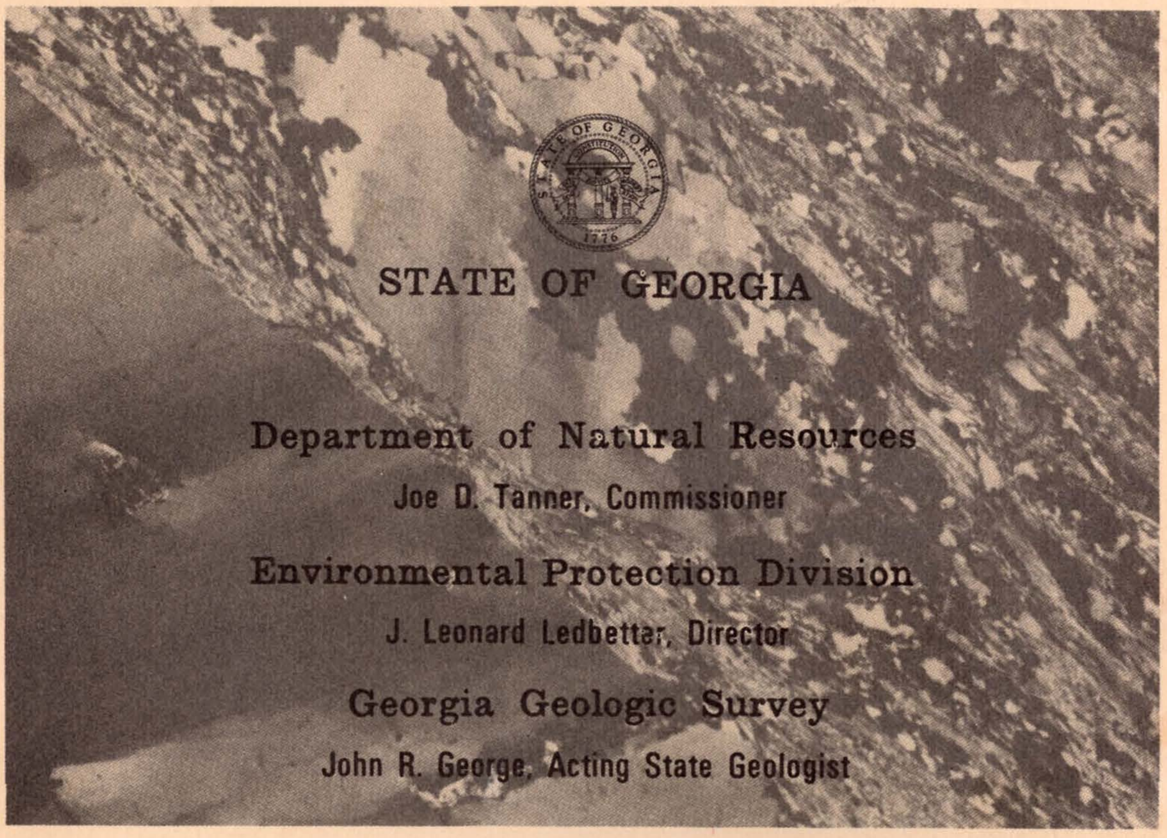
SHORT CONTRIBUTIONS TO THE GEOLOGY OF GEORGIA



ATLANTA
1978

BULLETIN 93

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SHORT CONTRIBUTIONS TO THE GEOLOGY OF GEORGIA

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THE HUBER FORMATION OF EASTERN CENTRAL GEORGIA

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ABSTRACT

The term "Huber Formation" is proposed for all of the post-Cretaceous pre-Jackson strata in the kaolin mining districts of Georgia, northeast of the Ocmulgee River. This term is justified for convenience in referring to a sequence of strata having incompletely known age, but occurring between two recognizable unconformities and constituting a mappable unit. This sequence of strata can be traced from the vicinity of Macon, on the Ocmulgee River, eastward through the Wrens District, near the Savannah River at the east margin of the State.

INTRODUCTION

The purpose of this paper is to name and describe the Huber Formation as a newly designated stratigraphic unit. This stratigraphic unit occupies the position between an unconformity that marks the top of the Cretaceous strata and an unconformity at the base of the Jackson strata. For many years the significance of the unconformity at the top of the Cretaceous was not generally recognized in the kaolin producing area, and all of the sequence of strata occupying the interval between the Jackson strata and the underlying crystalline rocks, in the up-dip portion of the section, was thought to be Cretaceous, or was left undesignated as to age. This sequence is now known to be Tertiary, on the basis of fossil evidence, and a formational name is desirable.

Both the Huber Formation and the underlying Cretaceous strata contain beds of commercial kaolin. Formerly all of the kaolin was believed to be Cretaceous. In the interpretation of drill data, geologists customarily took the uppermost occurrence of white kaolin as the top of the Cretaceous. In surface mapping at many places, the base of the fossiliferous Jackson strata was erroneously taken as the contact between the Tertiary and Cretaceous. Lamoreaux (1946) recognized local occurrences of materials that appeared to be different from either the Cretaceous or the Jackson, and referred to these materials as "channel sands". Some of his "channel sands" very probably represent part of what is now recognized as the Huber Formation. The principal reasons for the desirability of a new formational name is to contribute to the understanding of the stratigraphy and economic geology of the region. The base of the Huber Formation is an unconformity marking the break between Cretaceous (Mesozoic) and Tertiary (Cenozoic) time. Therefore the identification and mapping of the Huber Formation is essential to the understanding of the geology of this part of the Coastal Plain. Recognition of the Huber Formation contributes to the understanding of economic geology. Geologists have long known that kaolin deposits at different geographic and stratigraphic positions have different properties, without knowing the reason. It is now known that these physical differences correlate to a large degree with the age of the kaolin, Cretaceous or Tertiary. Accordingly, recognition of the Huber Formation will contribute to the solution of some of the prospecting, evaluation, and mining problems.

The portion of Georgia referred to in this paper is roughly the eastern half of the central part of the State. It is the inner part of the Coastal Plain, extending from the Ocmulgee River on the west to the Savannah River on the east. The area in which the Huber Formation crops out includes all or parts of these and possibly some additional counties: Bibb, Twiggs, Wilkinson, Jones, Baldwin, Washington, Jefferson, Glascock, Warren, McDuffie, Burke, and Richmond.

In the western part of this kaolin producing region most of the kaolin is Cretaceous. However, in the Wrens District near the eastern boundary of the State the kaolin is predominantly, if not entirely, in the Huber Formation.

The name Huber Formation is proposed after the placename and Post Office of Huber, Georgia, located on the Southern Railway in the northwestern part of Twiggs County. It was first studied in detail and its age determined from fossil collections in the area of the J. M. Huber Corporation mine, four to six miles (6.4 to 9.6 km) northeast of the Post Office and railroad crossing at Huber. The Huber Formation has become extensively exposed wherever mines have been opened from the vicinity of Dry Branch and Stone Creek Church on the northwest to Flat Creek on the southeast, a distance of about five miles. This can be considered as the type locality of the formation.

AGE AND STRATIGRAPHIC RELATIONS

Most geologists have followed Cooke (1943) in his usage of the name Tuscaloosa Formation for all of the Upper Cretaceous rocks in east central Georgia. Some, however, have used the term "undifferentiated Cretaceous" (Eargle, 1955). This term is more appropriate because in southwestern Georgia the Tuscaloosa is the oldest of six formations of Late Cretaceous age and beds correlating with some of the younger ones almost certainly are present east of the Ocmulgee River. Field work and stratigraphic correlations by Buie and Fountain (1967) also supported the need to avoid the term Tuscaloosa in referring to the kaolin-bearing strata of eastern central Georgia. Prior to 1964, several geologists, including Cooke (1943, p. 54) recognized or suspected the presence of Claiborne strata in eastern central Georgia but did not note that any of the Tertiary age deposits contained kaolin.

It appears that in mapping the sedimentary strata east of the Ocmulgee River, Cooke (1943) recognized and mapped the down-dip Claiborne strata as the McBean Formation but mistook the up-dip Claiborne with its beds of high-purity kaolin, for Cretaceous. If this interpretation is correct, the Huber Formation can perhaps be considered as the up-dip correlative of the McBean Formation.

In 1964, molds of a variety of small mollusks that appeared to be Tertiary were found by the writer in an argillaceous sand bed about one or two feet (less than one meter) thick beneath commercial kaolin being mined near Huber, Georgia. These fossil molds led to the first recognition that some of the commercial kaolin bearing beds in east central Georgia are younger than Cretaceous.

The age was determined by L. D. Toulmin who, by filling the molds with liquid latex, allowing the latex to become firm, and then brushing away the sand, was able to obtain identifiable replicas of the original fossils. In his report, Dr. Toulmin (1964) stated:

> It is difficult to determine species from fossil molds, and the bed containing the molds may be either the equivalent of the Gosport Sand or the lower part of the Moodys (Dellet Sand of Stenzel) from the genera recognized. It is faunally and lithologically different from any Moodys Branch beds I have seen in Georgia and therefore may be Gosport. Its stratigraphic relationship with the Moodys, if determinable, would be a better criterion of its age than the genera recognized. The fossils are *Nucula* probably *ovula*, *Nuculana*, *Pitar?*, *Crassatella*, *Linga pomilia alveata*, *Dentalium thalloides?*, *Mesalia vetusta?*, *Turritella.*
>
> Miss Gardner identified *Turritella carinata* and *Callucina alveata [Linga pomilia alveata]* in a bed on the Flint River 1½ miles above the bridge on the Americus-Vienna road. She says they are diagnostic of the Gosport Sand. (Southeastern Geological Society, 1944, Second field trip, Southwestern Georgia, p. 31.)

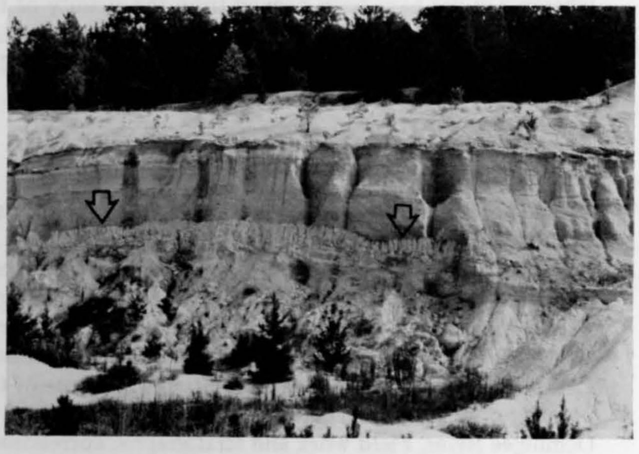

FIGURE 1 Contact of Huber Formation overlain unconformably by Clinchfield Sand. Vertical arrows mark Contact. Clay bed 2 to 3 feet thick is at top of Huber Fm.; fossil zone (2 ft. thick) is immediately beneath the clay bed. Georgia Kaolin Co. Mine 15, 5.6 miles (9.0 km) N45°E from railroad crossing at Huber.

The discontinuous bed containing the fossil molds, and the kaolin above it, are separated from the overlying typical marine fossiliferous Jackson strata by an unconformity. Hence, the identification of the kaolin as Claiborne rather than Jackson in age is supported by both paleontologic and stratigraphic evidence (Fig. 1 and 2).

Invertebrate marine fossils show that some of the kaolin is as young as Claiborne. In addition to the marine fossils, pollen of Tertiary age in the Huber Formation has been noted by several authors. Tschudy and Patterson (1975) identified Claiborne (middle Eocene) age pollen in samples from below a thick kaolin deposit and from beds not directly associated with kaolin at several localities. They also found Paleocene pollen in lignitic clays in a stratigraphic position between a thick white commercial kaolin and a stratigraphically higher kaolin. The Huber Formation, therefore, is known to contain beds of Paleocene and middle Eocene age. It probably also contains beds of Wilcox (lower? Eocene) age (Tschudy and Patterson, 1975). The range in ages of

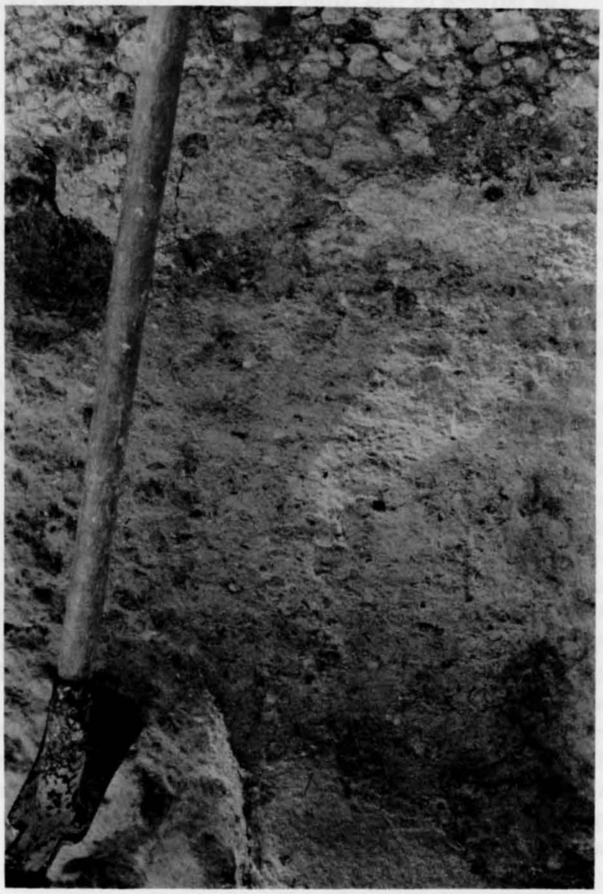

FIGURE 2 Close-up view of fossil zone in clayey sand, near top of Huber Formation. The small cavities are fossil molds; many of the cavity walls are limonite-stained. (Same location as Figure 1.)

beds in the Huber Formation represents approximately a 20-million year time interval (Fig. 2, p. 2-3, Huddlestun, Marsalis and Pickering, 1974). This interval includes all of post-Cretaceous and pre-Jackson time, and may appropriately be referred to as "pre-Jackson Tertiary".

Lithologies, stratigraphic relationships, and an unconformity at the top of the Huber Formation support a conclusion that the youngest beds are of middle Eocene rather than late Eocene age. The discontinuous kaolin and sand beds at the top of the Huber Formation are clearly similar to the lithologies of the lower beds and distinctly different from the fine-grained fossiliferous and marine characteristics of overlying beds. At different places in east central Georgia the younger Clinchfield Sand, Tivola Limestone, and Barnwell Formation directly overlie the Huber Formation. The definite marine characteristics of these beds contrast with those of the Huber Formation in which most of the beds have lithologies typical of nearshore deposition. In addition to the marked differences in lithologies an unconformity between the Huber Formation and overlying formations is indicated by an undulating contact (Fig. 1).

The Huber Formation is a wedge-shaped unit between the undifferentiated Cretaceous and the Jackson age strata. It wedges out updip where it is truncated by a middle Eocene or more recent erosion surface (Fig. 3). Irregularities in its thickness may be caused by uneven configuration of the unconformities, as well as by recent erosion. The formation is about 65 feet (19.8 meters) thick northeast of Huber, Georgia and increases to 94 feet (28.7 meters) thick 4.7 miles (7.5 km) downdip, in a mine in the Valley of Flat Creek north of Marion, Georgia. This is the maximum thickness measured and the thickness presumably continues to increase downdip, though not necessarily at the same rate.

The thickness of kaolin beds in the Huber Formation ranges typically from less than five feet (1.5 m) (usually not mineable), to a maximum of 20 to 30 feet (6 to 9 m). At one locality in western Wilkinson County a thickness of 60 feet (18 m) has been reported. A detailed stratigraphic section of the Huber Formation in the type locality is shown in Table 1.

