coarse-grained granites in the southern Appalachians described by Speer and others (1980) except that it is more aluminous with an average normative corundum content of 1.05% compared to the average of 0.64%

Table 3. Averaged probe analyses of muscovites, Portsmouth pluton, Va.

CP25A	1831.8	1837.0	1837.8	1856.5	1901.5	1920	1959.3
SiO_2	48.44	46.89	47.32	48.92	47.72	46.74	47.74
TiO_2	0.97	0.69	0.73	0.69	0.75	1.31	0.31
Al_2O_3	30.34	28.40	28.43	28.23	28.43	27.98	28.31
FeO*	4.91	5.53	5.76	5.77	5.54	5.30	5.87
MnO	0.03	0.15	0.19	0.10	0.16	0.06	0.09
MgO	1.80	1.83	1.83	2.09	2.32	1.84	2.01
CaO	0.0	0.0	0.02	0.0	0.0	0.0	0.0
BaO	0.0	0.0	0.0	0.0	0.0	0.0	0.0
Na_2O	0.13	0.22	0.27	0.17	0.25	0.11	0.13
K_2O	7.77	8.88	10.44	8.74	9.07	8.24	8.66
F	0.55	0.83	0.97	0.75	1.07	0.44	0.65
Cl	0.0	0.0	0.0	0.0	0.0	0.13	0.0
H_2O**	4.20	3.92	3.92	4.07	3.88	4.05	4.04
Total	98.93	96.99	99.47	99.54	98.75	96.19	97.54
-O=F+Cl	0.23	0.35	0.41	0.32	0.45	0.22	0.27
Total	98.93	96.99	99.47	99.23	98.75	95.98	97.54
number of cations on the basis of 24 (O,OH,F,Cl)							
Si	6.503	6.513	6.480	6.620	6.517	6.529	6.579
Al^{iv}	1.497	1.487	1.520	1.380	1.483	1.471	1.421
Al^{vi}	3.301	3.161	3.068	3.122	3.092	3.134	3.177
Ti	0.098	0.072	0.075	0.070	0.077	0.137	0.032
Fe^{+2}	0.552	0.642	0.660	0.654	0.633	0.618	0.677
Mn	0.004	0.017	0.021	0.011	0.018	0.008	0.011
Mg	0.359	0.379	0.374	0.422	0.471	0.383	0.412
Ca	0.0	0.0	0.003	0.0	0.0	0.0	0.0
Na	0.033	0.058	0.072	0.045	0.066	0.028	0.035
K	1.331	1.573	1.823	1.509	1.580	1.468	1.522
Cl	0.0	0.001	0.0	0.0	0.0	0.030	0.001
F	0.236	0.365	0.418	0.323	0.464	0.197	0.283
H	3.765	3.635	3.581	3.679	3.538	3.774	3.714
Fe/(Fe+Mg)	0.607	0.635	0.646	0.612	0.580	0.620	0.625
Number of analyses	2	3	2	2	5	2	2

*Total iron as FeO
**Calculated to give 4(OH+F+Cl)

ISOTOPE CHEMISTRY

Analytical methods

Six samples of the 1-1/2 inch diameter core, weighing about 1 kg each, were selected for Rb-Sr whole rock analysis. The number and size of samples were limited by the requirement that they be free of fractures and mineralized zones. All analyses were made on the Nuclide six-inch radius of curvature, 60° magnetic sector, therminonic source mass spectrometer installed in the Orogenic Studies Laboratory at VPI&SU in April 1979. It is interfaced to a model 9825A Hewlett Packard programmable calculator (Nuclide's PC/SIM-1 automated system) for magnetic peak stepping and data reduction. Fourteen analyses of SRM 987 standard carbonate which were run during this study yielded an $^{87}Sr/^{86}Sr$ isotopic composition of 0.71022 ± 0.00016 (2σ). All strontium isotopic analyses have been normalized to $^{86}Sr/^{88}Sr$ = 0.1194. The analytical uncertainties for $^{87}Sr/^{86}Sr$ and

$^{87}Rb/^{86}Sr$ measurements are 0.04% and 2% (both 2σ), based on 29 sets of duplicate analyses of granitic and felsic volcanic whole-rock samples. Rubidium and strontium concentrations were determined by isotope dilution using ^{84}Sr spike prepared by the National Bureau of Standards (SRM 988) and ^{87}Rb prepared at Oak Ridge National Laboratories. Ages and initial ratios

Table 4. Averaged probe analyses of chlorites, Portsmouth puton, Va.

CPA25A	1831.8	1898	1920	1959.3
SiO_2	27.05	28.81	26.76	26.13
TiO_2	0.10	0.14	0.16	0.03
Al_2O_3	18.86	17.32	18.98	20.64
FeO*	24.45	24.98	25.33	22.49
MnO	0.66	0.53	0.65	0.87
MgO	15.93	14.99	14.71	16.88
CaO	0.0	0.08	0.01	0.00
BaO	0.0	0.06	0.0	0.0
Na_2O	0.01	0.05	0.02	0.01
K_2O	0.0	0.06	0.08	0.0
F	0.15	0.09	0.22	0.34
Cl	0.0	0.02	0.07	0.01
H_2O**	11.26	11.28	11.10	11.26
Total	98.49	98.41	98.08	98.66
-O=F+Cl	0.06	0.04	0.11	0.15
Total	98.42	98.37	97.02	98.52
number of cations on thebasis of 18(O,OH,F,Cl)				
Si	2.856	3.044	2.850	2.729
Al^{iv}	1.144	0.956	1.150	1.271
Al^{vi}	1.202	1.201	1.231	1.269
Ti	0.008	0.011	0.013	0.002
Fe^{+2}	2.159	2.207	2.256	1.964
Mn	0.059	0.047	0.059	0.077
Mg	2.507	2.361	0.335	2.627
Ca	0.0	0.0	0.0	0.0
Na	0.002	0.010	0.004	0.002
Ba	0.0	0.002	0.0	0.0
Cl	0.0	0.004	0.012	0.002
F	0.050	0.030	0.073	0.112
H	7.934	7.966	7.915	7.886
Fe/(Fe+Mg)	0.469	0.489	0.498	0.437
Number of analyses	3	1	3	1

*Total iron as FeO
**Calculated to give 8(OH+F+Cl)

are calculated using the York model II equation (York, 1969). All errors are given at the 2σ confidence level. The weighted sum of the residuals squares (MSRS analogous to MSWD of the McIntyre and others, 1966, program) is given as a measure of the linearity of the data. Biotite/whole-rock ages were calculated as two-point isochrons.

Standard dissolution and ion exchange techniques were used. High purity water and HCl were obtained by tandem distillation of deionized water in quartz and teflon sub-boiling stills; other acids were distilled using the two-bottle teflon still described by Mattinson (1972). All procedures are carried out in 100+ clean air. Blanks for both rubidium and strontium are less than 200 picograms.

The results are listed in Table 7 and plotted on Figure 4. The whole rock isochron age is 263 ± 25 Ma (2σMSRS = 0.8186). This is interpreted as the age of crystallization of the granite. The initial $^{87}Sr/^{86}Sr$ composition of 0.7076 ± 0.0012 is higher than most other granites in the southern Appalachians (Fullagar and Butler, 1979). Biotites with $^{87}Rb/^{86}Sr$ values of 44.1 and 268, yield identical biotite/whole rocks ages of 267 Ma.

Table 5. Averaged probe analyses of epidotes, Portsmouth pluton, Va.

CP25A	1959.3
SiO_2	36.95
TiO_2	0.09
Al_2O_3	21.98
Fe_2O_3*	16.03
MnO	0.27
MgO	0.0
CaO	22.83
BaO	0.0
Na_2O	0.02
K_2O	0.0
F	0.14
Cl	0.0
H_2O**	1.80
Total	100.12
-O=F+Cl	0.06
Total	100.06
number of cations on the basis of 13 (O,OH,F,Cl)	
Si	2.962
Al^{iv}	0.038
Al^{vi}	2.038
Ti	0.005
Fe^{+3}	0.967
Mn	0.019
Ca	1.961
Na	0.002
K	0.0
Ba	0.0
Cl	0.0
F	0.037
H	0.963
Number of analyses	2

*Total iron as Fe_2O_3
**Calculated to give 1(OH+F+Cl)

Table 6. Major element analyses, Portsmouth pluton, Va.

CP25A	1848'	1916'	1981'
SiO_2	73.69	72.85	71.92
TiO_2	0.21	0.20	0.32
Al_2O_3	14.05	14.42	14.44
Fe_2O_3*	1.37	1.40	2.08
MnO	0.04	0.02	0.04
MgO	0.30	0.35	0.69
CaO	0.92	1.11	1.36
Na_2O	3.72	4.18	4.00
K_2O	4.51	4.55	4.40
P_2O_5	0.08	0.06	0.11
LOI**	0.77	0.76	0.78
Total	99.66	99.90	100.14
norms			
Q	32.71	28.42	28.14
C	1.57	0.75	0.89
Or	26.74	26.91	25.96
Ab	31.59	35.41	33.80
An	4.06	5.12	6.02
En	0.75	0.87	1.72
Hm	1.38	1.40	2.08
Il	0.09	0.04	0.09
Ru	0.17	0.18	0.28
Ap	0.19	0.14	0.26
Total	99.25	99.24	99.24

*Total iron as Fe_2O_3
**Loss on ignition.

Table 7. Rb and Sr data for whole rock and biotite separates from drillhole CP25A, Portsmouth pluton, Va.

Depth (ft.)	Rb (ppm)	Sr (ppm)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$
Whole rock				
1848	253	209	3.507	0.72064
1853	231	220	3.033	0.71907
1916	246	228	3.130	0.71935
1919	296	190	4.500	0.72435
1942	268	199	3.890	0.72244
1956	255	222	3.335	0.71993
Whole rock/biotite pairs				
1848				
WR	253	209	3.507	0.72064
Biot	551	36.7	44.127	0.87465
1956				
WR	255	222	3.335	0.71993
Biot	1.24	14.5	268.24	1.72538

Depth (ft.)	Biotite/Whole Rock Age	Calculated $(^{87}Sr/^{86}Sr)_{initial}$
1848	267 ± 7 m.y.	0.7073 ± 0.0016
1956	267 ± 5 m.y.	0.7073 ± 0.0015

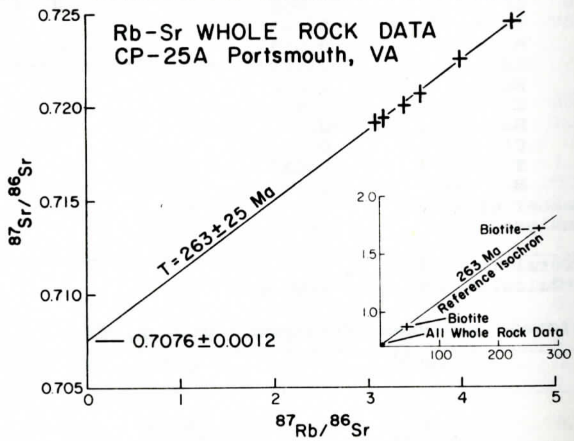

Figure 4. Rb–Sr isochron plot for whole rock samples from the Portsmouth monzogranite, Virginia. Inset shows the relation of the two biotite separates to the extension of the whole rock isochron.

This age is indistinguishable from the whole rock age and indicates that the Portsmouth pluton cooled rapidly past the blocking temperature of biotite, approximately 300°C (Dodson, 1979).

DISCUSSION

The Portsmouth pluton is similar in texture and mineralogy to the postmetamorphic, coarse-grained granitoid rocks of the southern Appalachians as described by Speer and others (1980). The Portsmouth joins this group as its easternmost member. It is among the youngest, only the Siloam, Georgia (269 ± 3 Ma, Jones and Walker, 1973) and the Edgefield, South Carolina (254 ± 11 Ma, Snoke and others, 1980) are as young. The evolved character of the Portsmouth granite, indicated by its leucocratic nature, peraluminous chemistry, F-rich silicate minerals, extensive late-magmatic or subsolidus reactions and mineral assemblages, and abundant quartz veins and high U contents, suggests that the drillhole encountered the late crystallizing and fluid-rich portion, or cupola, of the magma chamber. The well-defined circular gravity anomaly of the Portsmouth pluton and the undeformed texture of the granite suggest that the Portsmouth is a postmetamorphic pluton. The close agreement of the whole-rock isochron age and biotite ages indicate that the country rocks were not hot enough, at the time of intrusion, to retard cooling.

Several drillholes which encounter basement rocks in the vicinity of the Portsmouth pluton are shown in Figure 2 and summarized in Table 8. The Portsmouth pluton is located in a terrane of low-to-medium-grade metavolcanic, metagranitic and metapelitic rocks. To the south, three drillholes may have encountered Masozoic sedimentary and igneous rocks (holes 6, 10, 11; Figure 2). A Rb-Sr model age of the metavolcanic rocks in drillhole no. 12 is 408 ± 40 Ma (Dennison and others, 1967). This age may be the rocks' age of igneous crystallization, a cooling age of metamorphism, or a mixed age. A K-Ar muscovite age of 253 ± 5 Ma for the schist in drillhole no. 13 is reported by Dennison and others (1967). This would presently be interpreted as a metamorphic cooling age, and is similar to the biotite cooling age of the Portsmouth pluton. The region of the Portsmouth pluton was metamorphosed prior to the emplacement of the granite at 263 Ma and subsequently cooled rapidly, perhaps as a result of uplift during the Alleghanian orogeny.

Table 8. Summary of drillcore data for holes located in Figure 2.