LITHOLOGIC FEATURES

The lithology of the Huber Formation as a whole is very diverse, ranging from beds of high-purity and sandy kaolin to thick, cross-bedded members of coarse, pebbly sand, and even conglomerate composed of boulders of pisolitic kaolin (Fig. 4). However, the portion of the formation with which we are mostly concerned — the upper portion in which most of the Huber Formation kaolin that has been mined occurs — is more consistent. The accompanying description of a section (Table 1), measured about one and one-half miles southeast of Stone Creek Church, is typical for this part of the formation. Figure 5 is a photograph of the southeast wall of the mine where the section was measured.

A minor stratigraphic break is usually present at or slightly above the middle of the section. The lithology differs slightly above and below the break; fossil molds are commonly found above the break, but not below it; mica flakes and dark minerals are usually more abundant, and the sediments more laminated, below the break.

Distinctive features that aid in the recognition of the Huber Formation and in distinguishing it from the overlying Jackson formations include:

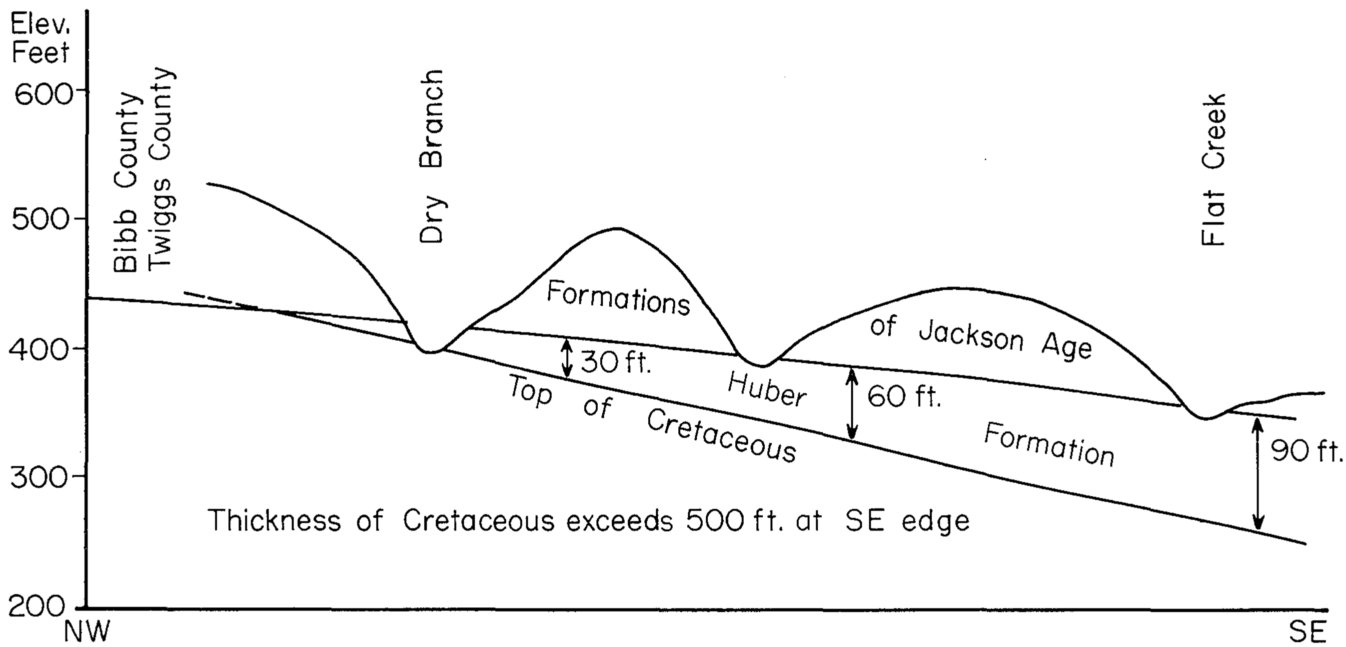

FIGURE 3 Schematic diagram of Cretaceous and Tertiary strata, Macon-Gordon area, Georgia.

TABLE 1 Detailed stratigraphic section of Huber Formation measured in J. M. Huber Corporation Mine No. 30, 5.8 miles (9.3 km) N55°E of railroad crossing at Huber Post Office, Twiggs County, Georgia.

	Feet	Meters
TERTIARY		
Jackson: Fossiliferous sands and clays.		
——— unconformity ———		
Claiborne:		
White kaolin with numerous kaolin-filled tubular forms about 0.1 inch (2.5 mm) diameter; also, cracks filled with sand and fossil fragments from the overlying unit.	5.0	1.5
Quartz sand with imprints and molds of small mollusks.	0.5	0.2
Medium to coarse grained white, kaolinitic, cross-bedded quartz sand with flakes of mica; dark minerals along cross-bedding. A few kaolin balls, 0.5 inch (1.3 cm) diameter, in basal 2 ft. (0.5 m) of coarse sand.	12.5	3.8
——— minor stratigraphic break ———		
Claiborne or older:		
Medium grained white micaceous quartz sand with laminae of dark minerals; a few kaolin balls to 0.5 inch (1.3 cm) diameter; some cross-bedded lenses of small pebbles of quartz; gradational with underlying bed.	4.8	1.5
Kaolinitic very coarse quartz sand and granules; cross-bedded; discoidal balls of kaolin having maximum dimensions of 1 inch (2.5 cm) by 2 inches (5 cm), in lower half.	3.6	1.1
Discontinuous layer of clay balls, 0.2 to 2 inches (0.5 to 5 cm), and quartz pebbles having smaller maximum diameter; dark minerals conspicuous, possibly 5% of total. Much of the quartz, especially that of the milky quartz pebbles, is friable.	1.7	0.5
Coarse grained kaolinitic quartz sand, with abundant dark mineral grains. Basal 1 ft. (30 cm) very coarse. Includes 1 to 2 inches (2.5 to 5 cm) of limonite and limonite-cemented sand at basal contact.	5.3	1.6
——— unconformity ———		
CRETACEOUS		
Kaolin; white, except for 2 inches (5.0 cm) limonite stain at top. About	20.0	6.1

1. The presence, at or near the top of the sequence, of a bed of kaolin, that is usually sandy, white to buff color and containing in most occurrences peculiar spaghetti-like forms of kaolin. These forms vary in abundance and may be absent or unrecognizable where much sand is present in the kaolin, or where the kaolin has been bauxitized and resilicated. These forms are usually about 0.1 inch (2.5 mm) in diameter, and may be as much as an inch (2.5 cm) or more in length. They appear to have been tubular cavities now filled with kaolin. Some have a branching form, suggestive of bryozoa; others appear to have resulted from the filling of small tubular borings. Some of the gray kaolin in deposits that have not been subjected to oxidizing conditions contain tubular-form openings that are partially filled with iron sulfide.

2. At many localities the Jackson strata, especially the calcareous portions, are abundantly and distinctively fossiliferous. The Claiborne strata contain far fewer, and different, fossils.

3. A thin zone of slightly kaolinitic quartz sand, rarely more than two feet (60 cm) thick, contains imprints and external molds of small pelecypods and gastropods; but no pectens, echinoids, oysters, or shark teeth, which are common in the Jackson strata.

4. The Huber Formation is devoid of calcareous materials, whereas the Jackson strata contain appreciable amounts of limestone and calcareous sandstone.

5. An abundance of mica flakes, now largely altered to hydromica, occur as laminae in the sand and sandy clay of the Huber Formation.

6. A much greater concentration of dark minerals than is usually seen in the Jackson strata is commonly present in the Huber Formation. (Where weathered, these minerals impart a deep red color to the sediments, thus causing some difficulty in distinguishing this unit from the weathered overlying Jackson strata.)

7. Present in the updip sections of the Huber Formation is a layer of quartz sand about four feet (1.2 m) thick replete with tubular borings *(Callianassa ?* burrows) that run in all directions, and which become conspicuous where slight erosion removes the less cemented interstitial sand.

8. Beds and lenses of bauxite, and in some instances pisolitic kaolin, formed from re-kaolinization of bauxite, are common at the tops of kaolin beds in the Claiborne and Wilcox (?) sections. Bauxite is not found in the Jackson strata.

9. The Jackson strata are more uniform — being of marine deposition — have a darker green or brownish green color, and usually contain smectite rather than kaolinite as the dominant clay mineral.

Features that distinguish the Huber Formation from the undifferentiated Cretaceous beds include: 1) a greater concentration of heavy dark minerals in the Huber Formation; and 2) a thin layer of limonite less than one inch (2.5 cm) thick at the top of the Cretaceous clay. This is generally present where the Cretaceous clay is overlain by sand of the Huber Formation. Other distinguishing features are mainly related to differences in the two kinds of kaolin, as summarized in Table 2.

FIGURE 4 Road cut exposure of updip Huber Formation, showing boulders of pisolitic clay or bauxite; 1.5 miles north of Bethlehem Church in southeast corner of Jones County. (Pick handle is 38 inches long.)

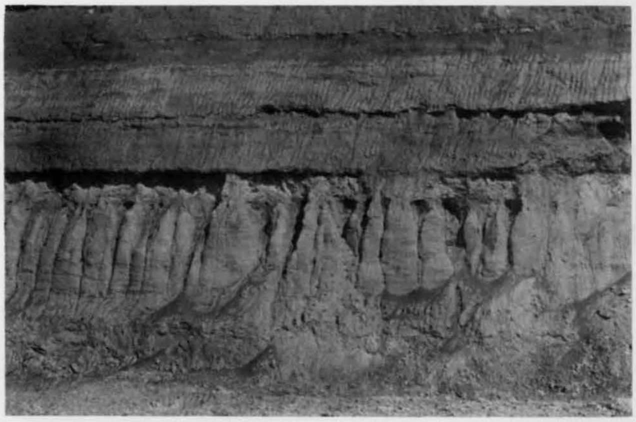

FIGURE 5 Photograph of measured section (Table 1) on southeast wall of Mine 30, J. M. Huber Corp. The Huber Formation extends from the top of the Cretaceous kaolin at the bottom of the cut up to the overhanging bed at the middle of the photograph. Jackson strata occupy the upper half of the view.

The colors of Huber Formation clay mentioned in Table 2 refer to the Claiborne clay. At several localities, notably near Gordon, considerable thicknesses of magenta-colored clay occur below the typical Claiborne part of the section and above what is believed to be Cretaceous. These are thought possibly to be Wilcox strata. Elsewhere in the Huber Formation tan or light brown clay is found that is too strongly colored to be commercial at present.

The striking contrast in the fracture of what is referred to as "hard clay" (Tertiary) and "soft clay" (Cretaceous) is seen in Figure 6. Usage of the terms hard and soft has

TABLE 2 Characteristics of kaolins of the Cretaceous beds undifferentiated and Huber Formation.

General Characteristics	Kaolin in the Cretaceous beds undifferentiated	Kaolin in the Huber Formation
Color	white to cream or buff; natural brightness high	mainly white to cream, buff, or gray; commonly has a slight greenish tinge that changes to a faint pinkish tint upon exposure; some deposits are very light gray; natural brightness moderate to low
Fracture or parting	smooth; when dry, breaks easily into friable blocks having sharp angles and smooth flat or subconchoidal faces; commonly called "soft clay"	mostly breaks into irregular chunks with rough to hackly fracture; commonly called "hard clay"
Tubular structures resembling borings or bryozoa	absent	common
Pisolitic texture	rare	common in upper parts of deposits
Particle size	generally 65% or less < 2 micrometers; vermicular crystals and "books" common	generally more than 90% < 2 micrometers; vermicular crystals scarce
Crystal perfection	generally good	generally poor
TiO_2 content	about 1%	about 2%

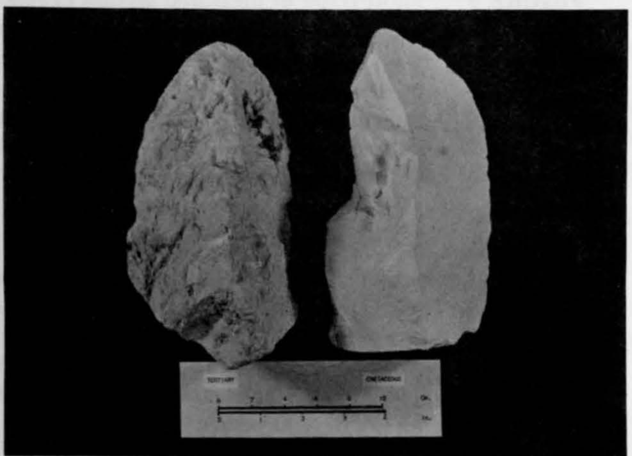

FIGURE 6 Contrast of fracture form in Tertiary and Cretaceous kaolin.

not been entirely consistent. As most often used, "hard clay" refers to a clay having high dry strength. Probably the most persistant feature of "hard clay" is its extremely fine particle size. Usually 90 percent or more of the clay component in a sample is composed of particles smaller than two micrometers in diameter. Not all of the Tertiary clay has characteristics typical of "hard clay". Some clay from intermediate depths in the Huber Formation has properties and particle size distribution somewhat like the Cretaceous clay.

Degree of crystal perfection varies greatly. On average it appears to be poorer in the Tertiary clay, but some specimens that show poorly outlined crystals at moderate magnification under the scanning electron microscope show much sharper crystal outlines at higher magnification. Figures 7 and 8 are scanning electron microscope (SEM) photographs of the same specimen at 12,000 and 60,000 diameters magnification. At the lower magnification crystal outlines are hardly visible, but at the higher magnification they are clearly seen.

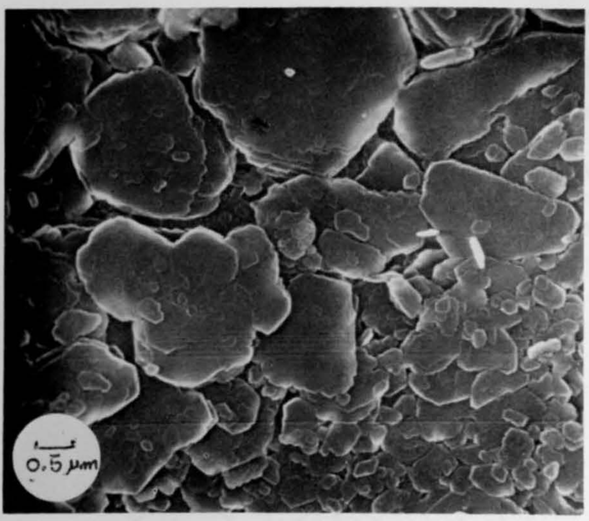

FIGURE 7 Scanning electron microscope (SEM) photograph of a typical specimen of Tertiary (Claiborne) kaolin. Note scarcity of recognizable crystal outlines, and absence of kaolin "books". 12,000X

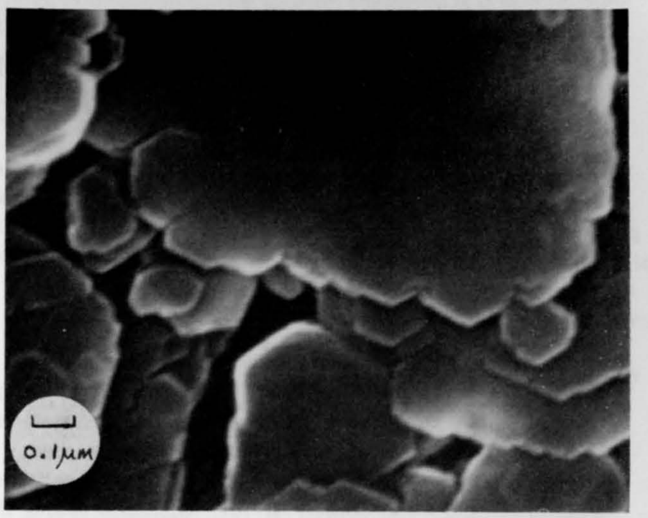

FIGURE 8 SEM photograph of same clay specimen as in Figure 6. Serrate edges of flakes are minute crystal outlines. 60,000X

The TiO_2 content of the Huber Formation kaolin is primarily in the form of anatase, occurring as particles smaller than one micrometer in diameter.

ACKNOWLEDGMENTS

This definition of the Huber Formation was prepared at the request of the Director of the Georgia Geologic and Water Resources Division, Department of Natural Resources. It is a part of a program to revise and update the stratigraphy of the Coastal Plain sedimentary rocks.

The management personnel of J. M. Huber Corporation, by granting a large degree of freedom in the geological work done while in their employ, made possible the compilation of information here reported. The unreserved cooperation of other mining companies in allowing access to their properties has been very helpful, and is appreciated. For identification of the fossils and for fixing their position precisely within the Tertiary sequence, the contribution by L. D. Toulmin, Emeritus Professor of Geology, Florida State University, is gratefully acknowledged.

Thanks are expressed to Sam H. Patterson, U. S. Geological Survey, for use of the photograph of the road cut exposure.

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ANOMALOUS DISTRIBUTION OF SINKS
IN THE UPPER LITTLE RIVER WATERSHED:
TIFT, TURNER, AND WORTH COUNTIES, GEORGIA

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ABSTRACT

The upper Little River watershed, located in central south Georgia, is considered typical of southeastern Atlantic Coastal Plain watersheds and is an area under observation by the U. S. Department of Agriculture. In the watershed, Miocene clastics overlie Cenozoic limestones that constitute a major ground-water aquifer. In four small areas along the east side of the Little River, Pleistocene eolian sands have accumulated on top of the Miocene clastics. One of these areas, comprising only 2.1 percent of the total area, includes 26 percent of the 105 sinks located in the watershed. The concentration of sinks in the other three areas covered by Pleistocene sand is not abnormally high with respect to the watershed as a whole, and the high concentration of sinks in one area of Pleistocene sand accumulation must involve more than high infiltration rates.

INTRODUCTION

The upper Little River watershed, in Tift, Turner, and Worth Counties, Georgia (Fig. 1), has been considered a typical small southeastern watershed by the Department of Agriculture and its hydrology is being intensively studied by the Department of Agriculture. In 1967, the author, under contract to the Department of Agriculture, made a reconnaissance of the upper watershed and surrounding area and reported general features of the geology (Carver, 1967a). Part of the data from that report was published by L. E. Asmussen in 1971. In the interest of a complete statement of the problem, additional data are presented in this short contribution.

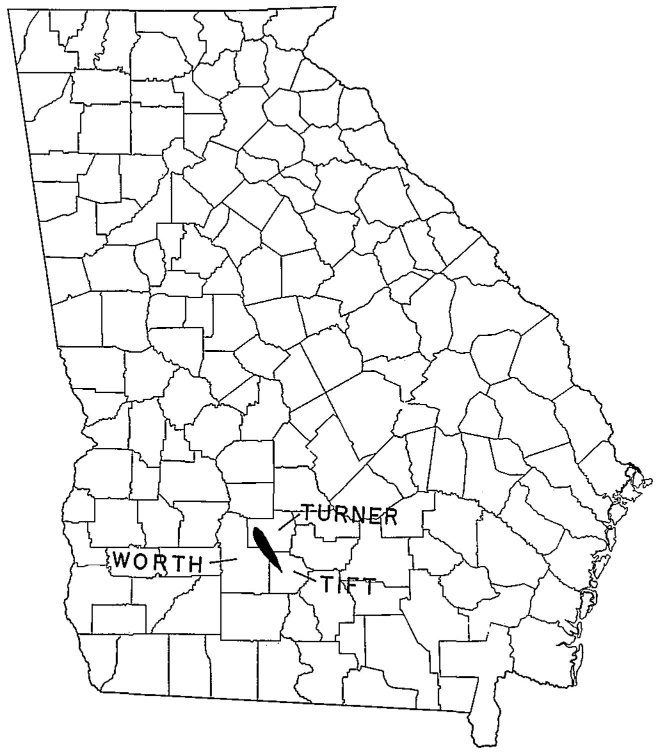

FIGURE 1 Index map showing the location of the Upper Little River watershed within the Coastal Plain of Georgia and identifying Tift, Turner, and Worth Counties.

GENERAL GEOLOGY

The upper Little River watershed lies entirely within the Tifton Upland physiographic district (Clark and Zisa, 1976). The surface formations in this area are currently mapped as Neogene Undifferentiated (Georgia Geological Survey, 1976) and consist primarily of interbedded sands and clays of the Hawthorn Formation and hard, sandy claystones of the Altamaha Grit. The Altamaha Grit was originally described by Dall and Harris (1892) and was informally described by Olson (1967) as the Ashburn Member of the Hawthorn Formation. The surficial units are underlain by a succession of Miocene, Oligocene, and Eocene limestones and coarse clastics that, together, make up the Floridan aquifer in this general area.

Piezometric surface maps of the area (Stringfield, 1966, p. 120 and 206; Carver, 1968) indicate that Tift and Turner Counties lie over a major pressure plateau in the aquifer that is produced by either a fault or a permeability barrier along a line running northeast and located approximately along the southeast border of Tift County. Between 1880 and 1942, deep ground-water pressure in this area declined approximately 50 feet in response to heavy pumping along the coast, but the general shape of the pressure plateau did not change, indicating that a physical restriction of ground-water movement is present (Carver, 1968). The free ground-water table in the area lies at depths of 0 to 19 feet, based on measurements of water levels in shallow wells and on the assumption that the Little River is an effluent stream. This water table probably is related to local surface drainage, rather than the Floridan aquifer, as noted below.

Carver (1967a) reported eolian sands overlying the sediments in four areas along the east side of the Little River

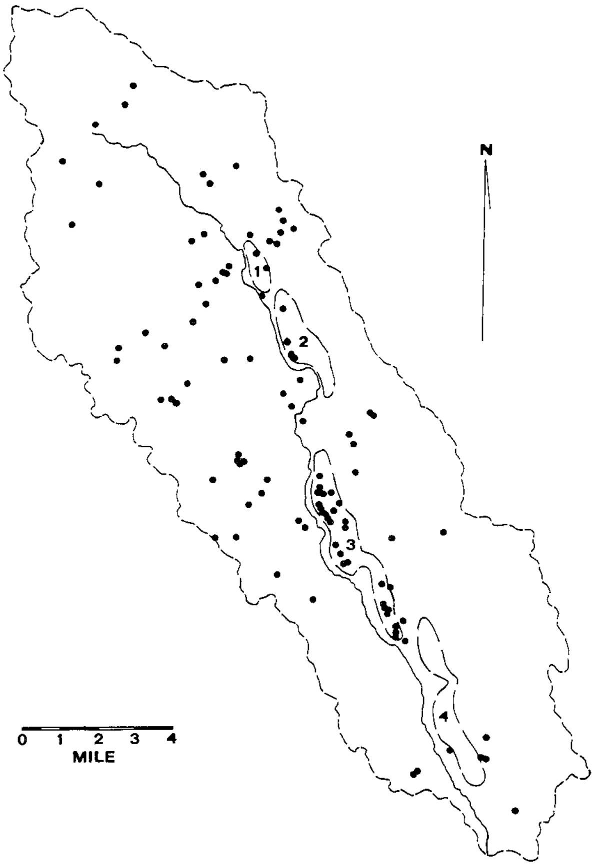

FIGURE 2 Outline map of the Upper Little River watershed, showing the locations of sinks (black dots), the main channel of Little River, and areas of Pleistocene eolian sand accumulation (numbered areas defined by lines with long dashes).

(Fig. 2). Eolian sands along the east, or northeast, sides of rivers are common in Georgia, and on the basis of their immature heavy mineral suite (Carver, 1967b) and the presence of artifacts (Kelley, 1967), they appear to be late Pleistocene in age. Asmussen (1971) ran three refraction seismic traverses across Area 3 of Figure 1 and reported sand thicknesses of 5 to 15 feet. Asmussen also noted that a perched water table occurs within the sand, held up by the impermeable Miocene sediments below.

CHARACTERISTICS OF THE SINKS

Examination of aerial photographs of the upper Little River watershed revealed 105 features that could be identified as sinks. The features are circular to irregular in shape, none of them have the smooth oval outline and raised sand rims that would suggest Carolina bays. One third of the sinks have overflow outlets and probably are not freely connected to the underlying limestone aquifer, and it is possible that this is true of most of the sinks in the watershed. However, Lime Creek Sink, just north of the watershed, takes the entire flow of Lime Creek and must have an open connection to the aquifer; therefore it is also possible that some of the sinks in the watershed are open to the aquifer. The size distribution of sinks is given in Table 1.

DISTRIBUTION OF THE SINKS

Of the 105 sinks in the upper Little River watershed, 27 sinks (26 percent) occur in one area of eolian sand accumulation, Area 3 of Figure 2, that constitutes only 2.1 percent of the total watershed area. The density of sinks in Area 3 is 8.8 per square mile, as opposed to 0.5 sinks per square mile for the watershed exclusive of Area 3. One might be tempted to relate the high density of sinks in Area 3 to higher infiltration rates and the perched water table within the surface sand body. However, according to Herrick (1961) lower Miocene limestones in this area are overlain by 100 to 300 feet of clay, sandy clay, or interbedded sand and clay. Based on observation of outcrops in the area of the upper Little River watershed, the interbedded sands are highly lenticular. The vertical and lateral permeability of the Miocene clastics should be very low and it is difficult to conceive that the presence of 5 to 15 feet of sand on the surface would have any major effect on the development of sinks that extend through the 100 to 300 feet of Miocene clastics. In addition, the density of sinks in the other three areas (Table 2) is not significantly greater than for the watershed as a whole and the presence of surface sands probably is not the cause of the concentration of sinks in Area 3.

The pronounced alignment of six sinks along the west edge of Area 3 suggests that some combination of joints and permeability of the surficial material may be involved, but it appears that the anomalous distribution of sinks in Area 3 cannot be adequately explained on the basis of available data.

TABLE 1 Size distribution of sinks in the upper Little River watershed.

Diameter (feet)	Number of sinks
200	32
200-400	42
400-800	24
800-1200	6
1200-1600	1
Total	105

TABLE 2 Distribution of sinks in the upper Little River watershed, by area shown in Figure 2.

	Surface Area, Square Miles	Percent of Total Area	Number of Sinks	Percent of Sinks	Sinks per Sq. Mi.
Area 1	0.6	0.4	2	2	3.3
Area 2	1.8	1.2	4	4	0.5
Area 3	3.1	2.1	27	26	8.8
Area 4	2.5	1.7	0	0	0
Total Watershed	147	100	105	100	0.71

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STRATIGRAPHIC SIGNIFICANCE OF HEAVY MINERALS IN ATLANTIC COASTAL PLAIN SEDIMENTS OF GEORGIA

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ABSTRACT

Composition of the heavy mineral suite of Atlantic Coastal Plain sediments is largely determined by the degree of intrastratal solution of the unstable heavy mineral suite: hornblende, epidote, and garnet. Piedmont rivers carry sediment rich in hornblende and epidote to the coast, but hornblende and epidote in this sediment are diluted by mixing of river sediment with more mature Coastal Plain sediment, and in addition, are depleted by post-depositional intrastratal solution. For late Pleistocene to Holocene sediments, the composition of the heavy mineral suite indicates the degree of mixing of sediment from Piedmont sources and Coastal Plain sources. In some cases, differences between heavy mineral suites of two bodies of sand are almost entirely due to the degree of intrastratal solution and are proportional to the age of the units.

INTRODUCTION

Heavy mineral studies have been useful in interpreting the geology of the southeast Atlantic Coastal Plain, primarily in determining patterns of sediment dispersal and age of sediments. Ilmenite, muscovite, biotite, garnet, staurolite, kyanite, sillimanite, brown tourmaline, zircon, and rutile are common to abundant in nearly all heavy mineral suites of the Coastal Plain. This heavy mineral assemblage indicates that the source area includes high and low grade metamorphic and acidic and basic igneous rocks. The source area is undoubtedly the Appalachian Highlands Major Division of Clark and Zisa (1976).

Pleistocene and Holocene sands of the Coastal Plain contain hornblende and epidote in addition to the minerals mentioned above. Hornblende and epidote are absent in older Coastal Plain sands because they are destroyed by intrastratal solution. Carver (1971) and Carver and Kaplan (1976) were able to distinguish Atlantic Shelf sands, derived from the Piedmont and carried to the coast by major rivers, from Shelf sands derived from older Coastal Plain sediments on the basis of the hornblende content. The presence of abundant hornblende also distinguishes Pleistocene eolian sands along major rivers from older sediments (Carver, 1967a).

The actual percentage composition of clastic heavy mineral suites in Coastal Plain sediments depends on the interaction of three factors: 1) composition of the heavy mineral suite derived from the source area as modified during transport, 2) grain size of the sediment sample, and 3) intrastratal solution of the less stable elements of the heavy mineral suite. The first factor depends on composition of the source area and the weathering history of the sediment as it is transported to the site of deposition. Unfortunately, in the Atlantic Coastal Plain, the effect of intrastratal solution tends to completely overshadow the effect of original variation in source area and weathering in transport, severely limiting the opportunity to study subregional differences in source area and past differences in rate of transport or weathering.

Carver (1971) and Gadow (1972) have investigated the size distribution of heavy minerals in Holocene sediments from the coastal area of Georgia (see also Potter and Pettijohn, 1963, p. 192-3). Size selection of the available heavy minerals has an important effect on composition of the heavy-mineral suite in any given sample. Zircon, for instance, is much more abundant in fine-grained sediments than coarse-grained sediments, and the opposite is true of tourmaline and staurolite. In comparative studies, it is therefore necessary to study a single size fraction of the samples, preferably a relatively narrow size range. In recent years, most workers have selected the 2 Phi to 3 Phi size fraction, and data reported here relate to that fraction unless specifically noted.

The third factor that determines composition of the heavy mineral suite is intrastratal solution. The continuous circulation of large quantities of relatively warm ground water through near-surface, coarse clastic sediments of the Atlantic Coastal Plain produces, over long periods of time, significant dissolution of the less stable heavy minerals. The common minerals present and considered unstable by Scott (1976) are hornblende, epidote, and garnet, in order of increasing stability. Scott considers kyanite and staurolite as metastable minerals. The chemical mineralogy of garnets is complex and their inclusion in the unstable group might be questioned. Overall, intrastratal solution appears to be the dominant factor in differences in heavy mineral composition of southeast Atlantic Coastal Plain sediments.

HEAVY MINERALS OF HOLOCENE SEDIMENTS

Giles and Pilkey (1965), in an extensive study of heavy minerals of Holocene sediments of the Atlantic Coastal Plain, determined that the most significant differences in heavy mineral suites were between river sands in rivers with headwaters in the Piedmont and rivers with headwaters restricted to the Coastal Plain. The Piedmont river suites are dominated by hornblende, while the more stable epidote is the most common mineral in Coastal Plain rivers, as illustrated by analyses 1 and 2 of Table 1. The differences in Piedmont and Coastal Plain river heavy mineral suites are due to the fact that most of the sediment in Coastal Plain

TABLE 1. Analyses of non-opaque, non-micaceous heavy mineral fraction of Holocene and Pleistocene Atlantic Coastal Plain sands.