Hole	Basement Rock Encountered	Reference
Virginia		
1	granite	Johnson (1975)
2	granite	"
3	low-grade lithic/crystal tuff	"
4	metagranite	Gleason (1982)
5	granite	Johnson (1975)
6	Triassic(?) sandstone	"
North Carolina		
7	syndeformational granite (Dort)	Becker (1981)
8	metarhyolite	Coffey (1977)
9	quartzite	"
10	basalt/diabase (Triassic?)	"
11	red shale/basalt (Triassic?)	Richards (1954)
12	felsic lithic/crystal metatuff	Dennison and others (1967)
13	garnet-biotite-muscovite-schist	"

The late Paleozoic plutons of the southern Appalachians have recently been described as a Hercynian (or Alleghanian) magmatic arc extending from Maryland to Georgia (Sinha and Zietz, 1982). Sinha and Zietz (1982) plotted the geographic distribution of chemical and isotopic data and related the origin of the plutons to a westward-dipping subduction zone. Their conclusion is based on $^{87}Sr/^{86}Sr$ initial ratios and the $\delta^{18}O$ values increasing toward the west, suggesting greater involvement of continental crust. The Portsmouth pluton, located on the eastern edge of the "magmatic arc," has a much higher $^{87}Sr/^{86}Sr$ initial ration than would be predicted by the model of Sinha and Zietz (1982). Indeed, many of the late Paleozoic plutons with $^{87}Sr/^{86}Sr$ initial ratios greater than 0.706 occur on the eastern edge of exposed Piedmont along the Fall Line: Falmouth Intrusive Suite, Va. (0.7088, Pavlides and others, 1982), Castalia pluton, N. C. (0.7141, Fullagar and Butler, 1979), Lake Murray pluton, S. C. (0.7115, Fullagar and Butler, 1979), Edgefield pluton, S.C. (0.7107, Kish and Fullagar, 1978), and Clouds Creek pluton, S. C. (0.7099, Snoke and others, 1980). There are only two, presently known late Paleozoic plutons in the western Piedmont with such high $^{87}Sr/^{87}Sr$ initial ratios: the Cherryville pluton, N. C. (0.715, Kish, 1977) and the Stone Mountain pluton, Ga (0.7250, Whitney and others, 1976). These initial ratios must be caused by some interaction with crustal materials, either in the source region or by assimilation and requires additional complicating factors in the mechanism of magma generation and transport.

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ALLUVIAL MORPHOLOGY AND STRATIGRAPHY OF A MEANDERING SEGMENT OF THE AMITE RIVER, SOUTHEASTERN LOUISIANA

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ABSTRACT

Morphologic and stratigraphic patterns in the Amite River alluvial valley of southeastern Louisiana reflect its floodplain development by lateral accretion and chute cutoff processes. The alluvial fill is composed of a lower sandy facies and an upper silty facies. The upper surface of the buried sandy facies has a similar configuration to the land surface. Point bar and channel lag deposits are present in this facies. At some localities the overlying silty facies is in sharp contact with the sandy facies; at other localities this contact has a loamy gradation between alluvial units, with thicknesses ranging from a few centimeters to more than one meter.

The silty facies is composed of an upper brown and a lower gray silt unit. The brown silt directly overlies the sandy facies where the gray silt is absent. The brown silt, which is very porous and has a relatively low density, is generally highly bioturbated, but may display primary sedimentary structures. The brown silt is associated with natural levee and chute channel-fill deposits and with the upper part of abandoned channel-fill deposits. The gray silt has a truncated soil profile typified by blocky structure, common sesquioxide stains above the low water table, and numerous crack and ped fillings; the density of the gray silt is greater than that of brown silt. The gray silt generally occupies the lower parts of abandoned channel-fill deposits and some chute channel-fill deposits.

Morphologic and stratigraphic characteristics of the Amite River alluvial fill are strikingly similar to sedimentary facies in the modern Amite River. Active and abandoned point bars have a similar plan form geometry and comparable patterns of sedimentary structures and grain-size trends; active and abandoned channels also have similar dimensions and meander geometries. These observations indicate similar process and response mechanisms for the active and abandoned fluvial systems of the Amite River.

INTRODUCTION

The Amite River alluvial valley of southeastern Louisiana has developed a meandering-channel system with a coarse-grained bedload. Its morphologic and stratigraphic characteristics reflect the sedimentary processes which dominated its geologic development. A small segment of the alluvial valley with well-developed lateral accretion topography was selected for morphologic and stratigraphic analysis. The study area (Fig. 1) is located in the middle of the Amite River valley profile (zone 2 of Schumm, 1977), a short distance upsteam from the Amite-Comite River confluence. Lateral accretion topography is common for several miles upstream and downstream of the study area, and the features represented in this area appear to be typical of the entire reach.

In the study area the Amite River has incised the late Pleistocene Prairie Terrace which locally occupies interfluves. This terrace is in the Pine Meadows physiographic subsection (Thornbury, 1965). To the north, the Amite River dissects higher and older Pleistocene terraces (Fisk, 1938; Snead and McCulloh, 1984) in the Pine Hills physiographic subsection (Thornbury, 1965). To the south, the Amite River valley merges with the coastal swamps of the Mississippi River delta plain.

This investigation assesses the morphology and stratigraphy of a segment of the Amite River alluvial valley in order to interpret its fluvial processes and environments. The objectives are to 1) define sedimentary

Figure 1: Location of study area.

facies based on morphologic, geometric, and lithologic criteria and 2) infer depositional environments associated with sedimentary facies.

Previous investigations in the Amite River (McGowen and Garner, 1970; Autin and Fontana, 1980) have focused on active sedimentation, especially bedload deposits associated with channel and point bar facies. No other significant investigations of the alluvial morphology or stratigraphy of the Amite River have been published. No time-stratigraphic relationships have been established for the drainage basin; however, important chronologies in nearby locations are available for the Mississippi River delta plain (Frazier, 1967; Saucier, 1963), the fluvial terraces in the Tunica Hills (Alford, Kolb, and Holmes, 1983; Otvos, 1980; Delcourt and Delcourt, 1977), and tributary alluviation in the Bluff Hills of north-central Mississippi (Grissinger, Murphey, and Little, 1982).

Aerial photographs, topographic maps, and soil series maps were used to review alluvial morphology and select the study area. Field surveys of topography by plane table and alidade and transit methods were used with published maps to define morphologic features. Exposed sedimentary sequences along modern stream cutbanks and the walls of commercial sand and gravel pits were used to observe and describe lithologic properties and the nature of stratigraphic contacts. Borings 2 to 8 m (6-25 ft) deep were collected with a Giddings hydraulic probe to construct stratigraphic cross sections. The clay content of selected vertical profiles was analyzed to check field differentiation of stratigraphic units.

TOPOGRAPHY

The Prairie Terrace is a low-relief, constructional landform which has a predominately east-west orientation across southeastern Lousiana. Locally, it extends up the larger valleys as a fluvial terrace. A fluvial component of the Prairie Terrace exists in the Amite River north of the study area and along interfluves in the study area. In general, topographic highs on the terrace follow the meandering courses of abandoned streams. The Prairie Terrace is mantled by a blanket of loess, with thicknesses greater than 3 m (10 ft) near the Mississippi River alluvial valley and about 1 m (3 ft) in and around the study area (Miller and others, 1982).

Modern drainage networks incise the Prairie Terrace, resulting in sub-maturely dissected topography. The principal streams of the region, such as the Amite River, have developed as a result of entrenchment, which was followed by lateral planation while base level rose during drainage basin aggradation. Drainage patterns of tributary streams are not fully integrated into dendritic networks (Schumm, 1956).

In the study area, elevations on the Prairie Terrace reach slightly over 17 m (50 ft) on upland interfluves (Fig. 1). The terrace has been incised to elevations near 2 m (6 ft) MSL along the edge of the alluvial fill and possibly deeper near the center of the valley axis.

Landforms identified in the Amite River alluvial valley indicate floodplain development primarily by lateral accretion. Topographic variations on the floodplain floor are caused by the development of point bars and scroll bars and the filling of thalweg channels and chute channels. Local elevations reach 9 to 11 m (30-35 ft) on crests of point bars and major scroll bars and slightly less than 5 m (15 ft) in abandoned channels (Fig. 2). In plan view, shapes of point bars range from lenticular to crescentic, with maximum dimensions of up to 490 m (1600 ft). Scroll bars are smaller, being less than 300 m (1000 ft) long, and are mostly elongated in plan form. Remnants of abandoned meanders are arcuate in plan form and have maximum channel widths of 90 m (300 ft).

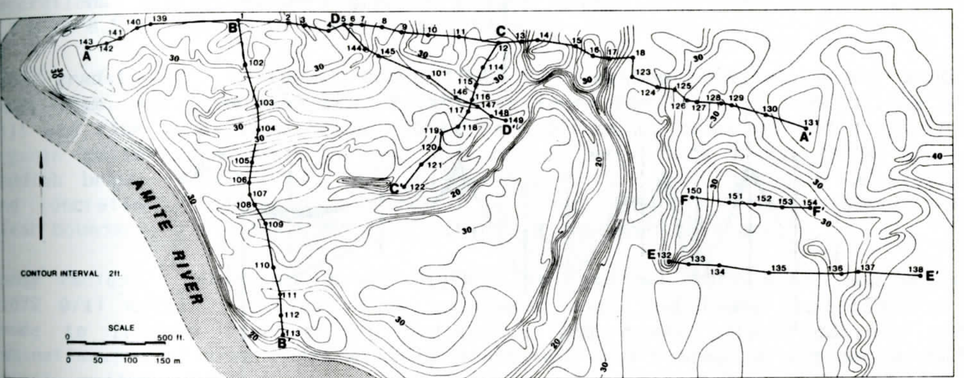

Figure 2: Study area topography and location of stratigraphic cross sections.

Scroll bars form within meandering-channel systems behind major point bars. Swales between scroll bars have widths less than 60 m (200 ft), but dimensions of multiple-channel sets can attain aggregate widths of up to 300 m (1000 ft). Scroll bars are initially formed during lateral accretion by a process similar to that described by Hickin (1974). During flood events, unconsolidated material along the concave bank is eroded. Patterns of current velocity and shear stress cause sediments to be deposited on the convex bank and form a longitudinal ridge along the channel bank. The rate of accretion on the convex bank is in equilibrium with the rate of erosion on the concave bank. Through successive flood events the channel maintains a nearly constant geometry. The result is a set of laterally stacked ridges, or scroll bars, with intervening swales between bars. The meander will develop a symmetrical set of ridges and swales during migration as long as the channel freely meanders under similar hydrologic conditions. During later flood events, the swales between scroll bars are used as chutes for secondary currents which split from the thalweg and cross the floodplain behind point bars.

STRATIGRAPHY

Sedimentary deposits in the study area can be divided into stratigraphic units based on variations in texture, color, sedimentary structures, degree of weathering, sediment body geometry, relative stratigraphic position, and topographic landscape position. Two lithostratigraphic units appear on uplands beneath the Prarie Terrace: the Prairie Formation and its overlying blanket of loess. Two lithostratigraphic units appear beneath the floodplain within the alluvial fill. Above the unconformable contact with the Prairie Formation, a lower sandy facies and an upper silty facies are identified. The upper silty facies can be divided into gray silt and brown silt. A loamy gradational zone is present between alluvial sedimentary units at some locations. Composite field descriptions of these units are provided in Table 1. Geometric relationships between lithologic units are illustrated in cross sections A-A' to F-F' (Figs. 3-8).

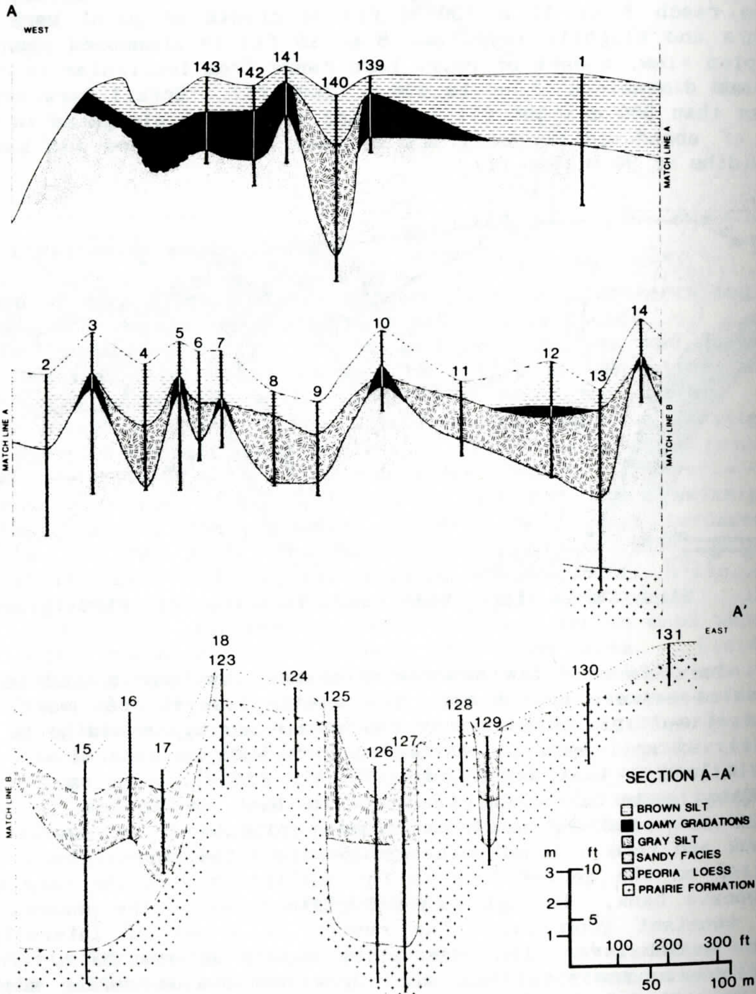

Figure 3: Cross section A-A' runs perpendicular to the valley axis from the modern channel to the outcrop of the Prairie Formation. The geometry of point bar, scroll bar, and abandoned channel-fill deposits are illustrated.

Upland Stratigraphy

In and near the study area the Prairie Formation is a greenish gray to light gray clay when unweathered. Samples commonly have thin laminations of clay and silt with few interbeds of fine sand. A complete stratigraphic section has not been described during this investigation, but Morgan (1961)

Table 1. COMPOSITE DESCRIPTION OF STRATIGRAPHIC UNITS

Brown silt: Yellowish brown (10YR 5/4) and brown (10YR 5/3) with white (10YR 8/2, 5YR 8/2), reddish brown (7.5YR 6/6), and strong brown (7.5YR 5/6) mottles; mostly silt loam with common loam, sandy loam and sandy clay loam textures; weak, fine to medium, crumb structure; friable, soft, or slightly hard consistency; laminations and interbeds of clay, silt, or fine sand; abundant plant roots; scattered decomposed organic matter; common worm burrows and pores; very dark brown (10YR 2/2) and red (2.5YR 4/8) sesquioxide stains and concretions; abrupt to clear boundary; 0.5-2.3 m (1.5-7 ft) thick in observed sections.