Analysis No.	1	2	3	4	5
Hornblende	46	7	26	12	5
Epidote	28	39	29	26	21
Kyanite	4	3	11	7	16
Sillimanite	7	10	19	24	28
Staurolite	4	13	6	10	13
Garnet	2	2	2	2	2
Zircon	5	13	2	5	6
Tourmaline	2	8	3	8	3
Rutile	2	3	0	2	4
Others	+	1	3	4	2
Total	100	99	101	100	100

Analysis 1: Average of 2.3 Phi to 3.0 Phi fraction of 17 samples from the lower Altamaha River (a Piedmont river) from Gadow (1972, Tables 2 and 5).

Analysis 2: Average of 2.3 Phi to 3.0 Phi fraction of 9 samples from the lower Ogeechee River (a Coastal Plain river) from Gadow (1972, Tables 2 and 4).

Analysis 3: Silver Bluff, Sample RB-3B, from Scott (1976, p. 115).

Analysis 4: Princess Anne, Sample SB-20C, from Scott (1976, p. 116).

Analysis 5: Pamlico, Sample SM-1, from Scott (1976, p. 116).

Analyses from Scott represent the 2 Phi to 3 Phi fraction. The analyses were selected to approximate an average from the sample locality and were rounded from Soctt's nearest 0.5 percent data.

rivers is derived from older Pleistocene, Miocene, Paleogene or Cretaceous sediments that have been through one cycle of weathering and transport and a long period of intrastratal solution.

The difference in mineralogical maturity of Piedmont and Coastal Plain sources of sand makes it possible to determine the source of Pleistocene sediments on the Atlantic Shelf. During Pleistocene and Holocene transgressions and regressions extensive erosion of older Coastal Plain sediments occurred and sediments deposited at the coast were, or are, predominantly from this source. Sediments deposited along the coast during Pleistocene or Holocene periods of transgression or regression therefore tend to have very low hornblende contents. Conversely, during periods of sea-level stability the major sources of sediment at the coast are the major rivers of the area which drain the Piedmont and Blue Ridge. These rivers carry sand that is rich in hornblende and sediments deposited along the coast during periods of sea-level stability contain high percentages of hornblende.

Carver and Kaplan (1976) were able to identify Atlantic Shelf sediments that had been deposited near the mouths of major rivers during periods of sea-level stability by mapping the hornblende content of Shelf sediments. The high-hornblende sediments extend south of the positions of river mouths projected for lower stands of the sea because much of the sediment was carried from the river mouths, or deltas, south, along the coast, by longshore drift.

HEAVY MINERALS OF PLEISTOCENE SEDIMENTS

Scott (1976) studied the heavy-mineral content of sands from Holocene and Pleistocene sediments of known age to determine if significant intrastratal solution took place within the Pleistocene. He found strong correlations between age of Holocene and Pleistocene sediments from Georgia and composition of the heavy mineral suites. Correlation coefficients exceeding 0.8 are presented in Table 2. The units sampled and assumed ages (Henry and Hoyt, 1968; Hails and

TABLE 2. Correlation of heavy mineralogy with age of sediment, from Scott (1976, p. 120-121). The correlations are between percentage of non-micaceous, non-opaque heavy minerals and age. The data are based on average heavy mineral percentages of samples from Holocene beach and dunes (assumed age 0 years B.P.), the Silver Bluff Terrace (25,000 years B.P.), the Princess Anne Terrace (80,000 years B.P.), and Pamlico Terrace (120,000 years B.P.).

Parameter	Correlation Coefficient
Percent Hornblende	-0.994
Percent Epidote	-0.901
Percent Kyanite	0.961
Percent Sillimanite	0.984
Percent Staurolite	0.912
Percent Garnet	0.957
Percent Tourmaline	0.924
Percent Hornblende + Epidote	-0.972
Percent Kyanite + Sillimanite + Staurolite	0.971
Percent ZTR	0.887

Hoyt, 1969) were Holocene beach and dune sands, 0 years B.P.; Silver Bluff, 25,000 years B.P.; Princess Anne, 80,000 years B.P.; and Pamlico, 120,000 years B.P. In general, the significant changes with age appear to involve only reduction in the percentages of hornblende and epidote, and a consequent increase in the percentages of all other heavy minerals. It is noted that the percentage of garnet increases with age of sediment at about the same rate that tourmaline does, and garnet probably should be considered a stable mineral in this context. Typical analyses for Silver Bluff, Princess Anne, and Pamlico terrace sands are given in Table 1.

Late Pleistocene age eolian sands occur along the east or northeast sides of many, perhaps most, of the larger and smaller streams of the Coastal Plain. Abundant Indian artifacts are present in some of the dune fields (Kelley, 1967) and the eolian sands are, at least in part, of late Pleistocene to Holocene age. Where eolian sands lie on much older sands flanking Piedmont rivers, the eolian sands can be distinguished by their immature heavy mineral suite, since the presence of 5 percent or more hornblende is sufficient to distinguish eolian sands from Paleogene or Cretaceous sediments (Carver, 1967a). Rivers with drainage basins confined to the Coastal Plain, however, derive all of their sediment from Cretaceous to Miocene clastic sediments and eolian sands flanking these rivers contain heavy-mineral suites similar to those of the source sediments (Carver, 1967b).

HEAVY MINERALS OF MIOCENE TO CRETACEOUS SEDIMENTS

Heavy mineral analyses of five samples ranging in age from late Eocene to perhaps late Cretaceous are given in Table 3. The first four analyses are similar in that they are dominated by the kyanite, sillimanite, staurolite and the zircon, tourmaline, and rutile (ZTR) groups, as are Miocene sediments from the Upper Little River Watershed in Tift, Turner, and Worth Counties (Carver, 1967b). There is considerable variation in the proportion of minerals within these groups, which perhaps holds promise for future studies of provenance. Hornblende is virtually absent from these samples and epidote is present in small quantities, if at all. Garnet is present as one percent or less of the non-opaque, non-micaceous fraction, as compared to 2 percent in Holocene sediments, suggesting that garnet is unstable over long periods of intrastratal solution.

Analysis 5 of Table 3 is from a sand lens entirely enclosed within a kaolin ore body. The lens is crossbedded, about one meter thick at maximum and extends approximately 15 meters along the face of the mine cut. A 5 cm thick lignite seam occurrs within the sand lens. Preservation of the original contacts and bedding structures indicates that there had been little, if any, volume change in the surrounding sediment, and the sand lens probably was unusually well protected from intrastratal solution. There are, therefore, significant percentages of surviving hornblende, epidote, and garnet present in the heavy mineral suite. If this represents the original sediment composition, the sediment was considerably more mature than Holocene Piedmont-derived sediments now reaching the coast, but this is impossible to determine on the basis of a single occurrence of apparently protected sediment.

SUMMARY

Heavy mineral content of Atlantic Coastal Plain sands is determined by provenance; weathering and sorting during transport; grain size of the sediment, or sediment fraction, examined; and the degree of intrastratal solution of unstable heavy minerals. Hornblende and epidote are abundant in sediments from rivers with headwaters in the Piedmont and virtually absent in Paleogene and Cretaceous Coastal Plain sands that have been subjected to long periods of intrastratal solution. Relative to Holocene coastal sediments, sands of the three youngest coastal Pleistocene terraces are progressively depleted in hornblende and epidote. It is therefore possible, under certain circumstances, to determine relative age of sand deposits on the basis of heavy mineral content, and it is also possible to define the relative contribution of Piedmont and Coastal Plain sources to Holocene coastal sediments. Studies of weathering of sediment in transport, presumably related to intensity of weathering and rate of transport, will probably depend on locating sediments that have been protected from the effect of intrastratal solution.

TABLE 3. Analyses of non-opaque, non-micaceous heavy mineral fraction of Eocene and Cretaceous Atlantic Coastal Plain sands.

Analysis No.	1	2	3	4	5
Hornblende	0	1	1	0	3
Epidote	8	4	3	0	12
Kyanite	7	20	29	25	5
Sillimanite	8	10	4	3	26
Staurolite	15	15	9	22	11
Garnet	1	1	0	0	4
Zircon	30	18	11	2	14
Tourmaline	16	10	17	21	12
Rutile	12	16	21	27	12
Others	3	5	5	0	1
Total percent	100	100	100	100	100

Analysis 1: Eocene, Irwinton Sand, Twiggs Co., Georgia, Sample No. WRR 30.

Analysis 2: Eocene, Claiborne undifferentiated, Macon Co., Georgia, Sample No. MAT 12.

Analysis 3: Cretaceous, Providence Sand, Schley Co., Georgia, Sample No. MAT 11.

Analysis 4: Cretaceous, Tuscaloosa Formation, Taylor Co., Georgia, Sample No. MAT 5.

Analysis 5: Late Cretaceous, Early Tertiary undifferentiated, from a sand lens within commercial kaolin, Twiggs Co., Georgia, Sample No. MAT 3.

Analyses are based on counts of 100 non-opaque, non-micaceous grains from the 2 Phi to 3 Phi size fraction of the sample.

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SOIL GEOCHEMISTRY OF THE FRANKLIN-CREIGHTON GOLD MINE, CHEROKEE COUNTY, GEORGIA

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ABSTRACT

The Franklin-Creighton mine, located in east-central Cherokee County, Georgia, was the most important gold mine in Georgia during the late 19th and early 20th centuries. Estimated production was between 622,000 and 833,000 grams from workings extending to a depth of 285 meters down pitch and 1500 meters along strike.

Ore bodies were relatively small, pipe-like shoots apparently occupying dilatant zones within a major shear. Individual ore bodies were characterized by unusual uniformity in grade and persistence with depth. Ore is mineralogically simple, consisting of quartz, pyrite and pyrrhotite with minor chalcopyrite, calcite, ankerite, and tourmaline. The grade of individual ore bodies varies from 9.0 to 30.2 grams of gold per ton.

The property is ideally suited for soil geochemical evaluation in deeply weathered terrane due to the uniformity of grade and well established size, outcrop location, and attitude of ore bodies. The potential for sample contamination is unusually low because of favorable mill and dump locations, and general inactivity on the property since abandonment.

Soil geochemical data for gold, copper, and arsenic were obtained from 198 samples collected on a grid with 30.5-meter station spacing. Maximum values reported are 2.7 ppm gold, 155 ppm copper, and 40 ppm arsenic. Gold data reflects an apparent bimodal population with thresholds of 0.4 and 0.8 ppm. Copper threshold is 80 ppm. Arsenic data do not exhibit an anomalous population.

Surface distribution of gold and copper in soils of the southeastern United States is complex and relationships between anomalies and outcropping mineralization are not straightforward. Although initial drill sites would be difficult to determine based solely on the soil geochemical data, combined gold and copper contours generally reflect the location of known ore body outcrops and the trend of the major shear structure. Significant anomaly offsets from ore body outcrops are attributed to dispersion and possible migration of gold and copper during soil profile evolution, topography, and to residual gold accumulation from previously weathered, up-pitch portions of the deposit. Anomalous gold content of soil in the immediate vicinity of ore body outcrops reflects a dispersion or depletion ratio during soil development of approximately 25:1 when compared with the grade of unweathered ore.

INTRODUCTION

The present study emphasizes both the importance and complexity of soil geochemistry as a rapid and inexpensive exploration technique for the deeply weathered terrane of the southeastern United States. While the value of soil geochemistry is already firmly established by those companies routinely engaged in base and precious metal exploration in this region, the results, techniques, and interpretive parameters of these programs are rarely published.

The Franklin-Creighton mine was the most extensively developed underground gold mine in Georgia and perhaps in the southeastern United States. The mine is inadequately described in the literature since the greatest period of production was subsequent to the publication of the comprehensive bulletin, "Gold Deposits of Georgia" (Yates and others, 1896). Flooding and abandonment precede by many years the work conducted for "Gold Deposits of the Southern Piedmont" (Pardee and Park, 1948).

The mine was developed along an approximately 1500 meter-long, northeast-trending mineralized zone within land lots 397, 399, 465, 466, 467, 472, and 473, 3rd district, in extreme eastern Cherokee County, Georgia (Figure 1). The property is approximately 65 kilometers north of Atlanta and 13 kilometers southeast of Ball Ground. The formerly important gold mining district at Dahlonega is 40 kilometers to the northeast.

The property is ideally suited for the determination of soil geochemical exploration parameters in the deeply weathered terrane of the southeastern United States for the following reasons:

A. The uniformity of grade within individual ore bodies permits correlation of geochemical data with ore grade.

B. The well established size, outcrop location, and attitude of ore bodies allows accurate interpretation of geochemical dispersion and signature.

C. The potential for possible sample contamination by dump material and mill tailings is low due to the utilization of mining methods resulting in very little waste, and the location of the chlorination and cyanide plants approximately 450 meters across strike and down slope from the mineralized zone.

D. After abandonment the mine area was allowed to revert to forest with the exception of minor intermittent truck farming and ranching in the southwestern portion of the grid area.

HISTORY AND DEVELOPMENT

The Franklin-Creighton gold mine was worked almost continuously for about 70 years commencing in 1840 (Becker, 1895; Blake, 1895; Nitze and Wilkins, 1897). The mine is acknowledged as the most important in Georgia during the early part of the 20th century and was one of the steady producers of gold in the South, having yielded a

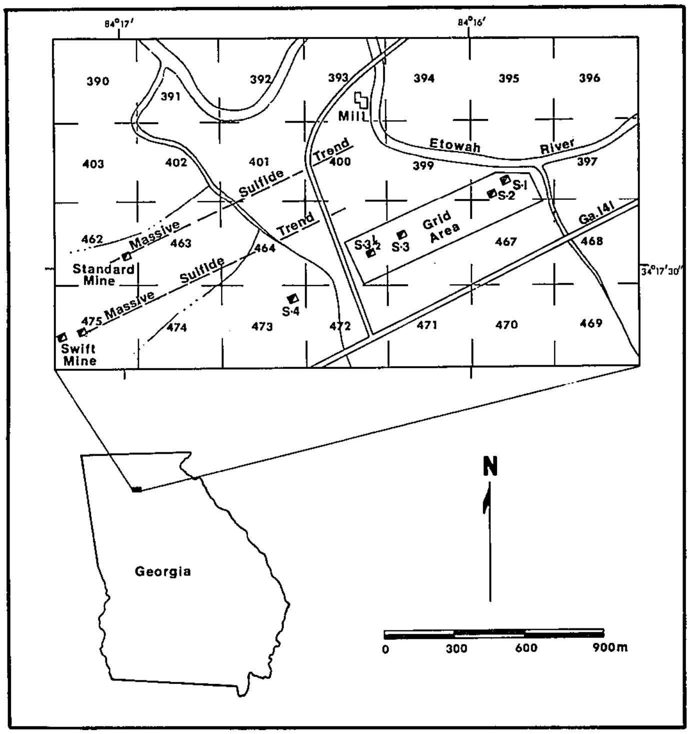

FIGURE 1. **Land lot map of the Franklin-Creighton Gold Mine and grid area. The five main shafts are indicated S-1 through S-4. Massive sulfide occurrences are at the Standard and Swift mines.**

conservatively estimated minimum of between 622,000 and 933,000 grams (Pardee and Park, 1948). Planimetric measurement of old stope maps indicates that this production was from approximately 180,000 tons of ore. Production figures for 1907, the last year for which complete data are available, indicate a daily production of 50 tons of ore containing 21.2 grams of gold per ton.

Although several closely spaced, parallel gold-bearing zones crop out on the property, most former production was from a persistent feature referred to in old reports and mine maps as the Franklin "vein". A total of 1160 meters of shafts and 2500 meters of drifts and crosscuts have been opened on this mineralized feature.

Massive pyrite for the production of sulfur was mined during World War I from several mineralized zones northwest of and parallel to the trend of the gold ore bodies (Fig. 1). Production was from two mines operated by the Standard Pyrites Company, and Swift and Company. Approximately 26,000 tons of pyrite concentrate were shipped from these two mines through October 1, 1917 (Shearer and Hull, 1918).

The most extensive underground development of the Franklin vein or zone was from two shafts approximately 60 meters southeast of the Etowah River (shafts No. 1 and No. 2). An idealized projection of these workings compiled from old descriptions and maps is shown in Figure 2. By 1895 four ore bodies had been developed from shafts No. 1 and No. 2 to depths varying from 60 to 120 meters along the pitch of the deposits (McCallie, 1907). A northeast-trending exploration drift was driven on the 106-meter level to a point 120 meters beyond the northeastern-most ore body. Another prospecting drift was extended to the northwest from the 60-meter level to the parallel McDonald "vein". By 1907, shaft No. 1 had been deepened along the pitch of the ore bodies to well below the 285-meter level.

Approximately 425 meters southwest of shaft No. 2, a shaft (No. 3, Fig. 2) was sunk to exploit three ore zones. All ore in the two largest bodies was stoped out above the 100 meter level. A fourth shaft (No. 3½, Fig. 2), located approximately 210 meters southwest of shaft No. 3, attained a total depth of about 150 meters measured along the pitch of the ore body. Limited stoping was conducted on the 27— and 60—meter levels. An additional shaft, located approximately 490 meters southwest of shaft No. 3½, was put down on the outcrop of an ore zone to a depth of 256 meters along the pitch. All ore above the 206—meter level was stoped out. Two additional ore zones that were apparently not exploited are indicated in the immediate vicinity of this shaft by McCallie (1907).

GEOLOGY

Although the general geology of the mine area has been shown on several regional maps, the detailed geology has not been presented in the literature or in unpublished reports. The geology of the area was originally mapped at a scale of 1:62,500 by Bayley (1928) in conjunction with work in the Tate, Georgia marble district. Detailed geology of adjacent Forsyth County, the western boundary of which is approximately 1.6 kilometers east of the mine, has been recently mapped by Murray (1973).

The mine area is within a large, northeast-trending sequence of thinly interbedded hornblende schist and gneiss, and amphibolite. This sequence is bounded by metagraywacke and various mica schists, gneisses, and thin interbedded quartzites. An extensive unit of biotite granite gneiss, formerly known as the Hightower granite, crops out approximately 0.8 km east of the mine. A major northeast-trending fault, the Dahlonega shear, has been mapped through the immediate mine area. The general strike of foliation and trend of all major rock units is approximately N60°E. Dips are generally steep to the southeast.

Ore bodies are bounded predominantly by hornblende-biotite gneiss containing locally abundant orthoclase and magnetite. Thin stringers of biotite granite and pegmatite are common. The ore bodies exploited in shaft No. 3½ are cut by two northwest-trending diabase dikes. These dikes are 5 cm and 45 cm wide, and dip uniformly at 73°NE.

Twelve gold ore bodies were exploited over a strike length of approximately 1500 meters along the Franklin vein. Outcrop locations of the nine northeastern-most ore bodies are shown in Figure 2. The ore bodies are series of lenticular shoots, pipes, or chimneys united by thin quartz stringers. These zones conform in dip and strike with the general schistosity of the host rocks, pitch approximately S80°E, and are bounded by clay-rich gouge. Individual ore bodies are generally small, attaining widths of up to 4.3

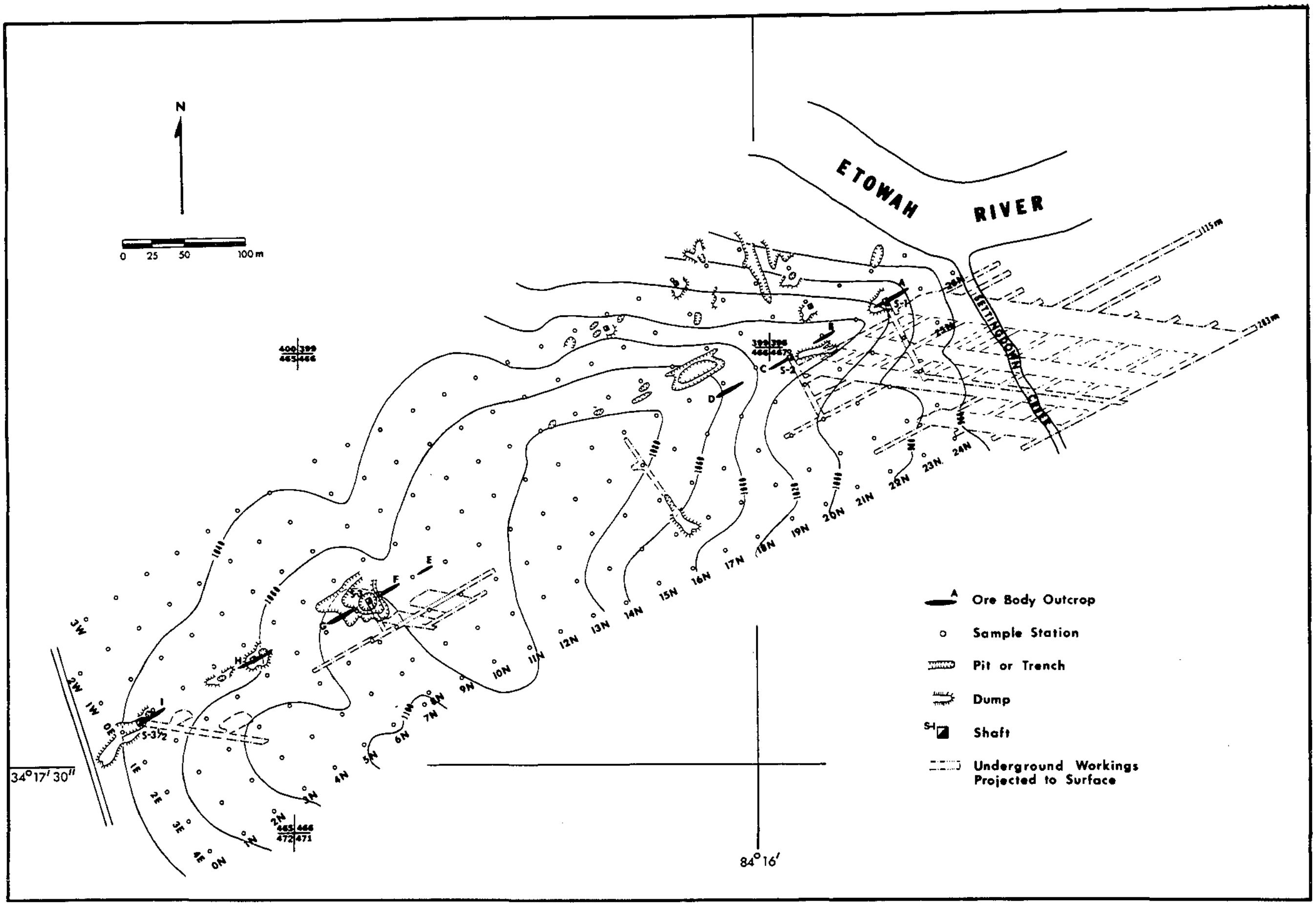

FIGURE 2. Grid area map with underground workings projected vertically to surface.

meters and lengths of about 30 meters. Similar gold mineralization occurs in a parallel zone known as the McDonald vein, approximately 30 meters northwest of the Franklin vein. Four other parallel gold-bearing zones have been prospected immediately southeast of the Franklin vein.

Perhaps the most significant characteristics of the individual ore zones are their extreme persistence with depth, and uniformity in grade and dimensions. Regularity of spacing along strike, persistence with depth, mineralogy, and presence of gouge and other cataclastic features suggest that the ore bodies occupy dilatant zones in a major shear.

Ore shoots within the Franklin vein consist of massive quartz containing both disseminated and coarse-grained, banded aggregates of pyrite and pyrrhotite. Arsenopyrite has not been observed but is a minor constituent of ores exploited at several mines a few kilometers to the northeast and in the Dahlonega district. Locally abundant accessory minerals are chalcopyrite, ankerite, and zoisite. The ore zones are locally cut, though not displaced, by fractures filled with coarse-grained calcite, pyrite, pyrrhotite, chalcopyrite, and tourmaline. Visible gold is of quite rare occurrence although mill data suggest that approximately 50% of the gold was free-milling, even in the unoxidized portions of the deposit. Estimated sulfide content of ore is between 5% and 10%. Mint data indicate a gold fineness from 980 to 989.

Data relative to the dimensions and grade of individual ore shoots are contained in a consulting report (1907) by former Georgia State Geologist, S. W. McCallie. Dimensions were determined by direct measurement during the period of active mining by the Creighton Mining Company (McCallie, 1907). Grade estimates are based on samples collected by McCallie and mill data. Data for the nine ore zones within the area of the soil geochemical grid are presented in Table 1.

Massive sulfide mineralization beyond the limit of the study area (Figure 1) is distinctly different from that of the relatively simple ore shoots of the Franklin vein. Formerly exploited massive sulfide zones were lenticular pods up to 245 meters long and 2.1 meters wide, lying parallel to the trend of, and completely enclosed within medium-grained albite-chlorite-quartz schist similar to altered rocks mapped in close association with massive sulfide deposits in Paulding and Haralson Counties (Hurst and Crawford, 1970). Mineralogy is dominantly granular pyrite with minor pyrrhotite. Diamond drilling has shown the presence of narrow polymetallic sulfide zones of unknown persistence. The best reported drill hole intersection is a 0.7 m width

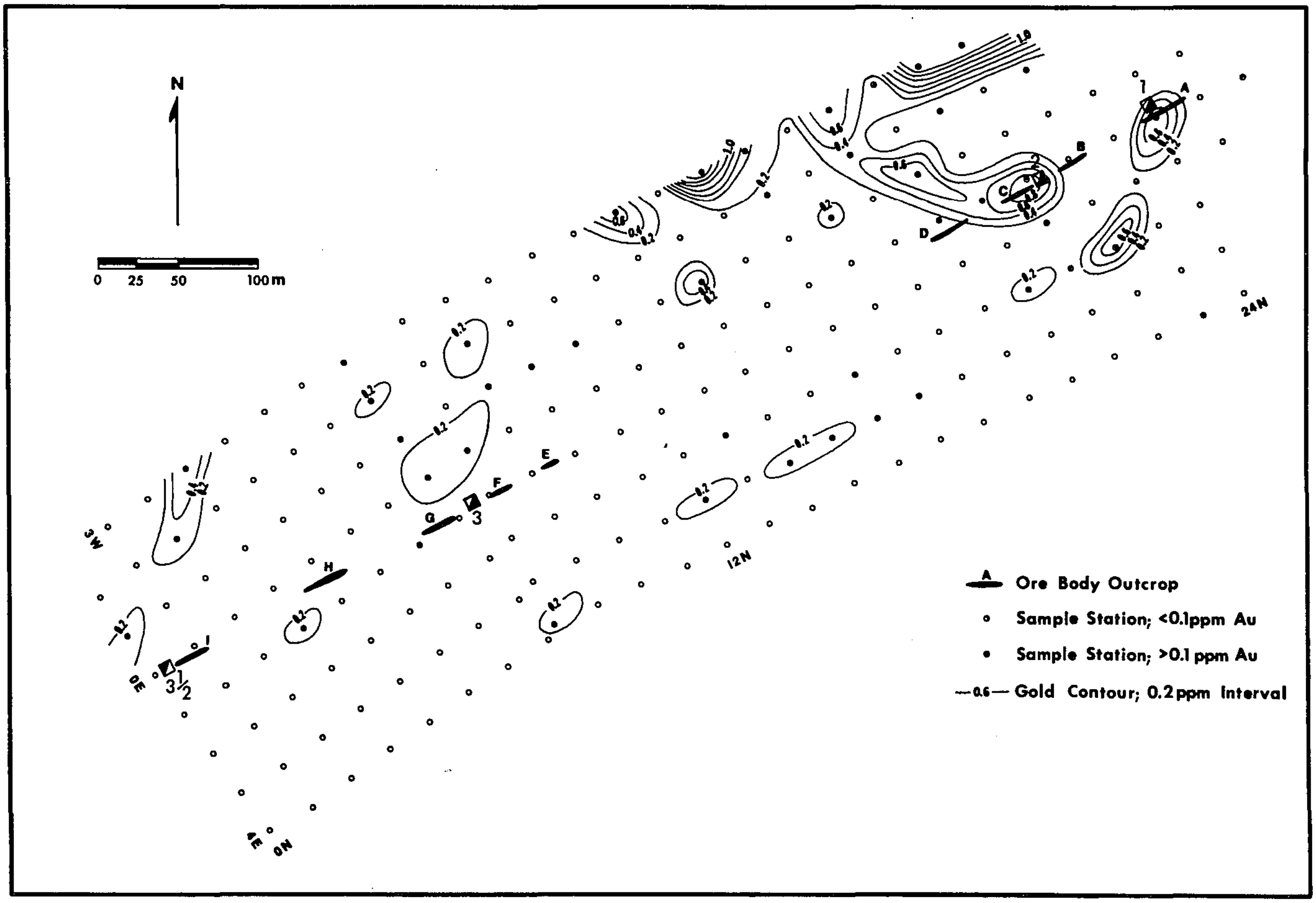

FIGURE 3. Contour map of "B" horizon soil gold content.

containing 9.08% zinc, 0.56% copper, 7.8 grams per ton silver, and 0.6 gram per ton gold. A genetic relationship, if any, between the adjacent, parallel massive sulfide horizons and gold mineralization associated with the Franklin vein has not been demonstrated.

SOIL GEOCHEMISTRY

A sampling grid of 198 stations on 30.5-meter centers was established along a N60°E base line extending for approximately 790 meters from a point 30 meters southwest of shaft No. 3½ to the confluence of Settingdown Creek and the Etowah River, approximately 60 meters northeast of shaft No. 1 (Fig. 2). Grid spacing was chosen to approximate the greatest outcrop dimension of the known ore zones. The grid encompasses the major workings along the Franklin vein (with the exception of shaft No. 4) and the projected locations of immediately adjacent, parallel gold mineralization both to the northwest and southeast.

Soil samples were collected from the upper portion of the "B" soil horizon with a stainless steel hand auger. Soil development is generally deep within the mine area. Saprolite is exposed in the deepest accessible portions of the slumped collar of shaft No. 3 and in the pit along lines 18N and 19N. Weathered muscovite schist is exposed in the face of the prospect adit at 16N-2E, a depth of 9 m below surface.

The -80 mesh fraction of each sample was analyzed for gold, copper, and arsenic by standard atomic absorption spectrophotometry. Detection limits are 0.02 ppm (parts per million) gold, 10 ppm copper, and 10 ppm arsenic. Maximum values reported are 2.7 ppm gold, 155 ppm copper, and 40 ppm arsenic. Data were analyzed by simple frequency distribution histograms and the log-normal probability technique of Lepeltier (1969).

Fifty-one percent of the samples contain detectable gold. Average background is slightly above the detection limit (0.02 ppm). Statistical analysis of data for samples containing detectable gold by the method suggested by Lepeltier (1969) indicate a significant gold threshold at 0.8 ppm. A second threshold at 0.4 ppm gold is suggested by a slight positive break in slope of the frequency distribution-log probability curve. This apparent bimodality may reflect bias in the sample population imposed by the restricted size of the sampling area relative to the wide overall distribution of anomalous values, and the high gold content of the grid area as a whole.