Gray silt: Gray (10YR 5/1, 5YR 5/1, 5G 5/1) with yellowish brown (10YR 5/6) and reddish brown (7.5YR 6/8) mottles; silt loam, sandy clay, loam, and loamy sand textures; weak, medium, subangular blocky structure; hard and firm or sticky and plastic consistency; common plant roots; common decomposed wood and organic matter; few worm burrows and pores; very dark brown (10YR 2/2), strong brown (7.5YR 5/8), and red (2.5YR 5/8) sesquioxide stains and concretions; scattered chert granule gravel; abrupt to gradual boundary; 1.0-2.7 m (3-8.2 ft) thick in observed sections.

Loamy gradations: Yellowish brown (10YR 5/4) without mottles; loamy sand, sandy loam, and loam textures; friable to loose consistency; laminations of silt and fine sand; rare grass roots; rare pieces of charcoal and scattered organic matter; few worm burrows and pores; very dark brown (10YR 2/2), reddish brown (7.5YR 6/6), and yellowish red (5YR 5/8) sesquioxide stains and concretions; sporadic occurrence of chert granule gravel; abrupt to clear boundary; 0.1-2 m (0.3-6.1 ft) thick in observed sections.

Sandy facies: Very pale brown (10YR 8/3), yellow (10YR 8/6), and white (10YR 8/1) without mottles; fine to coarse sand and loamy sand textures; loose to friable consistency; horizontal and cross stratifications and laminations; scattered chert gravel; scattered wood fragments and decomposed organic matter; part of the unit generally water-saturated; red (2.5YR 4/8) and reddish yellow (7.5YR 6/6) sesquioxide stains above the water table; abrupt boundary; 0.1-3.8+ m (0.3-11.6 ft) thick in observed sections.

Peoria loess: Yellowish brown (10YR 5/4) and brownish yellow (10YR 6/6) with white (10YR 8/2) and very pale brown (10YR 7/3) mottles; silt loam and silty clay loam textures; moderate, medium, subangular blocky structure; hard, firm consistency; common plant roots; common worm burrows and pores; very dark brown (10YR 2/2) sesquioxide stains and concretions; clear to diffuse boundary; 0.5-0.9 m (1.5-2.7 ft) thick in observed sections.

Prairie Formation: Light yellowish brown (10YR 6/4), light gray (10YR 7/1), and greenish gray (5G 5/1, 5BG 6/1) with strong brown (7.5YR 5/6) and white (10YR 8/1) mottles; silty clay loam, clay, and silt loam textures; moderate, medium, subangular blocky structure; hard to very hard or sticky to plastic consistency; common worm burrows and pores; horizontal lamintations of clay, silt, and local occurrence of fine sand below soil profile; rare chert granule gravel; very dark brown (10YR 2/2) and strong brown (7.5YR 5/8) sesquioxide stains; lower boundary not encountered.

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indicates that the expected thickness of the Prairie Formation should exceed 30 m (100 ft). The upper part of the formation has weathered to a yellowish brown silt loam or silty clay loam with brown and white mottling. Abundant sesquioxide stains and concretions and fragipan development occur in the soil B horizon.

Loess forms a thin 1- to 2-m (3-6ft) blanket which covers the weathered

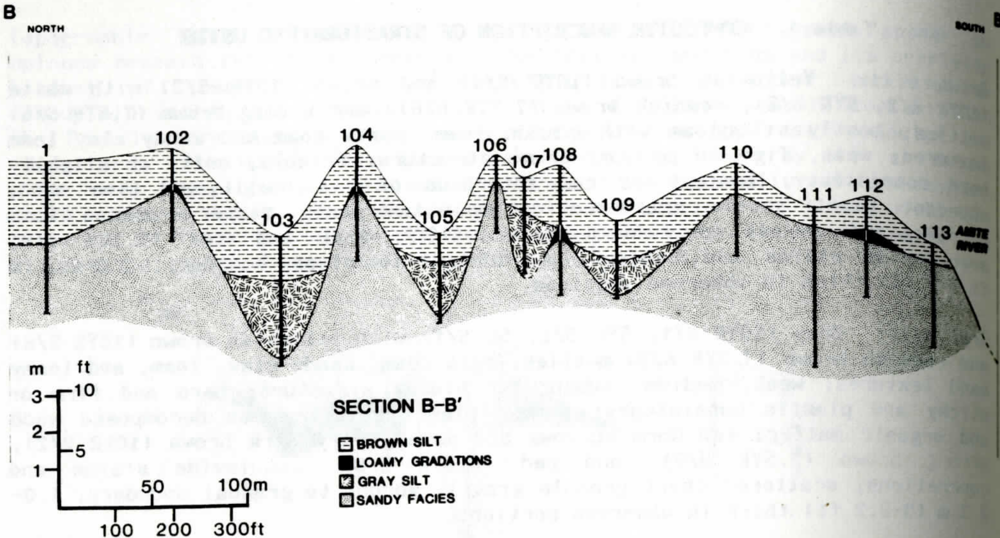

Figure 4: Cross section B-B' parallels the valley axis behind the largest relict point bar in the study area. The geometry of scroll bar and chute channel-fill deposits are illustrated.

surface of the Prairie Formation. In the study area, loess is a yellow-brown silt loam with brown and white mottles and common dark brown oxide stains and concretions. Plant roots, worm burrows, and a moderately developed subangular blocky structure typify weathering of the loess in the study area. The loess deposit in the study area has been correlated with the Peoria Loess of the Continental Interior (Miller and others, 1982).

Alluvial Stratigraphy

A sandy facies forms the lower member of the Amite River alluvial fill. This facies was found beneath overbank silts at every sampling site. The sand deposits are typically a coarse-to medium-grained quartz sand with granules of chert gravel commonly scattered in stringers or lenses. When exposed, a complete sequence displays well-developed tabular and trough cross-stratification overlain by ripple laminae in-phase, ripple-drift cross-laminations, and horizontal laminations. Arcuate and planar scour surfaces are common and generally form the bounding surfaces of related bedding sets. The base of the sandy facies has been encountered by drilling at five locations, indicating a maximum sand thickness of at least 6 m (18 ft). Total sand thickness is possibly greater near the center of the alluvial fill. The geometry of the base of the alluvial valley and the sedimentary character of the basal fill are poorly understood. The upper surface of the sandy facies undulates and generally conforms to the shape of the overlying land surface.

Overlying the sandy facies is a silty facies composed of an upper brown and a lower gray silt unit. The silty facies completely blankets the sandy facies. The brown silt is found directly beneath the land surface at all alluvial landscape positions and blankets the entire floodplain. The gray silt is buried by the brown silt and usually fills topographic lows. Maximum observed thicknesses of the silty facies approach 5 m (15 ft).

The brown silt is a yellowish brown to brown silt loam with either laminations or homogenized admixtures of fine to medium quartz sand. It commonly has faint white and brown mottles and red-brown stains. Most exposures and drill samples show heavy bioturbation caused by worm burrows and plant roots; however, the upper portion of this unit displays horizontal

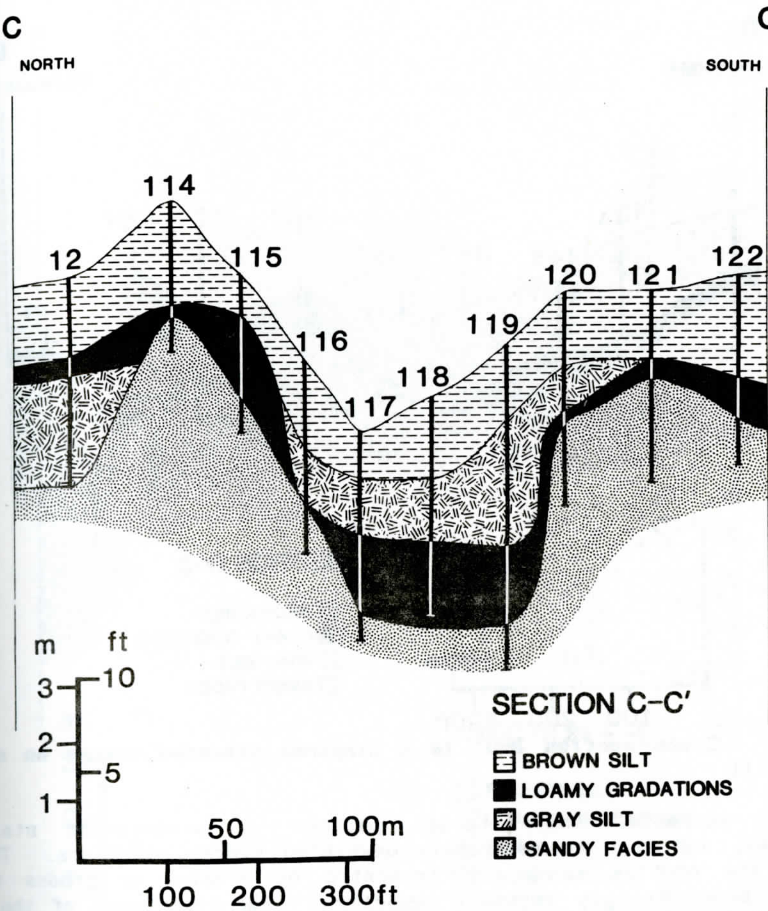

Figure 5: Cross section C–C' is a diagonal traverse across a relict point bar and adjacent abandoned channel fill.

laminations, ripple-drift cross-laminations, and ripple laminae in-phase. Primary sedimentary structures are especially well developed near the bank of the modern Amite River. After deposition, the sediment generally increases in porosity and decreases in density due to bioturbation. Maximum thickness of the brown silt approaches 2 m (6 ft). The lower contact of the brown silt is either gradational or in sharp contact with the gray silt or sandy facies.

At some locations where brown silt contacts gray silt, the brown silt has a basal sand unit which is generally less than 1 m (3 ft) thick. This sand unit is cross-bedded near its base and has common horizontal laminations, ripple laminae in-phase, and ripple-drift cross-laminations in its upper portions. This unit fines upward and grades into the typically bioturbated brown silt. At all locations where this sand unit was observed, the gray silt sequence is truncated by erosion, and the stratified sand unconformably lies on the scoured upper surface of the gray silt.

The gray silt commonly has a silt loam, loam, or sandy loam texture and typical yellow-brown and red-brown mottles. The unit has two distinct lithologic components: 1) an upper part, which has a relatively high density and a hard or firm consistency and shows distinct evidence of pedogenesis, and 2) a lower part, which is water saturated, has a plastic consistency, and is generally chemically reduced. Field evidence of soil profile development includes crack and root infilling, films of silt and clay which

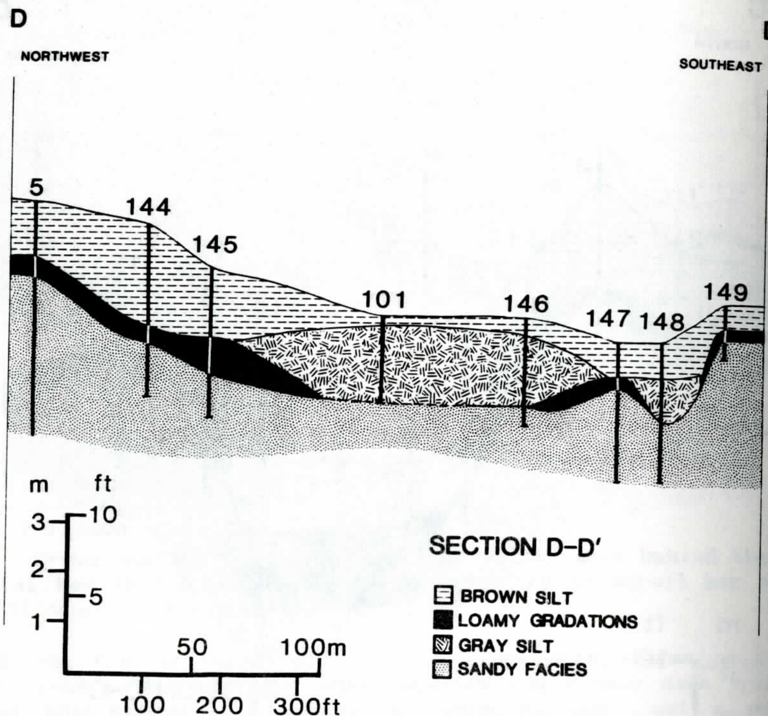

Figure 6: Cross section D-D' is a diagonal traverse across an abandoned channel fill.

coat ped surfaces, the internal pattern of sesquioxide stains and concretions, and weak to moderately developed blocky structure. The upper part of the profile is either truncated by erosion or grades into the overlying brown silt by vertical accretion. The lower part of the profile often contains fresh and macerated vegetable matter, such as wood, twigs, and leaves. Thickness of the gray silt ranges from nearly 1 m (3 ft) to almost 3 m (9 ft). The gray silt has an arcuate areal pattern and generally has a lens-shaped cross section because it fills low areas on the surface of the sandy facies. Gray silt generally coarsens towards the base, and the unit has either a gradual or abrupt contact with the sandy facies.

Some loamy gradations less than 2 m (6 ft) thick occur between major alluvial units. Gradational units generally are yellow-brown and have common faint laminations, but laminations may be homogenized by worm burrows and plant roots. Brown to red-brown oxide stains are also common.

A useful method for evaluating the placement of stratigraphic contacts by field observation is to plot changes in clay content with depth. Vertical sequences of brown silt overlying sandy facies and brown silt overlying gray silt were selected for evaluation (Fig. 9). In cores 3, 5, and 14 the brown silt overlies the sandy facies. At these sites a maximum in percent clay occurs in the brown silt within 1 m (3 ft) of the land surface. This increase in clay content with depth appears to indicate the translocation of clay from the upper to the lower portions of the brown silt. The sharp drop in clay content as the sandy facies is approached appears to be related to changes in sediment transport capacity. In cores 4, 15, and 16 the brown silt overlies the gray silt. At these sites a maximum in percent clay occurs in the brown silt within 1.5 m (5 ft) of the land surface. Clay content slowly decreases across the brown silt-gray silt boundary, but increases toward the base of the gray silt in cores 4 and 16.

There are two possible reasons for this variation across the brown silt-

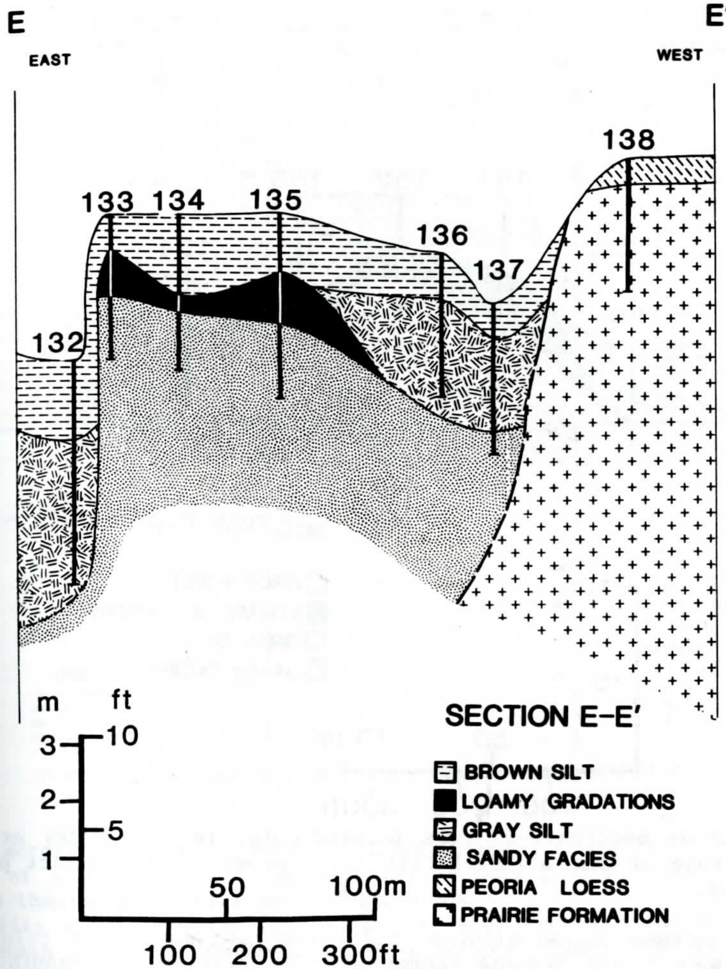

Figure 7: Cross section E-E' runs perpendicular to the valley axis along the eastern edge of the alluvial fill. The geometry of a relict point bar and abandoned channel fill are illustrated. The channel fill represented by core 132 appears to have cut the point bar deposit.

gray silt boundary. Changes in clay content are related to differences in sedimentary processes between the two units. The gray silt was deposited under a more variable set of sedimentary conditions than the brown silt. The development of the gray silt in landscape lows also causes it to be wetter than the overlying brown silt, possibly restricting the translocation of clays within the gray silt. However, other lines of evidence, such as the relative geometry of the two units and locally abrupt contacts between the units, suggest a sedimentologic difference between the gray and brown silts. Textural trends of additional vertical profiles need to be analyzed before these possibilities can be verified.

Trends of Cross Sections

Six cross sections were constructed from 72 borings to illustrate internal sediment body geometries of stratigraphic units (Fig. 2). The longest section, A-A', is oriented perpendicular to the valley axis and trends east-west along the northern edge of the study area (Fig. 3).

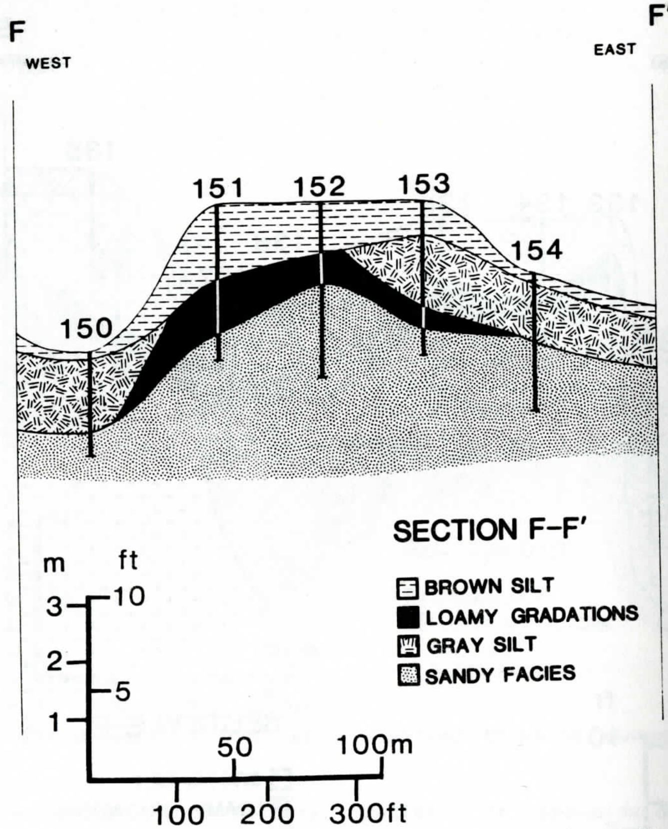

Figure 8: Cross section F-F' runs perpendicular to the valley axis along the eastern edge of the alluvial fill. The geometry of a relict point bar is illustrated.

Section A-A' transects the alluvial fill from behind a modern point bar to the outcrop area of the Prairie Formation. The primary north-south section, B-B', traverses well-developed scroll bar and chute channel topography behind the largest relict point bar of the study area (Fig. 4). Shorter cross sections C-C' and D-D' traverse an individual abandoned channel (Figs. 5 and 6), and sections E-E' and F-F' traverse portions of a relict point bar (Figs. 7 and 8).

Sections A-A' and E-E' illustrate the relationships between the alluvial fill and the Prairie Formation (Figs. 3 and 7). The Prairie Formation was encountered beneath alluvial sediments in five borings (13, 15, 126, 127, 129) of section A-A'. A major unconformity between the Prairie Formation and the alluvial fill is indicated by an abrupt lithologic change and scour at the contact between the two sedimentary units. The alluvial fill contacts the Prairie Formation at 2.0 to 2.4 m (6-8 ft) in four of the five borings, thus indicating that dissection progressed at least to this elevation prior to lateral planation of the valley floor when base level was at or near this position. However, all borings which penetrate this contact are along the eastern edge of the valley fill. It is possible that the base of the lower sandy facies is lower than the observed elevations near the center of the valley.

The upper surface of the sandy facies generally conforms to the land surface. The sandy facies is at higher elevations beneath topographic ridges than beneath topographic lows. The undulations of the upper surface of the sandy facies is best illustrated in sections A-A' and B-B' which

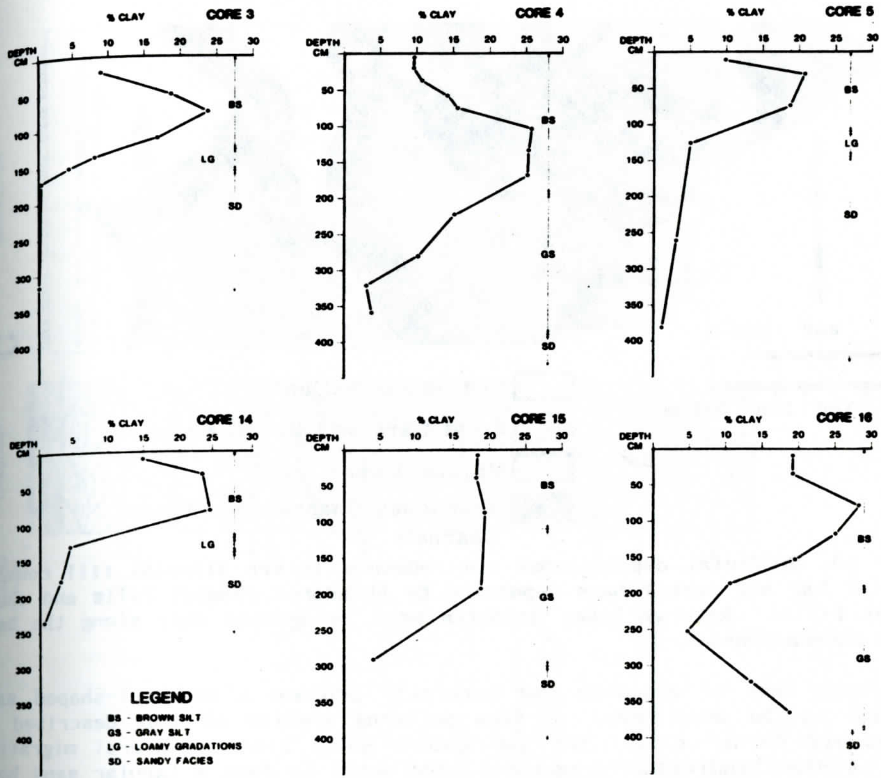

Figure 9: Percentage clay versus depth below land surface for selected cores.

transect scroll bars and chute channels (Figs. 3 and 4). The upward convexity of the top of the sandy facies is more gentle in sections across point bars than across scroll bars (Figs. 7 and 8).

The silty facies completely buries the sandy facies. The brown silt is found beneath bar and channel topographic positions, but it thins and thickens upredictably. Variations in brown silt thickness probably result from variations in the original thickness of the deposit and possibly from local scour and erosion. The gray silt is typically found beneath topographic lows and may overlap the edges of point bars and scroll bars. The lenticular to V-shaped geometry of gray silt bodies indicates that they formed as fills of abandoned channels. It is not certain whether the gray silt has this type of geometry at a distance from the study area. Gray silt may be tabular in areas where lateral accretion topography is not well developed.

SEDIMENTARY DEPOSITIONAL ENVIRONMENTS

Analysis of the characteristics of sedimentary facies in the study area permits interpretations of sedimentary depositional environments (Fig. 10). Environments recognized in the lower sandy facies include channel lags, point bars, and scroll bars. Environments recognized in the upper silty facies include abandoned channel fills of both thalweg and chute channels and natural levees.

The lower sandy facies developed primarily by deposition of river's traction load. The coarsest sediment is found near the base of the sandy facies and below abandoned thalweg channels. This deposit probably accumulated as a channel lag in scour pools. Point bar deposits develop on

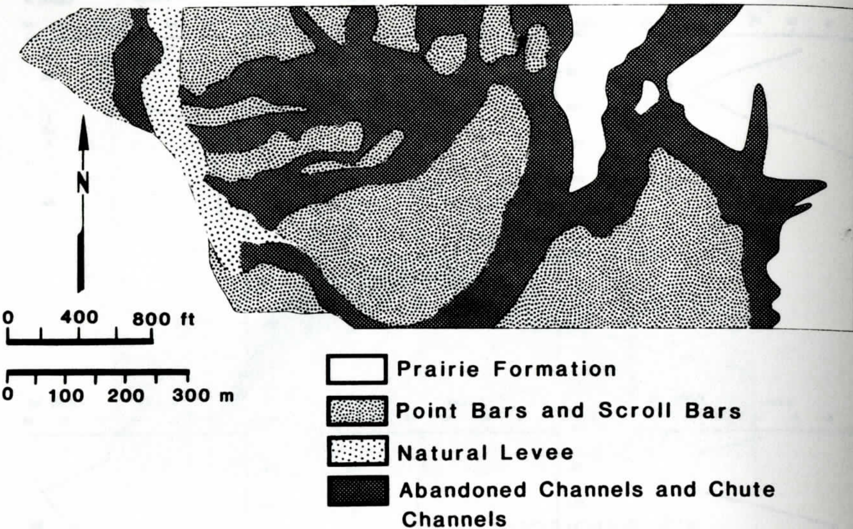

Figure 10: Surficial depositional environments in the alluvial fill consist of point bar and scroll bars separated by abandoned channel fills and chute channel fills. Natural levee deposits were recognized only along the bank of the modern channel.

the convex bank of meanders and generally produce a crescent-shaped sand body due to the development of flow patterns similar to that described by McGowen and Garner (1970) for the modern Amite River. Lateral migration tends to stack individual crescentic sand units to form a tabular sand body during development of the lower sandy facies. Point bars of the Amite River typically deviate from the classical fining-upward sequence. Relict point bars display variations in grain size and truncation of the complete fining-upward sequence, features analogous to modern point bars described by McGowen and Garner (1970) and Autin and Fontana (1980). Observed sedimentary sequences of the lower sandy facies are most similar to the lower sandy bar and upper bar subfacies (Fig. 11 a-b) of Autin and Fontana (1980). Chute bar subfacies (Fig. 11c) do not appear to be as common a subfacies as lower sandy bar and upper bar subfacies.

Immediately behind major point bars and along the convex bank lies an area of scroll bars which are separated by chute channels. Scroll bars are formed by stage variations that accompany flood events (Hickin, 1974). During flood events, a bar develops on the convex side of a meander and is left exposed as a ridge when the flood recedes. During later flood events, swales between scroll bars are utilized as chute channels for overbank flow. When a meander overextends itself, larger chute channels are likely to cut off the bend and become the thalweg channel.

In the study area, chute channels occur as multiple-channeled portions of the fluvial system activated at flood stages when secondary currents separate from the thalweg and pass behind a point bar. Currents which readily shift position across the floodplain follow existing patterns of scroll bar and chute channel topography. The resulting floodplain morphology is similar to the morphology in the channels of braided rivers. Channel and island morphology in streams such as the Donjek River, the Yukon (Williams and Rust, 1969), and the Brahmaputra River (Coleman, 1969) resembles the chute channel and scroll bar morphology of the Amite River floodplain. Because the sediment transport competence behind the point bar is probably lower than the competence in the thalweg, braided morphology does not dominate the alluvial landscape; it is only superimposed on the basic meander belt pattern.