TABLE 1. Data for nine ore zones within the Franklin vein (McCallie, 1907): location of zones is indicated in Figure 2

Zone No.	Width (m)	Length (m)	Depth Developed (m)	gm/ton Gold
A	1.8	27	285	27.1
B	1.5	16	285	22.1
C	2.6	20	285	22.7
D	1.8	24	285	16.5
E	1.2	9	61	30.2
F	1.8	40	98	12.1
G	2.4	30	107	9.0
H	0.6	30	61	22.7
I	0.8	15	152	30.2

Average background and anomaly threshold for copper are 40 ppm and 80 ppm respectively. These values are consistent with unpublished data for reconnaissance soil geochemical surveys over similar rocks elsewhere in the Georgia Piedmont.

Arsenic does not exhibit an anomalous population. All samples with the exception of the initial sample (ON-OE; 40 ppm) contain 10 ppm or less arsenic. The high arsenic content of the initial sample is attributed to contamination of either the auger or sample preparation equipment by arsenic-bearing material encountered in a prior study.

The surface distribution of gold geochemical data (Fig. 3) suggests several pertinent correlations. As would be expected, soils in the area of most intensive prior mining (shafts No. 1 and No. 2) are characterized by an abnormally high gold content, however, similar high areas which do not correspond to the outcrop trend of the Franklin vein lie to the west, essentially beyond the area of former mining. An up-pitch projection of ore body "C" (Fig. 3) above the present erosional surface can be extended to a point in space vertically above the approximate location of anomalous samples 21N-OE, 20N-OE, 19N-1W, 18N-2W, and 18N-3W. This anomalous area, which at first glance might be interpreted to reflect significant mineralization along the parallel McDonald vein, may actually reflect only the residual concentration of particulate gold from the weathered and eroded, up-pitch portions of ore body "C". Results of underground exploration of the McDonald vein indicate that residual gold prospected along this anomalous trend does not reflect primary mineralization at depth and thus supports this interpretation. Such an interpretation can be expanded to account for similar, though less well defined, anomalous areas reflected by stations 21N-3W, 20N-3W, 19N-3W, 17N-1W, 16N-3W, and 15N-3W. High gold content in samples collected at stations 15N-3W and 13N-3W may, however, indicate the location of a parallel mineralized zone, residual accumulation from the weathered, up-pitch projection of ore zones exploited in shaft No. 2, or a small, previously unrecognized ore zone situated approximately midway between shafts No. 2 and No. 3. A poorly defined though persistent zone of high gold values is parallel to the Franklin "vein" to the southeast. This feature, reflected by samples 8N-4E, 12N-3E, 14N-3E, 15N-3E, 20N-2E, and 22N-2E, may correspond to what has been referred to in early unpublished reports as the Creek "vein".

Comparison of the dump locations with topography and gold geochemical contours suggests that, for the most part, there has been little contamination of the gold geochemical signature by this material. On the other hand, there is an indication, particularly in the vicinity of shafts No. 3 and No. 3½, that high gold values are generally displaced down the topographic slope from ore body outcrops.

A general relationship between gold content of soil and grade of primary mineralization at depth is implied by comparison of geochemical data with the ore zone grade data of Table 1. Soil in the immediate vicinity of ore bodies at shafts No. 1 and No. 2 contain up to 1 ppm gold, approximately 3 to 4 percent of the gold content of unweathered ore. Similarly, soil in the area of lower grade ore bodies exploited at shaft No. 3 exhibit the same 3 to 4 percent relationship. If the projected outcrop location of similar mineralization in a deeply weathered prospect area is known, gold content of overlying soils may be used to roughly estimate the potential grade of primary mineralization at depth.

As might be expected, "B" horizon soil values represent a substantial depletion or dilution relative to grade of primary ore, apparently due to migration and dispersion of gold outward from the outcrop during development of the soil profile. Such outward dispersion is indicated by examination of the size of anomalous areas in comparison to the known cross-sectional areas of the ore bodies

(Fig. 3).

Residual enrichment of gold within specific, upper portions of soil profiles in the southeastern United States has been described by Kinkel and Lesure (1968), Cook (1970), and Lesure (1971). While these studies adequately explain the origin of residual surface and saprolitic ores formerly exploited in the southeast, they were conducted over areas in which primary mineralization was confined to small, erratic zones and did not relate soil values to well defined, uniform ore mineralization at depth. Although the gold content of soils overlying and in the immediate vicinity of the ore zones along the Franklin vein is substantially less than that of the primary ore, the gold content of these soils is quite high in comparison with the average background (0.02 ppm). It is significant in this regard that the highest gold geochemical values reported in this study are on the steepest topographic gradient of the grid area, generally downslope from the ore bodies exploited in shafts No. 1 and No. 2. These values imply residual enrichment of the "B" horizon during soil profile and topographic evolution.

There is general correspondence between the distribution of anomalous copper in soil and the trend of the Franklin "vein" as well as the reported parallel mineralization to the southeast (Fig. 4). Close agreement exists between anomalous copper and the outcrop location of ore bodies exploited in shafts No. 3 and No. 3½. In areas of relatively steep topography zones of anomalous copper appear to trend down slope from projected ore body outcrops. This is particularly striking immediately north of shafts No. 1 and No. 2 where the southern boundary of an apparent west-trending anomalous zone cuts across the general strike of the host rocks but corresponds in position to the 1000-foot topographic contour (Fig. 2 and 4).

A small, previously unrecognized mineralized zone approximately midway between shaft No. 3 and shaft No. 2 is suggested by anomalous copper centering at sample station 15N-0E and the gold content of samples 14N-1W and 13N-3W. The presence of ore in this vicinity is predictable based on the regular spacing of other ore zones along the Franklin vein.

CONCLUSIONS

The surface distribution of gold and copper in soil is complex, and relationships between anomalies and primary mineralization are not straightforward. Individually, neither gold nor copper contours adequately define known

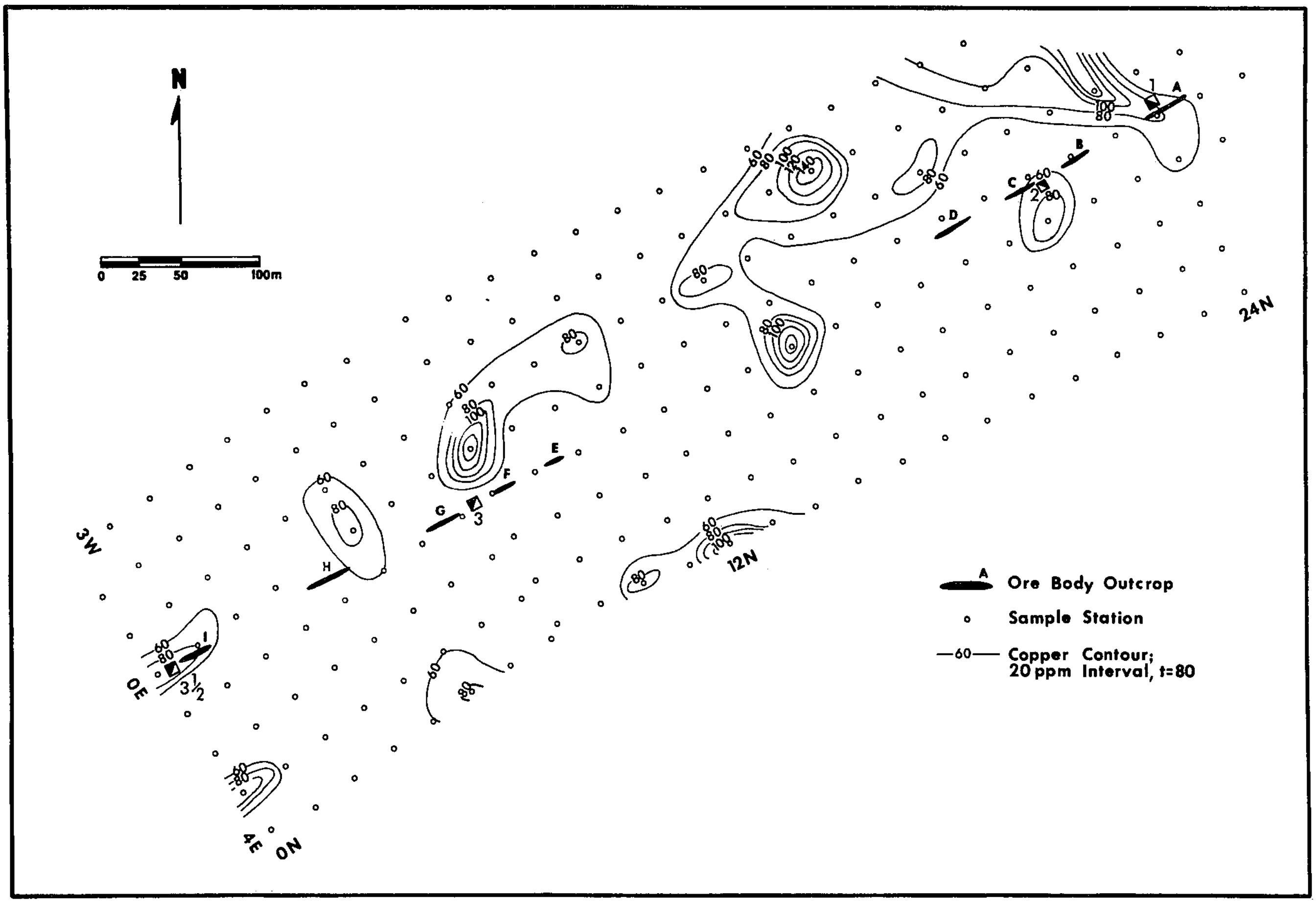

FIGURE 4. Contour map of "B" horizon soil copper content.

mineralization, although combined gold and copper soil geochemistry roughly reflects the general trend of the Franklin vein and the general locations of ore zones exploited by shafts No. 1, No. 2, No. 3, and No. 3½. Should soil geochemistry be the only exploration tool utilized prior to drilling, careful consideration must be given to anomaly offsets due to dispersion and possible migration during soil evolution, topography, and residual elemental accumulation downward from previously weathered, up-pitch portions of drill targets. The ore zones exploited by shafts No. 1 and No. 2 could have been easily missed in initial drill holes due to apparent anomaly displacement by residual accumulation of gold from previously weathered portions of the pitching ore shoots.

The indicated general relationship between gold content of soil in the immediate vicinity of ore zone outcrops and ore grade at depth is depletion by weathering in the ratio of approximately 1:25. While this relationship is considered applicable only for the immediate Franklin-Creighton mine vicinity, similar depletion of dispersion factors could be generated for specific project areas in conjunction with preliminary drilling. While the importance of determining this relationship is obvious, particular care must be exercised in the southeastern United States where specific soil types and weathering phenomena often lead to the abnormal accumulation of gold in the upper few feet of the soil horizon and attendant, potentially misleading geochemical data.

Although the copper content of ore zones is apparently quite low, copper geochemistry was found to compliment gold geochemical data, suggesting that copper content of soils be determined routinely in conjunction with precious metal geochemical exploration of similar occurrences in the southeastern United States.

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ORE MINERALOGY OF WEST-CENTRAL GEORGIA MASSIVE SULFIDE DEPOSITS

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ABSTRACT

Numerous massive sulfide deposits within the complex sequence of metavolcanic and metasedimentary rocks of west-central Georgia have been known and occasionally exploited since their discovery in the mid-1800's. Modern exploration programs including extensive diamond drilling have supplied sufficient data for the detailed description of the mineralogy of four of the more promising occurrences: the Little Bob, Swift, Tallapoosa, and Villa Rica deposits.

Mineralogically and texturally, ores of the Little Bob, Swift, and Villa Rica deposits are similar to those of the Ducktown, Tennessee, district consisting of abundant pyrite and pyrrhotite with lesser sphalerite, chalcopyrite, galena, cubanite, and valleriite. Variable amounts of supergene covellite and chalcocite occur in near surface ores. Magnetite and ilmenite are locally common. Gangue minerals include quartz, calcite, various amphibole and chlorite group minerals, gahnite, and pyralspite garnet. Rock and ore textures indicate several periods of cataclastic deformation as well as possible annealing recrystallization.

The Tallapoosa massive sulfide deposit is significantly different from the other deposits due primarily to the absence of pyrrhotite and the relative abundance of gangue dolomite. Accessory chalcopyrite, sphalerite, galena, tetradymite, gold, chalcocite, and covellite are locally conspicuous. Cataclastic textures are much less prominent than in the other deposits and there is a general absence of textures indicative of stress-induced or annealing recrystallization. Dominant gangue minerals are chlorite, quartz, and dolomite.

The complex post-depositional history of Little Bob, Swift, and Villa Rica ores, as indicated by textures, leaves the origin of the deposits open to considerable debate. The association of these three deposits with known metavolcanic rocks suggests that they may be classical volcanogenic massive sulfide deposits and that many textural features now present should be ascribed to post-depositional shearing and metamorphic episodes, rather than to original deposition. It is difficult, however, to attribute the Tallapoosa deposit to such an origin. This deposit gives every indication of being a simple shear zone replacement, post-dating or perhaps coinciding with the last major metamorphic event. Similarities in the four deposits likewise suggest that they may all be ascribed to shear zone replacement by hydrothermal sulfide-bearing fluids with post-depositional recrystallization of the Little Bob, Swift, and Villa Rica ores being the result of later readjustment during a period of metamorphism that generally diminished in intensity to the northwest, becoming of insufficient magnitude to affect ores of the Tallapoosa occurrence.

INTRODUCTION

Numerous massive sulfide occurrences have been prospected within the complex sequence of metavolcanic and metasedimentary rocks in west-central Georgia, predominantly in Paulding, Carroll, Douglas, and Haralson Counties. These deposits were first considered as potentially economic sources of copper during the short-lived period of extensive prospecting initiated by the discovery and development of the famous ore deposits at Ducktown, Tennessee in the mid 1800's. Several of the more promising occurrences, particularly the Little Bob, Swift, Tallapoosa and Villa Rica mines, were exploited in the early 20th century for the production of sulfuric acid from pyrite and pyrrhotite.

The better known sulfide mineral occurrences have been evaluated by industry in recent years as potential sources of copper, zinc, and the precious metals. This exploration has included at least 103 diamond drill holes, one program resulting in the acquisition of 26,303 feet of core from the Little Bob, Swift, and Tallapoosa mines alone.

The object of this paper is to briefly review the exploration and production history, and to present the detailed ore mineralogy of the better known occurrences.

The pyrite deposits of Georgia were first described by Shearer and Hull (1918) during the time of maximum sulfide production to meet the needs of World War I. These data have been reviewed and expanded by Hurst and Crawford (1970) to include drilling logs of the extensive New Jersey Zinc Company exploration program conducted in the early 1950's.

The detailed mineralogy and petrology of the four deposits discussed herein were the topic of a dissertation by the author (Cook, 1970). Selected diamond drill core sections representative of 26,303 feet of core obtained at the Little Bob, Swift, and Tallapoosa mines during the New Jersey Zinc Company exploration program, and selected samples from the Villa Rica mine acquired from the Tennessee Copper Company are the basis of this study. Ore and mineralized wall rock were examined in detail in thin and polished sections. Mineral identification is based on reflected and transmitted light optics, and x-ray defraction and electron microprobe analysis where appropriate.

EXPLORATION AND PRODUCTION HISTORY

Little Bob Mine. The Little Bob massive sulfide deposit centers within land lot 65, second district, 3rd section, of Paulding County, approximately 2 miles northwest of Hiram. The first production was in 1885 when approximately 1000 tons of pyritic ore were sold for its sulfur content. The mine was operated by Hannah Minerals Company continuously from April, 1917 through September, 1919. Production estimates for this period vary greatly from more than 200,000 (Hannah Minerals Company records, unpublished) tons to approximately 13,000 tons. A consulting report by Joel H. Watkins (1939) includes a monthly output summary for this period indicating a total production of 14,570 tons of ore. Calculation of ore body size from old mine maps indicates that no more than 20,000 tons could have been removed.