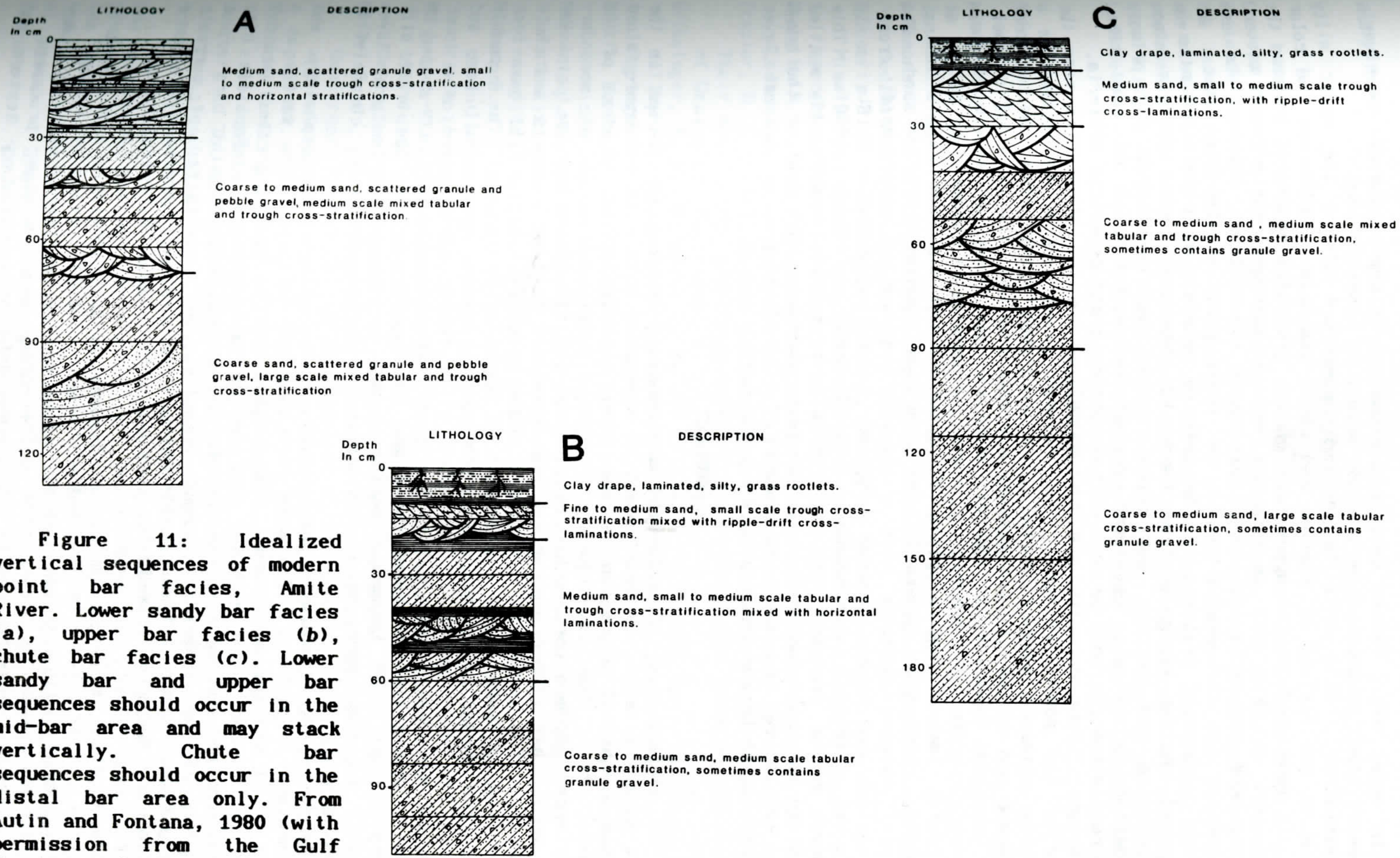

Figure 11: Idealized vertical sequences of modern point bar facies, Amite River. Lower sandy bar facies (*a*), upper bar facies (*b*), chute bar facies (*c*). Lower sandy bar and upper bar sequences should occur in the mid-bar area and may stack vertically. Chute bar sequences should occur in the distal bar area only. From Autin and Fontana, 1980 (with permission from the Gulf Coast Association of Geological Societies).

The upper silty facies developed primarily by deposition of the river's suspended load, accompanied by a minor amount of traction load sedimentation. Gray silt is commonly preserved within the upper silty facies by the filling of thalweg and chute channels. Oxidation and color mottling developed in the upper part of the gray silt sequence, whereas the lower part of this unit is commonly reduced and contains organic matter. Wetting and drying cycles which accompany water table fluctuations are responsible for the weathering of the upper part of the gray silt. No obvious lithologic criteria can be used to separate gray silt deposits in abandoned thalweg channels from those in abandoned chute channels since bioturbation and wetting and drying cycles have destroyed any primary sedimentary structures developed during sedimentation. At this time, differentiation can only be made with morphologic and geometric criteria.

Natural levee deposits are best developed in the brown silt near the banks of the modern channel. The natural levee can be recognized as a slightly elevated ridge 0.3 to 0.6 m (1-2 ft) higher than the adjacent landscape and drapes all sedimentary environments near the existing channel. The levee is about 50 m (150 ft) wide and parallels the modern channel. Sediments typically exhibit primary sedimentary structures, such as horizontal laminations, ripple-drift cross-laminations, and ripple laminae in-phase. Texture is generally silt with some interbeds of fine sand. The textural changes coincide with a stacked sequence of ripple-drift cross-laminations and ripple laminae in-phase, indicating repeated shifts in the ratio of traction to suspended load (Reineck and Singh, 1980). This sequence grades downward into the typically bioturbated and faintly mottled lithology of the brown silt. It is quite possible that most of the brown silt (exclusive of abandoned channel fills) developed as natural levee deposits which became homogenized by pedogenesis.

SUMMARY

The distribution of alluvial landforms and sedimentary facies in the Amite River reflects its development as a meander belt with a coarse-grained bedload. Lateral accretion is the primary constructional process of the alluvial fill; the development of chutes and cutoffs is of secondary importance. Common topographic features on the floodplain include point bar and scroll bar ridges and active and abandoned thalweg channels and chute channels. Two principal lithologic members of the valley fill have been identified, a lower sandy facies and an upper silty facies. The commonly stratified lower sandy facies is composed of point bar, scroll bar, and channel lag environments. The upper silty facies is composed of a lower gray and and upper brown silt. The gray silt is lenticular to V-shaped in geometry and fills abandoned chute and thalweg channels. The brown silt drapes all alluvial landscape positions and is composed of natural levee deposits and the upper portion of abandoned chute and thalweg channel-fill deposits.

Active and abandoned point bars have a similar morphology and sedimentary properties. The plan form geometry of point bars varies as a function of meander curvature. Sand bodies tend to have a greater vertical stacking inside loops with a low radius of curvature. Particle size variations and patterns of sedimentary structures within point bars are a function of water depth, current velocity, and orientation of currents during flood events. These similarities in active and abandoned point bars indicate a similarity in sediment load and hydrodynamic characteristics.

At this time, differentiation of abandoned thalweg channels and abandoned chute channels can only be made with morphologic and geometric criteria. Differences in the sedimentologic characteristics of thalweg and chute channel fills are not obvious based on field criteria alone. It may be possible to better define the differences in sedimentary processes in these types of channels with additional data on sedimentary textures and

analysis of microscale sedimentary structures.

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KARST TOPOGRAPHY OF CENTRAL GREENBRIER COUNTY, WEST VIRGINIA

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ABSTRACT

A mature karst topography is developed on marine limestones of Middle Mississippian age in central Greenbrier County, West Virginia. This study examines the surface karst features of the area in relation to the structural, stratigraphic, and hydrologic setting. Geologic structure controls the pattern of many surface and subsurface streams in the area. Strike-oriented reverse faults, however, seem to have little topographic expression except that caused by stratigraphic displacement or associated folding. The outcrop belt of the Greenbrier Group is bounded on the east and west by outcrops of clastic rocks; differential erosion is an important factor in the overall formation of the karst plain, the setting of blind valleys (capture points of drainage originating on the clastics), and the presence of "contact caves" which are perched on the underlying shale along the eastern boundary. Differential erosion among the limestone units within the Greenbrier Group determines whether the land surface is a smooth plateau or steep and karren-covered. Deep strike-oriented sinkholes along the Pickaway-Taggard-Denmar contacts are an expression of the differing resistance to erosion of the limestone units involved.

INTRODUCTION

A mature karst topography is developed on marine limestones of the Middle Mississippian Greenbrier Group in central Greenbrier County, West Virginia. The Greenbrier Group is approximately 250 meters thick, strikes about N25^{O}E, and generally dips less than 10 degrees to the northwest (Figures 1 and 2). Differential erosion of the Greenbrier Group and the clastic rocks bounding it has resulted in the formation of a broad karst valley parallel to regional strike. Scattered ridges and knobs capped by these clastics rise above the sinkhole-dotted karst plain. Drainage originating on the clastic rocks and flowing into the karst valley is captured by subsurface conduits. This process has resulted in the formation of numerous blind valleys and causes surface drainage on the karst plain to be very sparse.

Many of the karst features associated with the Greenbrier Group, particularly the cavern systems, have been well known for years (Davies, 1958). Recently, however, detailed geologic mapping and field observations made while working on a Ph.D. dissertation at West Virginia University (Heller, 1980) have supplied new information. Consequently, a more precise interpretation of some of the karst features is now possible. This paper describes the relationships between the karst landforms and the stratigraphy and structure of the exposed rock units.

STRUCTURAL INFLUENCES

The Greenbrier Valley lies in the poorly defined border area between the Valley and Ridge physiographic province to the southeast and the Appalachian Plateau to the northwest. The geology shows some structural characteristics of both provinces. Previously unmapped, strike-oriented reverse faults, most dipping southeast, displace strata that are otherwise relatively undeformed except for broad, asymmetric, northeast-southwest-trending folds of low plunge (Figure 2).

Regional strike controls the general trend of the Greenbrier Valley, outcrop pattern, and surface topography. This is primarily a result of differential erosion and is characteristic of the Appalachians. Strike is evident even on low-altitude aerial photographs by the orientation and

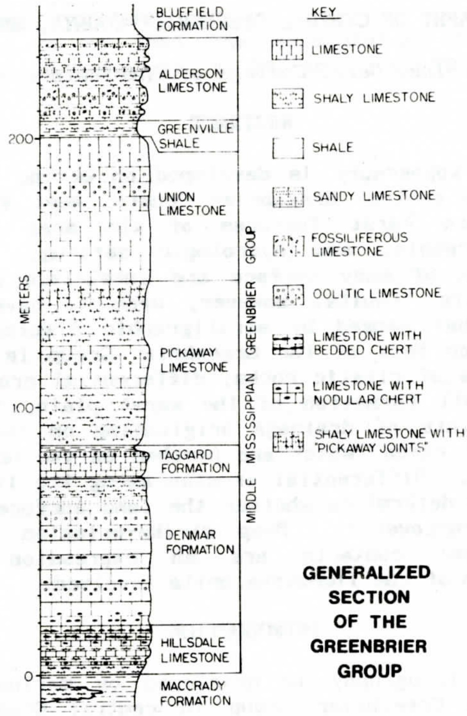

Figure 1: Generalized stratigraphic section of the Greenbrier Group in Greenbrier County, West Virginia.

alignment of sinkholes, roads, and property boundaries, and by the trends of surface streams (Figure 3).

Surprisingly, with a few exceptions there is little direct topographic expression of faults and folds. In a few cases, prominent linear dry valleys as much as 8 kilometers long and 60 meters deep extend along strike parallel to the structural grain. Two such valleys are shown in Figure 4. Dry valleys tend to develop along steeply dipping strata associated with folds or reverse faults. Steeply dipping strata allow water to move downward more freely than horizontal strata because water can move more easily parallel to bedding planes than across them. This allows rapid removal of material through the subsurface and promotes valley development.

In a few cases where a velley has developed adjacent to a reverse fault, the fault crops out along the side of the valley and does not of itself seem to promote erosion. Two caves adjacent to faults show little or no tendency to follow the fault zone directly, although the passages terminate abruptly against the faults (Figure 5). Other cave passages in the area follow faults (Palmer, 1974). Such faults, however, have smaller displacements and no surface expression, and are, therefore, generally not mappable. This suggests that the large-scale faults have little influence on surface and subsurface drainage, other than that caused by stratigraphic displacement or associated folding. The smaller scale faulting visible at places in the cave passages can, however, locally control passage development.

BLIND VALLEYS

Surface streams originating on the clastic rocks are low in dissolved carbonates; therefore, they are readily able to promote solutional openings in the limestone. This allows some or all of the surface flow to escape underground. Complete subsurface capture of these streams has resulted in

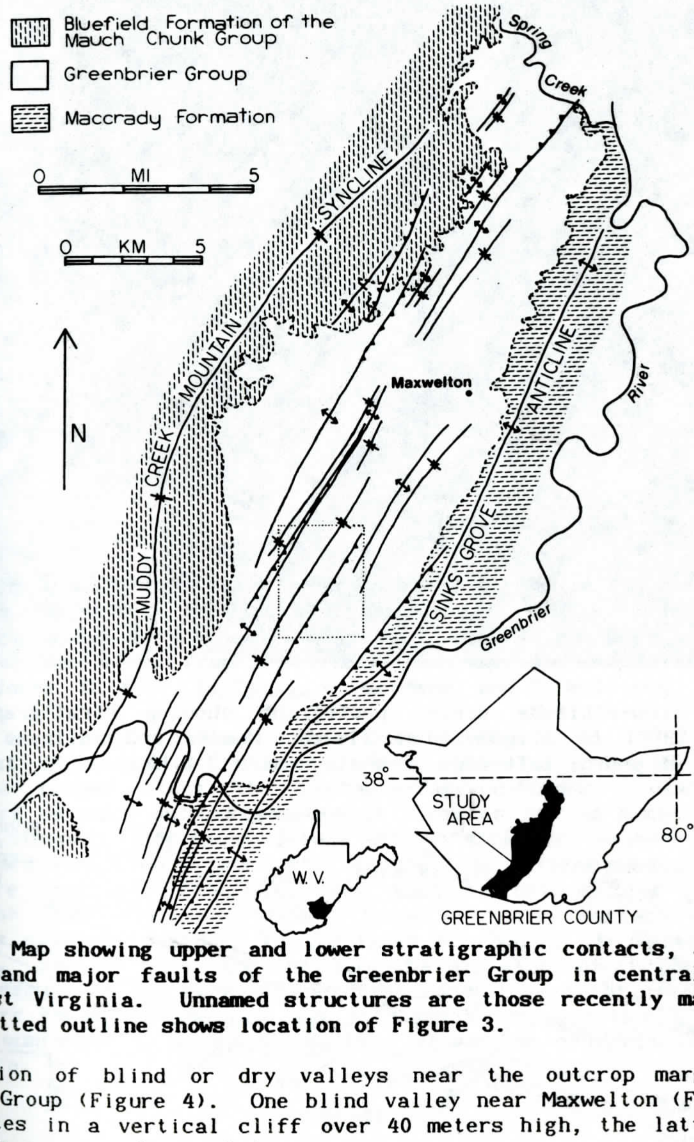

Figure 2: Map showing upper and lower stratigraphic contacts, axial traces of folds, and major faults of the Greenbrier Group in central Greenbrier County, West Virginia. Unnamed structures are those recently mapped by the writer. Dotted outline shows location of Figure 3.

the formation of blind or dry valleys near the outcrop margins of the Greenbrier Group (Figure 4). One blind valley near Maxwelton (Figures 2 and 4) terminates in a vertical cliff over 40 meters high, the latter of which towers above a cave entrance below.

Several streams flowing on the Maccrady shale at the basal contact of the limestone can be followed through laterally extensive cavern systems, known as contact caves; such caves are some of the longest in the United States. One of these, Organ Cave, has been known since the Civil War and has over 58 kilometers of mapped passages (Stevens, 1980). Caverns resulting from the capture of streams originating on clastic rocks of the Upper Mississippian Mauch Chunk Group above the limestone tend to have steeper gradients. This is due to the greater thickness of permeable limestone available and the large height of the surface above base level in this region (Figure 4).

Because of the rapid subsurface capture of streams and the high permeability of the karst plain, surface drainage in the Greenbrier Valley is very sparse. This presents a frustrating problem to land owners of the area, who in spite of over 80 centimeters of annual rainfall must obtain their water from great distances, invest in expensive (deep) wells, or plug

Figure 3: Low-altitude aerial photograph showing stratigraphic strike (locally N 29^{o}E) by alignment of fields, roads, and valleys. North is toward right of photo; left edge of photo covers 3.0 km.

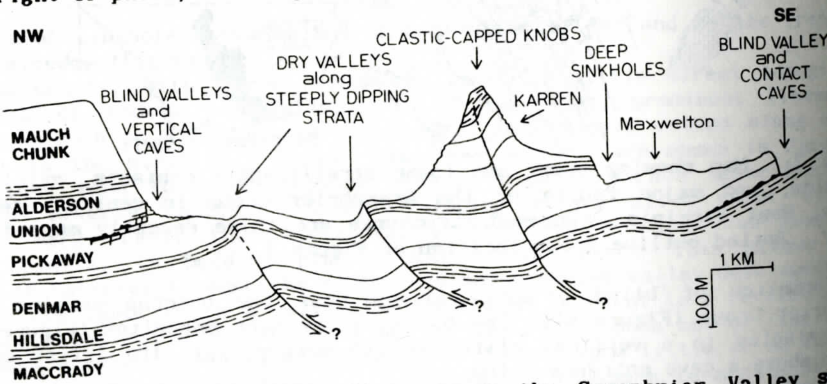

Figure 4: Schematic cross section across the Greenbrier Valley showing major karst landforms discussed. Scale shown is approximate.

up sinkholes to form ponds as a water supply. Most of the sinking streams return to the surface at large springs close to base level along Spring Creek and the Greenbrier River (Jones, 1973).

DIFFERENTIAL SOLUBILITY

Within the Greenbrier Group itself, some members are much more soluble than others. The Pickaway and Hillsdale Limestones (Figure 1), which have more insoluble detritus and chert, weather to form thick soils and smooth stable uplands that are excellent for farming. The more soluble rocks of the Union Limestone and the Denmar Formation form steeper sinkhole-pitted slopes with abundant exposures and thin soil cover, because material can be

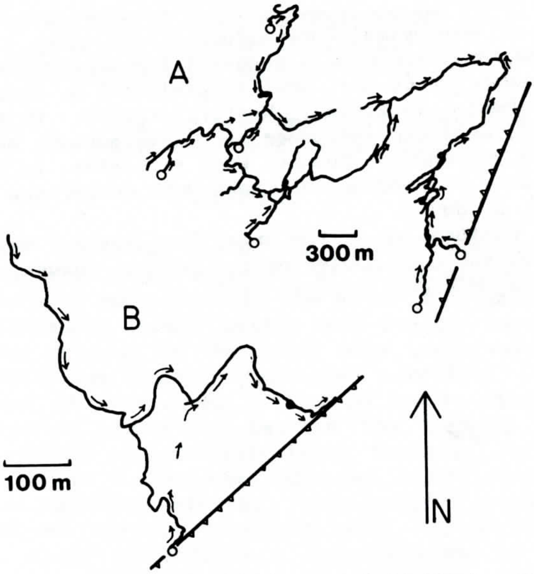

Figure 5: Maps of A) Culverson Creek Cave and B) Taylor Falls Cave showing the location of reverse faults. Arrows show flow directions of streams; entrances are shown by the small circles. Map A) has been modified from Jones (1973); map B) is from a November 1977 tape and compass survey by S. Heller, W. Jones, G. Kyle, D. Scott, D. Goldman, and C. Williams.

transported more readily through subsurface conduits.

Southeast of Maxwelton, elongate, deep sinkholes oriented parallel to strike have formed along the Pickaway-Taggard-Denmar contacts. Some sinkholes are as much as 1 kilometer in diameter and as much as 60 meters deep. The steepest (northwest) slope of the sinkholes exposes all or most of the Taggard Formation (Figure 4). Because the Pickaway and Taggard units contain more insoluble detritus, they tend to act as cap rock. This protects the less resistant Denmar Formation from surface erosion and concentrates solution features in the upper section of the Denmar, close to the contact. Surface water seeping down the dip slope of the Denmar is thus able to undercut the cap rock via solution conduits. In a sense, these deep sinkholes are formed in the same way as blind valleys, although none of the sinkholes terminates a permanent surface stream or contains an accessible cave entrance.

SUMMARY

The presence, location, and orientation of karst landforms, such as dry and blind valleys, contact caves, sinkholes, and karren, are strongly influenced by structural grain and differential erosion. Strike-oriented reverse faults in the region seem to have little topographic expression except that caused by stratigraphic displacement or associated folding.

ACKNOWLEDGEMENTS

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STRATIGRAPHY AND SOIL DEVELOPMENT IN UPLAND ALLUVIUM AND COLLUVIUM NORTH-CENTRAL VIRGINIA PIEDMONT

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ABSTRACT

A two-meter-deep pipeline ditch dug across north-central Virginia exposed saprolite, bedrock, and deposits of alluvium and colluvium. The colluvial mantle averaged approximately one meter deep except where it more than doubled in thickness along active and relict toeslopes. Shallow phyllitic bedrock was present in the ditch only under valley bottoms along large streams. Colluvium sharply truncated saprolitic structures which seldom showed evidence of soil creep. Alluvium lay in channels on erosional benches cut at several elevations between the highest divides and the floodplains. On certain intermediate-level benches, variations of soil color and percent clay in B horizons indicated significant age differences between alluvial and colluvial strata. Stratigraphic evidence of episodic colluviation existed at several locations with little indication of the timing or cause of slope instability. Several lines of evidence suggest that only a few tens of meters of relief formed in the study area during the Pleistocene.

INTRODUCTION

The stratigraphy of surficial deposits exposed in an excavation made by Colonial Pipeline Corporation across north-central Virginia proved to be surprisingly complex. This report outlines the distribution and stratigraphy of saprolite, alluvium and colluvium along the reach examined and the degree of soil development in each of these materials. Comparisons of these geologic and pedologic data with geomorphic models developed for surficial deposits in other Piedmont settings (*e.g.*, Overstreet and others, 1968; Eargle, 1977; Dunford-Jackson, 1978; Darmony and Foss, 1982; Costa and Cleaves, 1984) demonstrate the usefulness of saprolite, colluvium, and alluvium in deciphering the history of a generally eroded landscape.

EXPOSURES AND THE GEOLOGIC SETTING

Pipeline construction through the study area often paralleled the strike of regional Piedmont structures. Where dug mechanically, the pipe ditch had vertical sides 2.0 meters deep and 1.5 meters apart. Shallow bedrock required blasting which left wider and deeper excavations. At any one location, the ditch was open approximately three weeks before the pipe was buried. Because of the time constraints, this study was confined to 22 kilometers of exposures within Culpeper and Orange counties (Figure 1). In the study area the route crossed a well-dissected upland underlain by light grey-to-tan phyllite (Wissahickon schist of Calver, 1963). Triassic sedimentary and volcanic rocks lie less than six kilometers to the west in the Culpeper basin; Cretaceous Coastal Plain deposits cap hills thirty kilometers to the east (Calver, 1963). Although phyllite was the only weathered bedrock exposed in the ditch, saprolite formed from small intrusions of felsic and mafic rocks commonly produced significant changes in soil texture and color over very short lateral distances. Although no faulting was apparent from surface features or displaced beds, slickensides present along some of the many high-angle fractures visible in both bedrock and saprolite may indicate recent fault activity.

Over 60 meters of relief exist where the Rappahannock and Rapidan Rivers cross the phyllitic upland in the study area. Third- and fourth-order stream

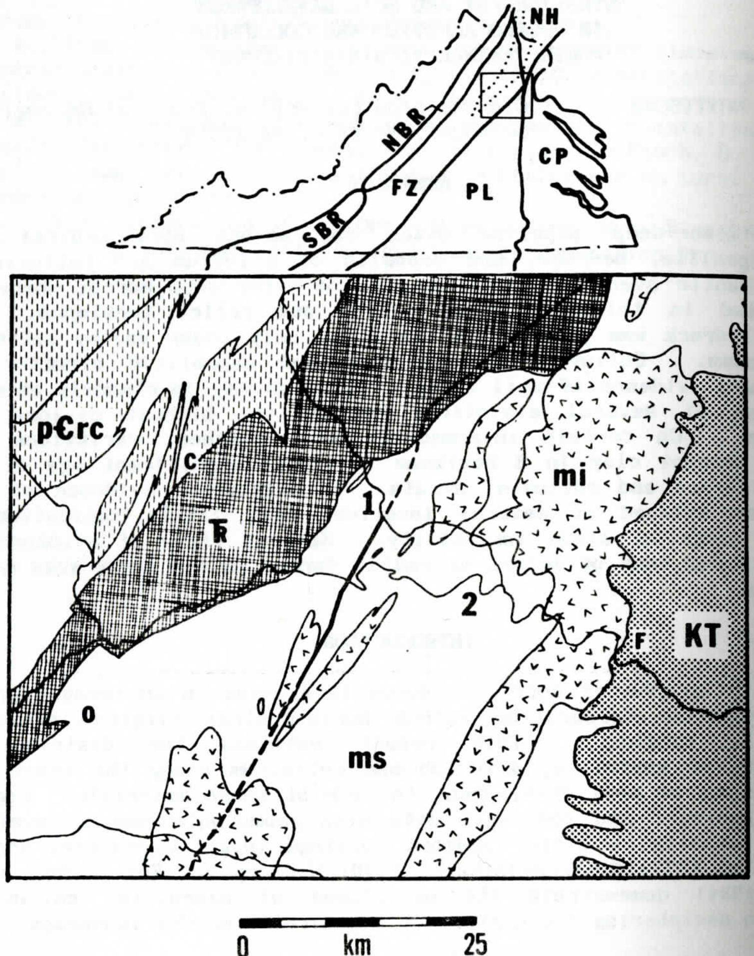

Figure 1. Map of study area showing path of pipeline excavation (dotted line), rivers, major towns, and generalized geology (from Calver, 1963). (1 = Rappahannock River; 2 = Rapidan River; C = Culpeper; F = Fredricksburg; O = Orange; ms (white) = metamorphosed sedimentary rocks; mi= metamorphosed igneous rocks; $P\epsilon_{rc}$ Robinson River and Catoctin formations; ^{T}R = Newark group sedimentary and igneous rocks; KT = Cretaceous and Tertiary sediment). Solid line along pipeline route indicates study area. Location map shows relationship of the area on the map to the physiographic provinces in Virginia (After Hack, 1982).

networks have thoroughly dissected the upland with a relatively high drainage density (8 km per sq. km) so that the resulting landscape has numerous short, steep slopes. Flat-topped terraces exist up to 20 meters above the level of most Rappahannock basin streams, according to Dunford-Jackson (1978).

METHODS

During an initial investigation of the excavation, I identified 23 locations with unusually thick colluvium or alluvium that merited closer study. At eleven of these sites, I took samples, made profile descriptions using standard soil science techniques (Table 1), and constructed stratigraphic sections (e.g., Figure 2). In the laboratory, sand-silt-clay

Table 1. Examples of soil profiles developed on divides, terraces, and floodplains.