A total of 43 diamond drill holes have been put down in the deposit during four separate exploration programs. The most extensive program was that conducted by New Jersey Zinc Company at which time 24 holes were drilled along the trend of the deposit. Drilling indicates the presence of a complex series of massive sulfide lenses enclosed in and conformable with intermediate-to-mafic metavolcanic rocks. A moderate quantity of ore containing marginal amounts of copper, zinc, gold, and silver is indicated by results of the various drilling programs.

Swift Mine. The Swift massive sulfide deposit is within the 19th district, 3rd section of extreme southwestern Paulding County, approximately 1.5 miles east of Draketown and only a short distance from the Paulding-Haralson County line. Initial copper prospecting began on the property in 1858. The only reported production was in 1905 when Swift and Company shipped approximately 300 tons of massive pyrite-pyrrhotite ore for the production of sulfuric acid.

The property has been explored by diamond drilling on two separate occasions. The most extensive program was that of the New Jersey Zinc Company which resulted in the drilling of 15 core holes. This drilling program delineated a small massive sulfide lense enclosed within mafic metavolcanic rocks. Local high grade drill hole intersections contained substantial zinc and lesser copper mineralization.

Tallapoosa Mine. The Tallapoosa massive sulfide deposit is within the 20th district, 3rd section, of the northeast corner of Haralson County, approximately 2.5 miles north of Draketown and 1500 feet northeast of the Tallapoosa River. Extensive copper prospecting was carried out in 1874 by the Middle Georgia Mineral Association, although no production was reported until 1881 when approximately 22,000 tons of ore were shipped.

The property has been explored by diamond drilling on four separate occasions. The United States Bureau of Mines evaluated the property in 1947 (Ballard and McIntosh, 1948) at which time four holes were put down. Eight additional holes were drilled in 1954 by the New Jersey Zinc Company. Several more core holes were drilled by the Tennessee Copper Company in 1960. Drilling data indicate a rather small though persistent vein-like lens of massive sulfide ore within thinly bedded actinolite-quartz-chlorite schist and granular quartzite.

Villa Rica Mine. The Villa Rica massive sulfide deposit, the most extensively developed in the district, is in the northwestern corner of Douglas County, 3 miles north-northwest of Villa Rica and adjacent to Georgia Highway 61. Active exploration of this deposit began in 1890 when a 300 foot shaft was sunk and considerable drifting carried out. The property was acquired by the Virginia-Carolina Chemical Company in 1895. Production was continuous from 1899 until June, 1917. Ore was produced to a depth of 500 feet. Examination of mine maps indicates that a minimum of 280,000 tons of ore were produced during this period.

The property was diamond drilled in 1960 by the Tennessee Copper Company. This and other work defined the presence of a vein-like lens of massive sulfides within hornblende-rich metavolcanic rocks which are cut by numerous, narrow pegmatite dikes. Copper, zinc, and precious metal values are generally low although local, apparently small zones contain significant amounts of chalcopyrite and sphalerite.

MINERALOGY

Mineralogically and texturally, ores of the Little Bob, Swift, and Villa Rica deposits are quite similar to ores of the Ducktown, Tennessee district as described by Carpenter (1965). Primary sulfide minerals in decreasing order of abundance are pyrite, pyrrhotite, sphalerite, chalcopyrite, galena, cubanite, and valleriite. Variable amounts of supergene covellite and chalcocite are conspicuous in near-surface ores. Magnetite and ilmenite are locally common, particularly in the Little Bob and Swift deposits. Gangue is dominantly quartz, locally abundant calcite, large irregular masses and cataclastically deformed fragments of wall rock, individual grains of minerals comprising altered wall rock, both euhedral and poikilitic gahnite, and pyralspite garnet. Several periods of cataclastic deformation as well as possible annealing recrystallization are indicated texturally.

The Tallapoosa massive sulfide deposit is significantly different from the other deposits due primarily to the absence of pyrrhotite and the relative abundance of gangue dolomite. Ore of this deposit consists of very abundant pyrite with minor chalcopyrite, sphalerite, galena, tetradymite, gold, and locally important supergene chalcocite and covellite. Cataclastic textures are much less prominent in Tallapoosa ore than in the other deposits and there is a general absence of textures indicative of stress-induced or annealing recrystallization. Dominant gangue minerals are chlorite, quartz, and dolomite.

Pyrite - FeS_2. Pyrite of the Little Bob, Swift and Villa Rica deposits occurs as cubes, angular fragments of broken grains, and irregular masses in a matrix of other sulfides, quartz, and carbonates. Grain size varies from 9.0 mm for unbroken cubes to less than 0.1 mm in zones in intense deformation. Cataclastic deformation of pyrite is most evident within the Little Bob deposit, but is locally conspicuous in the Swift and Villa Rica ores. Pyrite grains of the more massive ores exhibit sheeting (Kinkel, 1962), irregular internal fracturing, or peripheral shattering when in contact with other pyrite grains. Elongate fragments are

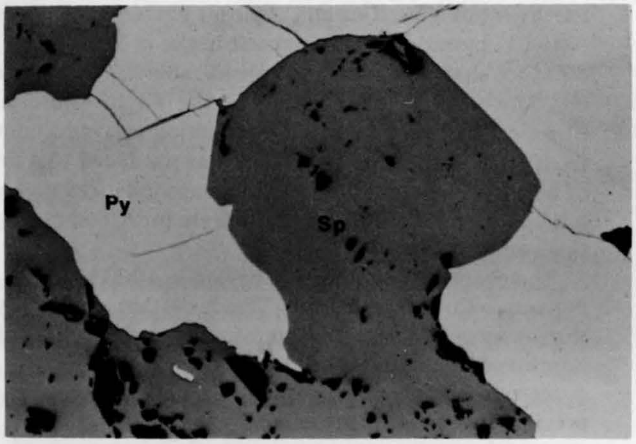

FIGURE 1. Sphalerite (Sp) embayment into pyrite (Py); Little Bob deposit.

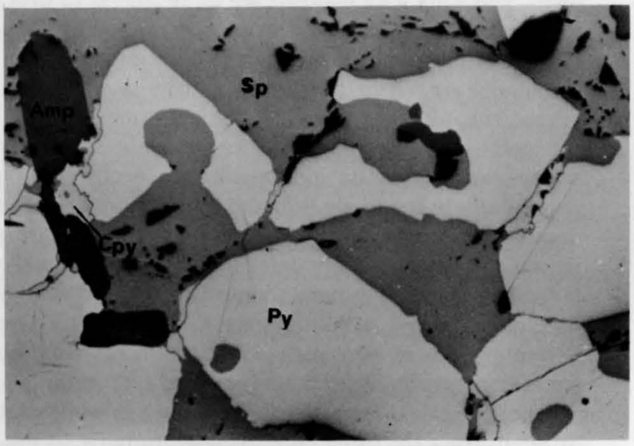

FIGURE 2. Sphalerite (Sp) deeply embaying pyrite (Py). Note lack of chalcopyrite exsolution blebs and lamellae within sphalerite, although chalcopyrite (Cpy) separates pyrite grains and partially replaces amphibole (Amp); Little Bob deposit.

commonly oriented parallel to the local schistosity and fractured perpendicular to elongation of grains. Other fragments of shattered grains appear to have floated out into surrounding sphalerite and chalcopyrite.

Embayment and veining of pyrite by chalcopyrite and sphalerite are locally extensive (Fig. 1, 2, and 3). Irregular remnants of pyrite euhedra in coarse-grained sphalerite matrix are typically coated by narrow rims of chalcopyrite. Conversely, partially replaced pyrite grains in chalcopyrite matrix are surrounded by narrow rims of sphalerite. Such textures have been shown to result from migration of pre-existing sulfide phases during their formation (Roberts, 1965). Atol textures (Edwards, 1947) are locally developed between subhedral pyrite and matrix sphalerite. Slight embayment of euhedral pyrite by galena and interstitial quartz is common.

Pyrite grains are characterized by a large variety of inclusions. Euhedral magnetite grains up to 0.5 mm are common in undeformed pyrite. Hornblende, chlorite, and biotite occur in parallel orientation within larger pyrite grains. The orientation of these lineated inclusions corresponds with local foliation of the country rocks in weakly deformed or undeformed zones; however, a random orientation exists between inclusions in individual pyrite fragments in zones of intense deformation. Many of the included biotite and hornblende grains are corroded and veined along cleavage planes by pyrite. These relationships suggest that pyrite crystallized after regional metamorphism and was later subjected to local shearing, causing a displacement of individual grains.

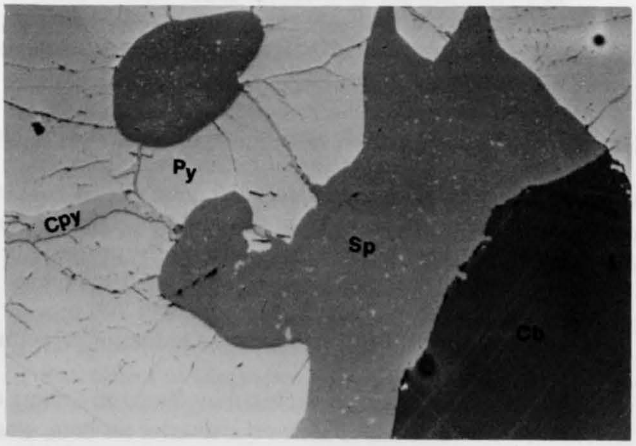

FIGURE 3. Apparent core and rim replacement of cataclastically deformed pyrite (Py) by sphalerite (Sp). Chalcopyrite (Cpy) fills fractures in pyrite. Note minor replacement of carbonate (Cb) along twin planes by sphalerite; Little Bob deposit.

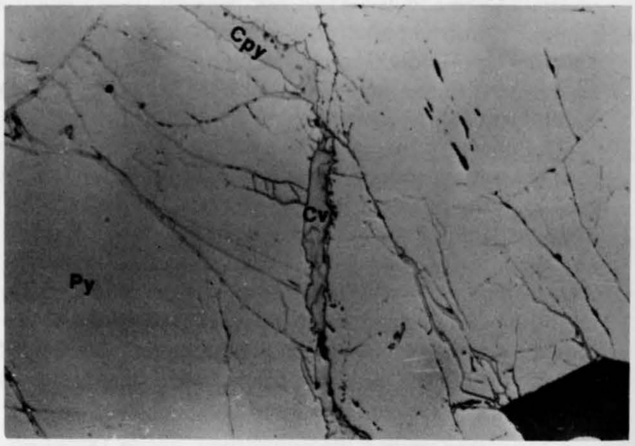

FIGURE 4. Cataclastically deformed pyrite (Py) "healed" by chalcopyrite (Cpy). Chalcopyrite is partially replaced by covellite (Cv); Tallapoosa deposit. Note quartz (Q) grain.

Carbonate grains and euhedral quartz grains averaging 0.05 mm in length are locally included in pyrite. Small ellipsoidal inclusions of sphalerite, pyrrhotite, chalcopyrite, and galena are characteristic of pyrite-rich zones throughout the deposits. Bi-mineral inclusions of sphalerite and chalcopyrite are common. Sphalerite, chalcopyrite, and pyrrhotite locally co-exist within single inclusions, particularly in the Little Bob and Swift ores. Sphalerite inclusions within unshattered pyrite grains exhibit fine networks of chalcopyrite blebs and lamellae. Such textural relationships imply an early deposition of pyrite with subsequent crystallization of sphalerite, chalcopyrite, and pyrrhotite.

Pyrite is the only iron sulfide mineral present in the Tallapoosa deposit, locally constituting up to 85% of the massive ore. Pyrite occurs as medium-grained, close-packed aggregates of broken grains (Fig. 4) and as euhedral cubes isolated within dolomite-rich or chloritic wall rock. Fragmented pyrite grains range up to 8.0 mm in diameter and are commonly cemented by interstitial dolomite and quartz. Pyrite cubes up to 1.0 cm across occur in gangue-rich portions of the ore and commonly include well foliated chlorite and actinolite masses.

Cataclastically deformed pyrite of the Tallapoosa deposit is commonly veined and embayed by sphalerite, chalcopyrite (Fig. 4) and galena. Small peripheral fragments of large, shattered pyrite grains exhibit slight displacement into matrix or interstitial dolomite, sphalerite, and chalcopyrite. Such textures suggest that pyrite crystallized early in the depositional sequence, preceding the other sulfides.

Pyrite of the Tallapoosa mine is characterized by numerous inclusions of both gangue and ore minerals. Ovate inclusions of sphalerite, chalcopyrite, and galena are present in most pyrite grains of the ore zone but are rare in wall- and country-rock pyrite. Highly irregular inclusions of quartz and dolomite are locally abundant. Euhedral magnetite grains averaging approximately 0.1 mm in diameter are included in pyrite, but are of much rarer occurrence than in the other three deposits.

Pyrrhotite - $Fe_{1-x}S_x$. Pyrrhotite is very abundant in ores of the Little Bob, Swift and Villa Rica deposits, but is completely absent in the Tallapoosa deposit. The composition of 25 pyrrhotite samples from the 3 deposits as determined by measurement of the d(102) spacing reflect a range of 47.29 to 47.79 atomic percent iron. All pyrrhotite exhibits hexagonal symmetry.

Pyrrhotite occurs in three distinct modes in ores of the Little Bob deposit. It occurs abundantly as rounded inclusions in pyrite grains and frequently co-exists in these inclusions with chalcopyrite, sphalerite and magnetite. Pyrrhotite also occurs abundantly as foliated masses near massive sulfide zone margins and in immediately adjacent, altered wall rock (Fig. 11, 12). A third mode of pyrrhotite occurrence is in highly irregular intergrowths with chalcopyrite and late-stage magnetite near the central portions of thicker massive sulfide zones.

Pyrrhotite occurs in the massive sulfide ore of the Swift deposit as coarse-grained interlocking aggregates. Pyrrhotite grains contain randomly distributed inclusions of slightly corroded, coarse-grained subhedral pyrite in textures reminiscent of Ducktown ore. Pyrrhotite is commonly veined and embayed by both sphalerite and chalcopyrite. Individual pyrrhotite grains locally contain blebs of chalcopyrite concentrated along basal planes. This texture may represent chalcopyrite exsolution from pyrrhotite or preferential pyrrhotite replacement by chalcopyrite along susceptible planes.

Massive and foliated pyrrhotite occurs throughout the ore zone of the Villa Rica deposit. The more massive ore is dominated by large aggregates of 0.8 by 0.4 mm interlocking pyrrhotite grains. This pyrrhotite is characterized by inclusions of corroded, subhedral pyrite and numerous amphibole, biotite and chlorite grains. The included pyrite grains reach a maximum size of 1.0 cm in diameter and are randomly distributed throughout the matrix pyrrhotite. Biotite, chlorite, and amphibole inclusions are typically veined and embayed by pyrrhotite. Larger pyrrhotite grains contain small ovate inclusions of chalcopyrite concentrated along basal planes.

Foliated masses of pyrrhotite are intergrown with well aligned silicate grains in gangue-rich portions of Villa Rica ore (Fig. 11, 12). This pyrrhotite is similar to that of the sheared wall rock, consisting of elongate aggregates of equigranular, interlocking grains. Grain sizes range from 0.04 to 0.07 mm in diameter. Twinning and inclusions of other grains are rare.

Foliated pyrrhotite of the wall rock and pyrrhotite of cataclastically deformed zones in more massive ore of the Little Bob, Swift, and Villa Rica deposits show numerous deformation effects. Deformation twinning, undulose extinction, preferred orientation of grains and twins, and possible sub-grain formation are typical Tapered deformation twins occur either perpendicular or parallel to the inferred plane of shearing. These twin lamellae locally become close-spaced, eventually giving way to patches of untwinned grains. Patches of apparently undeformed pyrrhotite within zones of intense deformation may be the result of stress-induced recrystallization of subgrain formation (Stanton and Gorman, 1968).