HORIZON	DEPTH	DESCRIPTION	PERCENT OF SAND-SILT-CLAY <2 mm FRACTION
ALLUVIUM ON INTERSTREAM DIVIDE (Site 5, No. 2; Elevation 104 m)			
Ap	0- 20	Red (2.5YR4/6)[a] clay; weak medium granular; friable; pH 5.1; abrupt smooth boundary	7-40-53
B22t	20- 40	Red (10R4/8) clay; moderate medium subangular blocky; firm; pH 4.8; continuous thick clay films; clear wavy boundary; alluvium	8-40-53
B23t	40- 90	Red (2.5YR4/8) clay; moderate coarse subangular blocky; pH 5.2; firm; continuous thick clay films; gradual irregular boundary; alluvium	15-36-49
II B3t	90-120	Red (2.5YR4/8) stony sandy clay loam; weak medium subangular blocky; friable, pH 5.1; discontinuous wavy sandy clay loam lens from 95-105 cm depth; many subangular to rounded quartz pebbles and cobbles; many thick clay films; clear wavy boundary; alluvium	49-21-30
III B3t	120-200+	Red (2.5YR5/8) silt loam; weak fine subangular blocky; very friable; common yellow (10YR7/8) mottles; common thick clay films; pH 5.3; saprolite	23-61-16
ALLUVIUM ON TERRACE (Site 4, No. 7, Elevation 77 m)			
Ap	0- 32	Dark greyish brown (10YR4/2) clay; weak medium granular to platy, friable; pH 5.2; a few subrounded quartz pebbles; abrupt smooth boundary	20-34-46
B22t	32- 80	Brownish yellow (10YR6/6) clay; strong medium subangular blocky; extremely firm; pH 4.6; a few red (10R4/8) and yellowish red (5YR5/6) mottles; a few subrounded quartz pebbles and cobbles; continuous very thick clay films; diffuse irregular boundary; alluvium	23-21-56
B23t	80-155	Brownish yellow (10YR6/6) clay; strong coarse angular blocky; extremely firm; pH 4.4; many red (10R4/8) and pale yellow (2.5Y8/4) mottles; a few subrounded quartz pebbles and cobbles; continuous very thick clay films; abrupt irregular boundary; alluvium	31-25-44
II B3t	155-185	Strong brown (7.5YR5/6) loam; weak medium subangular blocky; friable; pH 4.3; common red (10R4/8) and pale yellow (2.5Y8/4) mottles; a few rounded quartz cobbles; many thick clay films; abrupt smooth boundary; alluvium	43-25-32
II B3tg	185-210+	Yellowish brown (10YR5/6) sandy clay loam; weak fine-subangular blocky; friable; pH 4.7; common white (2.5Y8/2) mottles; a few subrounded quartz pebbles; many thick clay films; alluvium	50-21-29
RELATIVELY OLD COLLUVIUM OVER ALLUVIUM ON TERRACE (Site 4, No. 4; Elevation 79 m)			
Ap	0- 32	Yellowish brown (10YR5/4) clay; weak medium granular to platy; friable; pH 5.0; abrupt smooth boundary	23-34-43
B1t	32- 65	Reddish yellow (7.5YR6/6) clay; weak fine subangular blocky; firm; pH 4.9; common red (2.5YR4/6) mottles; a few subangular quartz pebbles; many thick clay films; clear wavy boundary; colluvium	21-46-43
B2t	65-109	Strong brown (7.5YR5/6) silty clay; moderate medium subangular blocky; firm; pH 4.8; many red (2.5YR4/6) and pale red (10R6/4) mottles; common subangular quartz pebbles; many thick clay films; clear wavy boundary; colluvium	10-50-40
RELATIVELY OLD COLLUVIUM OVER ALLUVIUM ON TERRACE, continued			
B3tg	109-160	Strong brown (7.5YR5/6) silty clay loam; moderate coarse subangular blocky; firm; pH 4.8; many pale red (10R6/4) and white (10YR8/1) mottles; common subangular quartz pebbles; many thick clay films; clear wavy boundary; colluvium	18-44-38
B3g	140-180	Strong brown (7.5YR5/6) stony silt loam; moderate coarse subangular blocky; firm; pH 4.9; many subangular quartz pebbles; many thick clay films; clear irregular boundary; colluvium	20-76-40
II B2tb	180-210+	Red (10R4/8) clay; strong medium angular blocky; extremely firm; pH 4.1; many brownish yellow (10YR6/6) mottles; continuous very thick clay films; alluvium	16-28-56
RELATIVELY YOUNG COLLUVIUM OVER ALLUVIUM AND SAPROLITE ON TERRACE (Site 4, No.8; Elevation 75			
Ap	0- 21	Yellowish brown (10YR5/4) silt loam; weak medium granular; friable; pH 4.7; few subrounded quartz pebbles; clear wavy boundary	25-54-21
A2	21- 40	Very pale brown (10YR8/4) loam; weak fine platy to massive; firm; pH 5.2; a few subrounded quartz pebbles; clear wavy boundary; colluvium	31-47-22
B21t	40- 75	Light yellowish brown (10YR6/4) stony clay; moderate medium subangular blocky; firm; pH 4.9; many subangular quartz pebbles; abrupt wavy boundary; colluvium	23-28-49
II B22t	75- 93	Light brown (7.5YR6/4) clay; strong medium angular blocky; firm; pH 4.9; common subangular quartz pebbles; many thick clay films; abrupt wavy boundary; stony alluvium	28-16-56
III B22t	93-134	Reddish brown (5YR5/4) clay; strong medium subangular blocky; very firm; pH 4.6; common light yellowish brown (10YR6/4) mottles; continuous thick clay films; abrupt irregular boundary; alluvium	34-17-49
IV B3t	134-155	Reddish brown (5YR5/2) stony, sandy clay loam; weak fine subangular blocky; friable; pH 4.8; many rounded to angular pebbles and cobbles of gneiss, schist, quartzite, greenstone, red shale, basalt; continuous thick clay films; abrupt irregular boundary; alluvium	49-25-26
V B3t	155-200+	Strong brown (7.5YR5/7) silt loam; weak fine subangular blocky; friable; pH 4.7; common thick clay films; saprolite	22-64-14
ALLUVIUM ON VALLEY BOTTOM (Site 12, No.9; Elevation 68 m)			
Ap	0- 10	Light brownish grey (10YR6/2) sandy loam; weak fine granular; very friable; pH 5.2; abrupt smooth boundary; alluvium	53-31-16
C	10-210	Yellow (10YR7/6) to yellowish brown (10YR5/4) sandy clay loam to silt loam; massive; firm; a few rounded cobbles; pH 5.0; abrupt wavy boundary; alluvium	67-10-23 to 41-54- 5
II C	210+	Very dark greyish brown (10YR3/2) gravelly sandy loam; massive to granular; loose; pH 4.9; continuous rounded pebbles and cobbles of gneiss, quartzite, basalt, red shale, and greenstone; alluvium	78-11-11

[a]All Munsell colors are for moist samples

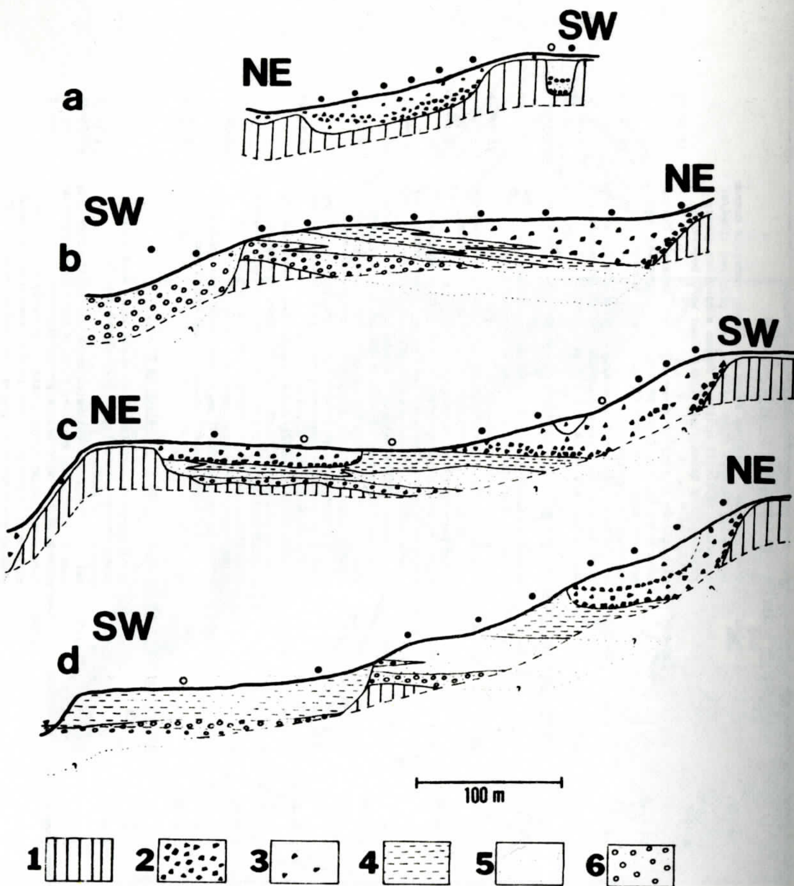

Figure 2. Schematic cross-sections showing stratigraphy of surficial sediments at four sites (1 = saprolite; 2 = stony silty colluvium; 3 = stone-poor silty colluvium; 4 = fine-grained alluvium; 5 = sandy alluvium; 6 = sand and gravel alluvium.) Dots denote position of measured soil profiles and stratigraphic sections; open dots, profiles described in Table 1. Exposures are 2 meters in height.

a. Site 5, 2.9 km southwest of Rappahannock River; 108 to 100 m elevation.
b. Site 21, 7.6 km southwest of Rapidan River; 113 to 100 m elevation.
c. Site 4, 1.0 km southwest of Rappahannock River; 97 to 84 m elevation.
d. Site 12, 0.5 km northeast of Rapidan River; 97 to 65 m elevation.

ratios were measured with hydrometer techniques and by sieving at half-phi intervals. Water pH values were obtained with a Corning probe.

DISTRIBUTION OF SURFICIAL MATERIALS

Saprolite

Saprolite overprinted with a pedogenic solum comprised roughly 20% of the ditch exposures. Saprolite is the isovolumetric residuum formed at shallow depths by geochemical alterations of easily weatherable rocks; further pedologic alterations active close to the surface form the solum by decomposing even the most resistant primary minerals (Newell and others, 1980). Under the highest, gently sloping divides, residual soils (solum plus saprolite) were often thicker than the exposures were deep (2 meters) and showed no visible stratification or stonelines. On particularly steep slopes where geologic and cultural erosion had removed the solum, saprolite lay at the surface. Encroachment of the clay-enriched solum into the low-density saprolite occasionally followed planes of weakness such as foliation and sheet joints that formed before saprolitization. Exposures of slightly weathered or fresh bedrock commonly showed 5 cm thick clay skins filling these cracks as much as 4 meters below the surface.

Shallow bedrock was exposed in the Culpeper ditch only beneath large streams near the base of long or steep slopes descending to those streams. The few strongly convex slope segments in the landscape generally were controlled by the location of shallow bedrock or saprolite thinly blanketed with colluvium and were present on the downhill side of gently sloping benches or terraces close to the valley bottom.

Colluvium

Stony, silty colluvium blanketed more than 60% of the landscape cut by the trench with a very uniform 0.5 to 1.0 meter thickness (Figure 2a). The abrupt base of the colluvium appeared as a change in color, texture, or stoniness that truncated saprolite, bedrock or alluvium. At the bottom of the colluvium and below, few veins, dikes, or foliations showed signs of warping caused by soil creep.

The colluvium was derived principally from phyllite and schist saprolite and from quartz veins. Within a colluvial mass, angular blocks of phyllite were present only near the saprolite-colluvium interface at the uphill end of the deposit. Pebbles of subangular quartz less than 5 cm in diameter studded the fine-grained matrix and appeared in stone lines extending up to 10 meters in thin discontinuous beds parallel to the hillslope. Occasionally an irregularly-thick bed of vein quartz fragments lined the base of the colluvium. No buried soils were noted within the colluvium wedges although certain reaches of the exposure did have distinct breaks in colluvial texture (e.g., stone-poor silty colluvium over stony silty colluvium, Figure 2a).

Deposition of colluvium at the sites studied was both episodic in time and migratory in location. Where the pipeline route crossed the paths of first-order valleys, shallow channel-fill sequences often appeared in the exposures. At one site (Figure 2c) two small channels were filled with massive colluvium that contained fewer angular quartz pebbles and less soil development than the underlying material. Material in channels at another site (Figure 2d) graded laterally between thin, bedded alluvium and massive stony colluvium. In its uphill exposures this colluvium had a silty matrix with abundant angular quartz pebbles throughout and numerous blocks of phyllite near its base. Thin lenses of sand and rounded pebbles composed of non-local rock types were increasingly abundant further downhill so that the sediments in the lower margin of the channel appeared fluvial in charcter. In the top of the channel above this variable unit lay a uniformly massive colluvial bed with a quartz stoneline at its base. The thick deposits below this channel-fill sequence also made an interbedded transition from colluvium to alluvium.

Alluvium and Related Colluvium

Usually thick accumulations of colluvium and alluvium lay on some summits, on terraces and benches, and on concave valley bottoms. At the crests of two major interfluves stony deposits filled channels 1.0 to 1.5 meters deep and no more than 8 meters wide at their tops (Figure 2a). Each deposit was identified as alluvium because of the partially rounded pebbles in thin beds at their bases and the diminishing quantity of sand in successively higher layers above the gravel. Beneath the summits and terraces, the deposits and underlying saprolite were impregnated with pedogenic clay throughout their exposures and stained with reddish iron oxides. Soil survey and geomorphic reports (e.g., Carter and others, 1971; Elder and Pettry, 1975; Dunford-Jackson, 1978) indicate that such divide-capping alluvium is common in the Piedmont of Virginia; some of these deposits lie at elevations considerably higher than encountered in this study. Presumably an exhaustive search for potential hilltop channels would locate more examples than were encountered along this pipeline ditch.

The thickest colluvium and alluvium beds present along side-slopes were most often beneath and uphill from gently-sloping benches 5 to 12 meters above the modern floodplain. In a few locations sideslopes cut across buried benches of saprolite and alluvium without changing slope. Figures 2b-d illustrate the varied stratigraphy encountered in this landscape position. Wedges of reddish-brown clay-rich colluvium commonly buried fining-upwards sequences of alluvium, also heavily weathered. At several sites, small channels cut into these beds were filled with younger, less weathered slope debris. All of these saprolite-supported bench deposits were older and higher than the modern valley-bottom alluvium.

The alluvium beneath sideslopes and benches contained rounded and subrounded pebbles and cobbles of quartz, quartzite, green stone, red shale, schist and gneiss which graded upwards through beds of stone-free sand and silty clay. The thickness of the alluvium varied according to the size of the nearest stream but ranged from 1 to 3 meters. Many of the non-quartz gravel clasts were very friable due to post-depositional weathering; quartz pebbles commonly had iron-rich rinds up to 8 mm thick. In places thin pedogenic clay skins and iron- and magnanese-rich oxides coated alluvial gravels and sands up to 2 meters below the surface.

Where exposed by the pipeline ditch, the deposits in the low terraces and floodplains along major streams consisted of relatively unweathered fluvial sequences. Basal gravels containing vein quartz, quartzite, and gneiss with minor amounts of red shale, greenstone, and basalt lay upon unweathered bedrock. Beds of poorly-sorted sand and silt blanketed the gravels. Although water in the excavation usually obscured the total thickness of these fluvial sediments, some sequences exceeded 4 meters in measured thickness and rose to more than 3 meters above their parent stream.

RELATIVE AGE DATA

The relative ages of surficial deposits are established by stratigraphic relationships, geomorphic positions, and soil development criteria. Depositional and erosional boundaries clearly exposed in the ditch indicate the relative ordering of many geomorphic events. The indices of weathering available for this study (B horizon structures, thicknesses, colors, and clay contents; degree of clast weathering) might suggest the timing of those events.

The principle of geomorphic ascendancy dictates that the hilltop channels were the oldest deposits found in the study area. Formed by small streams which left an armor of relatively insoluable quartz pebbles and sand, the channel-fills now sit in an inverted topographic position. The very high clay content (65%) in the once-sandy alluvium, the reddish-brown to red (2.5 YR to 10R) soil colors (Table 1), and the landscape position of the channels imply a great age for the deposits. B horizon thicknesses were difficult to assess because the topsoil and some subsoil horizons had been eroded.

If this relative age interpretation for the hilltop channels is correct, the alluvium on the sideslopes and benches should be measureably less weathered than the hilltop channel alluvium.. It seems clear that the observed benches of bedrock (saprolite) which lie at 75 meters elevation were once part of a valley floor system (Terrace Level 3 of Dunford-Jackson, 1978), much younger than the higher channels at 104 and 128 meters elevation. With the available soils data, though, it is difficult to distinguish the hilltop alluvium from the terrace deposits. On the sideslopes and benches the full depth of B horizon clay penetration was at least 200 cm but may have been as deep as 400 cm; maximum clay contents were comparable (50-60%) at like depths in similar materials. Although soil colors were somewhat less reddish in the bench deposits (5YR to 2.5YR) than in the hilltop channels, this color distinction may be the result of differing histories of soil drainage conditions. Relative measures of pebble weathering were not useful in establishing relative ages because the hilltop channels contained nothing except vein quartz. Thicknesses of iron oxide rinds on rounded quartz pebbles ranges from 0 to 8 mm in both landscape settings; therefore, due to the lack of time, numerous systematic measurements of these rinds were not taken. Future research may find such data valuable, however.

Even though the soil data on the two sets of weathered alluvium did not suggest marked differences in age, contrasting soil characteristics between the bench alluvium and overlying colluvium did indicate some variations in age. In Figure 2b, the colluvial wedge was weathered to the same degree as the underlying alluvium; similar weathering criteria were found in the colluvial and alluvial facies represented in the oldest strata of Figure 2d and the oldest alluvium of Figure 2c. However, the massive colluvium shown in Figure 2c seemed significantly less weathered than the alluvium at that site. In deposits of similar texture, the alluvium had better developed blocky structure plus more and thicker clay films than the overlying colluvium. The stoneline and clear boundary at the base of the colluvium plus the change in clay content also indicate that the colluvium buried a paleosol formed on the alluvium.

Young colluvium filling the swales cut into these older bench deposits was less weathered than the underlying materials. Not all of the young swale fills were equally weathered at every site. Some of the deposits apparently accumulated in the recent past, judging by their high content of disseminated organic matter in the moderately well-drained sites; others had filled now-relict drainageways presently truncated by swales following new slope directions. All of these younger colluvial masses had more brown hues, lower clay precentages, and thinner B horizons than stratigraphically older deposits at the same site.

Compared to the bench and sideslope deposits, the valley bottom sediments were virtually unweathered. The amounts of clay present rarely exceed 15% even in the finest floodplain sediments; the lack of soil structures suggested that the clay in these deposits was non-pedogenic. Brown colors (10YR) indicated a minor amount of oxidation in these soils but for all practical purposes the gravels were unaltered. The freshness of these class contrasted with the moderately weathered, friable gneiss and basalt pebbles found in older alluvial gravels.

DISCUSSION

Colluvium is widespread in the unglaciated Piedmont (e.g., Cain, 1944; Conley and Drummond, 1965; Overstreet and others, 1968; Eargle, 1977; Ciolkosz and others, 1979; Newell and others, 1980; Darmody and Foss, 1982). Most authors consider the saprolite, which mantles bedrock in the Piedmont, to be the source of the silty matrix and angular stony fragments found in most colluvium, although Darmody and Foss (1982) argued that a significant amount of the silty colluvium in the Maryland Piedmont is reworked loess. Darmody and Foss cited lithologic breaks visible within the colluvium that matched changes in stoniness and clay mineralogy to substantiate the loessal hypothesis. In Culpeper County, Virginia, only a few sites with such lithologic distinctions were noted along the 22 kilometers examined. Although future studies may confirm the reworking of loess into the colluvium in Virginia, the available stratigraphic data do not suggest that the loess was thick or well preserved.

The colluvium cover in the Piedmont is generally thought to be thin especially when compared to 18-meter-thick colluvial wedges in the Inner Coastal Plain of Georgia (Newell and others, 1980). However, Eargle (1977) reported 3 to 6 meters of colluvium in swales and on sideslope benches in South Carolina while thicknesses of 1 to 30 meters have been cited in Pennsylvania's Piedmont (Ciolkosz and others, 1979). The massive colluvium found during the present study exceeded 2 meters only where it lay upon active or relict toeslopes.

Alluvium is as widespread as colluvium and saprolite in some Piedmont areas (*e.g.*, Eargle, 1977). In north-central Virginia, though, alluvium formed the parent material for soils in only approximately 10% of all ditch exposures; colluvium, 60%. The relatively sparse distribution of alluvium in Culpeper and Orange Counties is interesting given the numerous terrace deposits noted by Dunford-Jackson (1978) in the same areas. This dearth may be due to the large number of steep slopes along the pipeline route caused by the incision and close proximity of Rapidan and Rappahannock Rivers (Stan Dunford-Jackson, personal communication, 1980). These slopes probably induced more stripping of surficial sediments along the narrow divides in the study area than occurrred elsewhere in the region.

Several features found to be common in the South Carolina and North Carolina Piedmont were not present in the Virginia pipeline exposures. Near Spartanburg, South Carolina, Eargle (1977) discovered large numbers of colluvial masses burying both weathered alluvium and mats of organic debris up to 3 meters thick. During the Pleistocene, these depsists filled steep-sided swales carved into the saprolite. The massive colluvium proved resistant to subsequent re-entrenchment and forced valleys to develop in the adjacent saprolite thereby forming benches of valley-fill materials. Surficial materials and stratigraphic situations similar to these in South Carolina also exist along small streams in central North Carolina (Overstreet and others, 1968; Whittecar, 1984). Neither organic mats nor substantial colluvium-armored benches were seen along the ditch in north-central Virginia. As noted in Figure 2 colluvium did bury alluvium along relict toeslopes in north-central Virginia, yet because of the relative thinness of most colluvial deposits when compared to the size of adjacent streams, any effect of the observed colluvial masses upon the location of stream incision is believed to have been minimal. Perhaps the lack of visible organic mats was due to the limited exposure afforded by the pipeline excavation in poorly-drained materials, in contrast to the deep gully walls used by some earlier workers. Most sites examined were sufficiently drained to oxidize any carbonaceous sediments. The only "organic deposits" found along the pipeline route filled broad depressions that occasionally appeared only on one side of the ditch. Artifacts, highly-disturbed stratification, and charred stumps confirmed the reports of local residents that the apparent "channels" were disposal pits dug during construction of the original pipeline in the 1960s.

Many recent studies indicate that changes in soil profile development can indicate the relative age of some landscape features (*e.g.*, Gile, 1970; Bockheim, 1980; Whittecar and others, 1982; Birkeland, 1984; Foss and Segovia, 1984; Mills and Wagner, 1985). However it is very difficult to assign absolute ages to weathered deposits by using soil criteria exclusively (Vreeken, 1984). Rate and types of soil-forming processes on a single hillslope can vary dramatically according to landscape position, drainage, slope angle or parent material. Soils on stable uplands or interfluves will tend to be overprinted with soil characteristics formed under climatic or geomorphic conditions different from the present. Furthermore, under a monogenetic environment the rates of change for many soil characteristics are non-linear and should not be used to calculate soil age without additional relative dating information. Firm chronologic control is only established by correlating a soil with dated sedimentary sequences or well-established geomorphic surfaces. Even with such control, great caution should be used when extending soil-age comparisons away from the area of empirical reference chronosequences to different climatic or geologic situations (Vreeken, 1984). With the potential for so many problems in interpretation, soil age estimates are rarely better than "order of magnitude" in precision.

Any use of soil development data to indicate relative age must be supported by a combination of geomorphic, stratigraphic, sedimentologic, or petrographic information. For example, a clear distinction must be drawn between the age of a geomorphic surface and the sediments and soils underlying that surface (Ruhe, 1969; Daniels and others, 1971). Almost all slopes in the Piedmont are composed of one or more erosion surfaces which can truncate geologic structures, sedimentary sequences and soil profiles. Fortunately, because alluvial point bar deposits exhibit fining-upwards sequences approximately equal in thickness to the depth of the river channel (Wolman and Leopold, 1957), it is possible to estimate the amount of deposit stripped from a given terrace surface. In the study area the degree of erosion on alluvial deposits generally increases with elevation. The youngest valley-bottom sediments showed no apparent erosion; the hilltop channel fills were severely eroded with half or more of the original deposits missing. Alluvial sequences on intermediate-level terraces were nearly complete beneath old colluvial masses close to the valley side but moderately to severely eroded elsewhere. Although the amount of erosion was less than one meter on most of these terrace deposits, the tops of fining-upwards sequences were rarely expressed as low-gradient surfaces. The range of soil development upon a single terrace level can be understood only with knowledge of the site's sediments and stratigraphy.

The study of Rappahannock River basin done by Dunford-Jackson (1978) relied upon elevation analyses and recognition of sedimentary sequences as well as differences in soil development to distinguish six alluvial surfaces present within 120 meters of relief. Increasing soil development was indicated by more red hues, greater clay precentages in the B horizon, and total solum thicknesses. The alluvial benches at 75 meter elevations in the study area correspond to Dunford-Jackson's Terrace Level 3. On surfaces at this elevation, Dunford-Jackson reported up to 80% clay in B horizons with 2.5YR colors Stream deposits on major interstream divides contained up to 90% clay and 10R colors throughout. As indicated in Table 1, clay contents in the hilltop and alluvial deposits described earlier in this paper contained only 50-60% clay, more like soils Dunford-Jackson described on younger, lower surfaces. These differences in soil clay content could be due to variations in parent material or to the erosion of horizons at the sites I studied. Soil colors and pH values found in the pipeline exposures generally match those predicted by Dunford-Jackson (1978) although his soil pH values showed significant decreases at the base of the B horizon that were not found in the ditch profiles.

Several lines of evidence suggest that the deposits on the 75 meter benches may date from at least the early Pleistocene or before. Dunford-Jackson (1978) obtained estimates of terrace age by projecting the Piedmont terrace levels down-valley and correlating them with Coastal Plain terraces. He suggested that the widespread Terrace Level 3 formed at the same time as the "Coharie" terrace, one of several Coastal Plain surfaces thought to be pre-Pleistocene in age (*e.g.*, Hack, 1955). Markewich, Pavich, and Gonzales (1983), who studied soils in dated sandy Coastal Plain deposits, found that thickness of the total soil, thickness of the B horizon, clay content, and various elemental ratios in secondary soil minerals increase with time of soil development. They found that soils in many deposits one million years old were 2-to-4 meters thick and had over 40% clay in the B horizon, whereas 60,000-to-500,000 year old soils contained 15-25% clay. Other studies in Virginia (Pavich and Obermeier, 1977; Pavich and Markewich, 1979; Pavich, in press) suggest that a 4-meter thick mantle of saprolite can form in crystalline bedrock during a million years of landscape stability. Sheet joints and potholes preserved on the saprolite beneath the 75 meter terraces in the present study strongly suggest that saprolite developed after the alluvium was deposited. Comparison of the sola and saprolite on the benches with those described in reports cited above indicates that the bench soils formed over a span of at least 500,000 years and perhaps much longer. Therefore an early Pleistocene date, or even Pliocene as suggested for these surfaces by Dunford-Jackson (1978), seems possible.

In north-central Virginia, bedrock rarely appeared in the 2-meter deep ditch at more than 10 meters elevation above the nearest large stream. The presence of unweathered bedrock only beneath incised valley bottoms in the Piedmont, also noted in Georgia (Newell and others, 1980) and in Maryland (Costa and Cleaves, 1984), indicates that stream erosion has removed regolith there more rapidly than saprolite can reform. Thus, based upon the rate of saprolite formation mentioned earlier, this active fluvial erosion occurred during the late Pleistocene in the lower portions of major valleys of the Rappahannock River system.

Although some authors expect to find only one generation of soil development in the Piedmont (e.g., Newell and others, 1980), buried paleosols are reported from several locations in the Piedmont and Blue Ridge (e.g., Overstreet and others, 1968; Eargle, 1977; Mills, 1977, 1982). Because a well-developed soil profile only forms upon a relatively stable surface, Ciolkosz and others (1979) concluded that the presence of buried paleosols in colluvium implies episodic periods of slope instability alternating with times of stable slopes. Mills (1977) came to the same conclusion based upon three traceable slope deposits exhibiting different degrees of weathering in the South Carolina Blue Ridge. Using discriminant analyses of weathering phenomenon from 135 sites in that area, Mills (1982) provided further support for repeated episodes of rapid colluviation.

Climatic change is often cited as the reason for such episodic slope instability. Ciolkosz and others (1979) used indirect evidence of widespread periglaciation in Pennsylvania (Marchand and others, 1978) to argue for frost-accentuated solifluction during the Pleistocene. Citing reports of spruce-fir pollen from colluvium and from organic mats buried beneath colluvium, several authors have suggested that cooler Pleistocene climates induced rapid colluviation in the Piedmont of the Carolinas (Cain, 1944; Whitehead and Barghoorn, 1962; Overstreet and others, 1968; Eargle, 1977; also see Pewe, 1984).

In north-central Virginia, the different degrees of weathering between alluvial deposits at several sites and the buried paleosols on some alluvium indicate that along the pipeline traverse more colluvium was deposited during some periods than others. Two explanations are possible for the stratigraphy observed in the ditch. Because of the possible shifts in the positions of valleys and toeslopes in the study area due to progressive landscape erosion, the hillslope colluvium might have changed its accumulation areas through time (e.g., Newell and others, 1980) even while its rate of activity remained relatively constant. If this were true, any arbitrarily-chosen traverse may encounter only parts of the entire colluvial sequence present in the landscape thereby giving an illusion of apparent pulsations in slope stability. The evidence gathered in the present study is insufficient to refute this "steady down-wasting" interpretation. On the other hand, studies in the Piedmont cited above suggest that Pleistocene climate changes strongly affected the rates of hillslope processes in the region. Some observations and conclusions in this report may support an interpretation of the superimposed colluvial wedges as glacial-age deposits. Based upon the soil-development arguments stated earlier, most of the slope deposits are Pleistocene in age. Unfortunately, clear and direct indications of solifluction or periglaciation, such as ice-wedge casts or convolute structures, were absent in the Piedmont pipeline exposures. Future research may reveal other paleoenvironmental indicators useful in establishing the timing of colluviation at the study sites.

SUMMARY

The surficial sediments which mantle the northern Virginia Piedmont landscape are relatively thin except along active or relict toeslopes. Colluvium buries alluvium and well-developed paleosols at many of those sites. Alluvium is often present beneath hillslope benches lying at several levels above the valley bottom but it also exists is patches which have no geomorphic expression.

Differential soil development between deposits at various elevations and on materials at the same level reflects episodic downcutting and subsequent deposition by fluvial and slope processes. The older deposits may be at least early Pleistocene in age based upon both the presence of shallow bedrock only in valley bottoms and the rates of saprolite formation reported by others. Although evidence is mounting from many neighboring areas that cold Pleistocene climates induced relatively rapid mass wasting in the Virginia Piedmont, the causes of the episodic colluviation in the study area are not clearly understood.

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