A study of the possible effects of annealing recrystallization by triple junction point angle measurements (Stanton and Gorman, 1968) was performed on polished sections containing abundant pyrrhotite. These data suggest that coarse-grained, equigranular pyrrhotite commonly forming the matrix for coarse-grained, subhedral pyrite has undergone annealing recrystallization. Triple junction point measurements for fine-grained, equigranular pyrrhotite aggregates in sheared massive sulfide ore, weakly foliated wall rock, and massive sulfide zone margins returned inconsistent data.

Sphalerite - ZnS. Ore of the Little Bob, Swift, and Villa Rica deposits locally contain very abundant sphalerite. Sphalerite of these deposits exhibits identical morphological, textural, compositional, and association characteristics, and differs markedly from sphalerite of the Tallapoosa deposit.

The iron content of sphalerite was determined by cell edge measurements and confirmed by electron microprobe analysis. Sphalerite of the Tallapoosa deposit contains an average of 4.8 mole percent FeS as compared with 11.9, 14.0, and 13.0 percent for the Little Bob, Swift, and Villa Rica deposits respectively.

Sphalerite of the Little Bob, Villa Rica, and Swift deposits occurs predominantly as an interstitial matrix for pyrite grains (Fig. 1, 2, and 3), and as individual crystals and groups within gangue-rich ore and immediately adjacent wall rocks. Sphalerite of the massive sulfide ores exhibits well defined embayment of pyrite (Fig. 1, 2, and 3). Shattered pyrite grains are commonly veined by fine-grained sphalerite. Individual pyrite fragments in matrix sphalerite show displacement with respect to their position in the original coherent grain. This feature is unusual in regard to the low crystallizaing power of sphalerite and may be the result of post-ore recrystallization (Gill, 1969).

Magnetite, ilmenite, quartz, calcite, and fractured and bent chlorite, sericite, biotite and hornblende are commonly included within individual sphalerite grains. These minerals show surface corrosion, embayments, and veining along cleavage planes.

Crystallographically oriented blebs and lamellae of chalcopyrite are common features of the sphalerite in these deposits. This texture has been described in Ducktown ores (Carpenter, 1965) and is believed to be the result of chalcopyrite exsolution from sphalerite. Exsolved chalcopyrite may be evenly distributed throughout the host grain (Fig. 3) or concentrated toward the central portions. Other sphalerite grains exhibit barren margins but are bounded by irregular chalcopyrite masses (Fig. 2). Such textures suggest both core and rim replacement of sphalerite by chalcopyrite, but may be the result of crystallization and mobilization or exsolution-migration as pointed out by Roberts (1965). Gangue mineral grains included in sphalerite are coated by thin films of chalcopyrite, with a corresponding decrease in exsolution blebs and lamellae in the immediate vicinity. Such textures may be the result of chalcopyrite migration during slow cooling, or mobilization during deformation or recrystallization (Roberts, 1965; Gill, 1969). Shattered sphalerite grains are locally healed and slightly corroded or embayed by chalcopyrite.

Sphalerite of the Little Bob, Swift, and Villa Rica deposits exhibits a high incidence of twinning when etched with nitric and iodic acids. Broad primary twins are conspicuous. Such twins are commonly 0.5 mm wide, transect entire crystal grains and show no preferred orientation with respect to adjacent twinned crystals. They occur as single parallel groups across individual grains or as complex triangular networks oriented parallel to planes of chalcopyrite exsolution. Other twins, believed to be the result of deformation, characterize zones of cataclasis. These twins constitute narrow, close-spaced networks of parallel lamellae, often terminated before reaching the grain boundary, and exhibit imperfect parallel alignment between aggregates of grains.

Deformation effects are well developed in sphalerite of the Little Bob, Swift, and Villa Rica deposits. Primary twin lamellae are often bent and warped as are chalcopyrite exsolution lamellae. Aggregates of small sphalerite grains averaging 0.2 mm in diameter occur at the margins of highly deformed masses of larger sphalerite grains. Deformation twins within the larger grains terminate at the margins of the fine-grain aggregates. These fine-grained aggregates contain narrow twin lamellae identical to primary twins, but no chalcopyrite exsolution lamellae. Such grains may be the result of sub-grain formation during deformation. The fine twin lamellae may represent post-ore annealing.

Annealing recrystallization of Little Bob and Swift deposits sphalerite is suggested by the results of triple junction point angle measurements resulting in a standard deviation of 11 for granular, medium-grained sphalerite from the central portion of a 145 foot ore zone in the Little Bob deposit, and a standard deviation of 12.5 for equigranular, medium-grained sphalerite from the Swift deposit.

In the Tallapoosa deposit, sphalerite is most abundant in dolomitic zones. Irregular veins of sphalerite penetrate and embay crystalline aggregates of dolomite along grain boundaries. Cataclastically deformed zones contain twinned dolomite rhombohedra included within masses of sphalerite. These dolomite grains exhibit surface corrosion and embayment along twin planes. In pyrite-rich zones, sphalerite locally forms an intergranular matrix for shattered pyrite fragments.

Sphalerite of the Tallapoosa deposit is characterized by abundant, well-oriented lamellae and blebs of chalcopyrite. Textures suggestive of chalcopyrite migration to sphalerite grain boundaries, cores, and the surface of inclusions are conspicuously rare in Tallapoosa ore. The textural evidence suggests that the chalcopyrite blebs and lamellae formed by exsolution from sphalerite, but that post-ore conditions were not of sufficient intensity to produce chalcopyrite migration.

Twinning is conspicuously rare in Tallapoosa mine sphalerite in comparison with the other deposits. Narrow deformation twins occur only in sphalerite bordering cataclastic zones. No twins suggestive of annealing were observed and triple junction point angle measurements on sphalerite of this deposit exhibit a randomly distributed population contraindicative of annealing recrystallization.

Chalcopyrite - $CuFeS_2$. Chalcopyrite is a locally important, primary constituent of both the ores and altered wall rock of the four deposits. It occurs predominantly in sphalerite-rich zones and, in the Little Bob deposit, intergrown with pyrrhotite and late-stage magnetite. Lamellar twinning is often conspicuous, although other effects attributable to deformation are uncommon. Valleriite and cubanite lamellae are present in larger chalcopyrite masses and are probably the result of exsolution.

Replacement textures are well developed between chalcopyrite and pyrite, hornblende, biotite, sericite and carbonates. Included hornblende, biotite and sericite grains show surface corrosion, embayments, and extensive veining along cleavage planes. Carbonate grains are embayed by chalcopyrite, particularly along cleavage and twin planes

(Fig. 5, 6). Shattered pyrite grains are distinctly corroded, embayed, and veined in zones of cataclasis (Fig. 4). Chalcopyrite is locally embayed by galena and veined by supergene chalcocite and covellite.

Galena - PbS. Very minor galena is characteristic of sphalerite-rich portions of the Little Bob, Swift, and Villa Rica deposits. Galena typically occurs as 0.1 mm rounded inclusions in pyrite, irregular patches and segregations within sphalerite, and embayments and veins in cataclastically deformed chalcopyrite and pyrite. Individual galena grains commonly exhibit cubic step-like growth into adjacent carbonate grains. Close-packed, cataclastically deformed pyrite grains are locally invaded by interstitial galena in ores of the Swift deposit. The effects of mechanical deformation are inconspicuous or absent in galena of these ores. Etching with iodic and nitric acids reveals no twinning.

Although a minor ore mineral, galena is more common in ores of the Tallapoosa mine where its abundance locally approaches 1 percent. It occurs as irregular masses up to 3.0 mm in diameter within aggregates of sphalerite and chalcopyrite. Galena also occurs within pyrite grains as separate or bi-mineral inclusions with sphalerite and chalcopyrite. Textural relationships suggest that galena crystallized slightly later in the depositional sequence than chalcopyrite and sphalerite.

Cubanite - $CuFe_2S_2$. Chalcopyrite-rich portions of the Little Bob, Swift, and Villa Rica deposits contain minor cubanite. The mineral is characterized by a color intermediate between chalcopyrite and pyrrhotite, moderate anisotropism with blue-white polarization colors, and a Vicker's hardness of approximately 200.

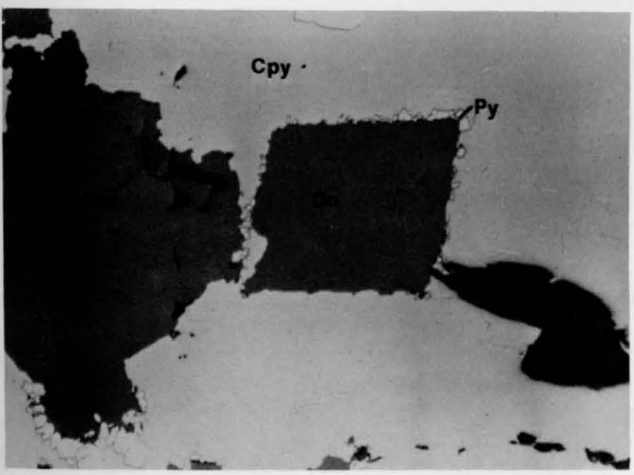

FIGURE 5. Euhedral dolomite (Do) partially replaced by chalcopyrite (Cpy). Dolomite crystal is rimmed by apparently cataclastic pyrite (Py) fragments; Tallapoosa deposit.

Cubanite occurs as tapered lamellae and threads in chalcopyrite and as irregular, anhedral grains around the margins of chalcopyrite and pyrrhotite aggregates. Chalcopyrite grains locally contain cubanite lamellae arranged radially around portions of the grain boundaries. Cubanite lamellae reach maximum lengths of 0.25 mm. Similar cubanite and chalcopyrite have been described for Ducktown ores by Carpenter (1965) and may be the product of solid solution unmixing (Ramdohr, 1960). The granular phase may be the product of direct crystallization or the migration and recrystallization of exsolved cubanite.

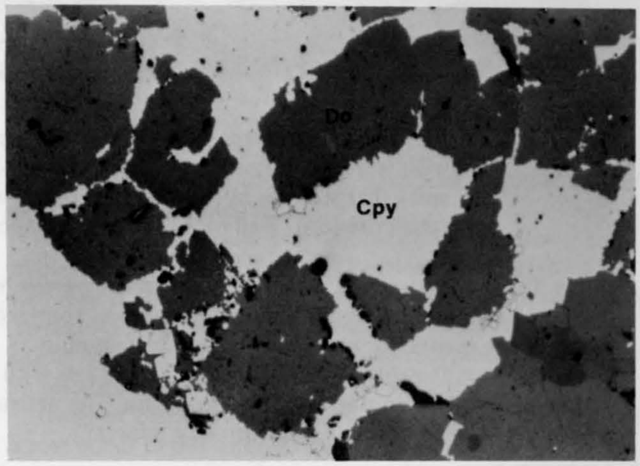

FIGURE 6. Dolomite (Do) replacement by chalcopyrite (Cpy); Tallapoosa deposit.

Valleriite - $4(Fe,Cu)S \cdot 3(Mg,Al)(OH)_2$. Valleriite has been tentatively identified as a very minor constituent of both Little Bob and Swift mine ores. Identification is based on optic properties and hardness. In reflected light the mineral is light brown and exhibits extreme anisotropism from white to dark grey. Its hardness is approximately that of chalcopyrite. The mineral occurs as oriented, lamellar intergrowths within chalcopyrite. The lamellae are generally less than 0.1 mm wide and lie en echelon within the enclosing chalcopyrite. Textural relations suggest that the valleriite is the product of exsolution.

Tetradymite - Bi_2Te_2S. Tetradymite is a very minor constituent of Tallapoosa mine ore. It occurs as 0.01 mm long, lath-like inclusions within most galena grains and is distinguished from galena by its creamy white color, moderate anisotropism, and lamellar twinning. Electron microprobe analysis, corrected for absorption and atomic weight effects, yields the following composition: Bi = 57.2%, Te = 35.2%, and S = 6.5%.

Gold - Au. Low though persistent gold values characterize the massive sulfide ores of the four deposits. Native gold was observed in a single polished surface of chalcopyrite-rich ore taken from New Jersey Zinc Company drill hole number 11

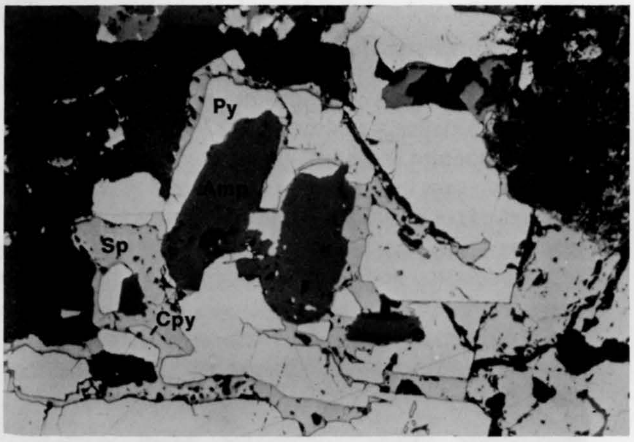

FIGURE 7. Pyrite (Py) replacement of amphibole (Amp) with later replacement of pyrite, amphibole, and other silicates by chalcopyrite (Cpy) and sphalerite (Sp). Wall rock, Little Bob deposit.

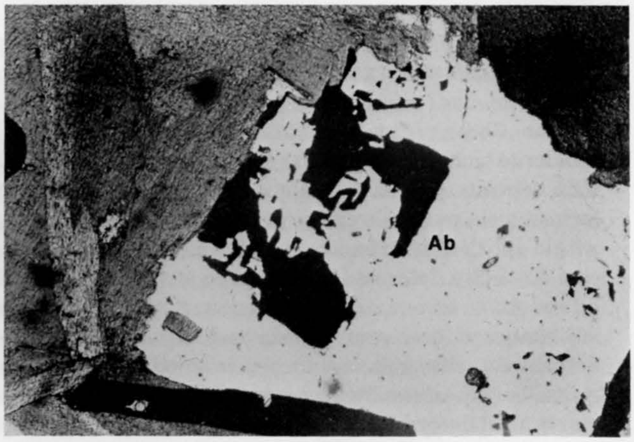

FIGURE 8. Skeletal sphalerite crystal in albite (Ab). Wall rock, Little Bob deposit.

in the Tallapoosa deposit. Individual gold grains average less than 0.01 mm in diameter. The mineral occurs as rounded grains at the contact of chalcopyrite- and tetradymite-bearing galena. It exhibits irregular, mutual contacts with galena and straight contacts with chalcopyrite. The native gold is believed to be a late-stage primary mineral.

Covellite - CuS. Covellite and lesser chalcocite are locally present as supergene enrichment minerals in near-surface ores of the deposits, although no well defined enrichment zones are indicated by drill core analyses. Covellite typically replaces chalcopyrite veins and aggregates in cataclastically deformed pyrite (Fig. 4). Sphalerite and pyrite grains are locally coated and embayed by thin films of covellite. Covellite in the Villa Rica deposit is particularly noticeable where narrow chalcopyrite veins in silicates and pyrite are replaced by covellite at the chalcopyrite-silicate and chalcopyrite-pyrite interfaces. In polished surfaces covellite exhibits extreme pleochroism and a tendency to tarnish rapidly when exposed to air.

Chalcocite - Cu_2S. Although much less common than covellite, chalcocite is ubiquitous in zones containing the former mineral. Chalcocite typically occurs as narrow, apparently replacement networks within larger masses of covellite, particularly in near-surface ores of the Tallapoosa deposit.

Magnetite - $FeFe_2O_4$. Magnetite is a locally important, early accessory mineral in ores of the Little Bob, Swift and

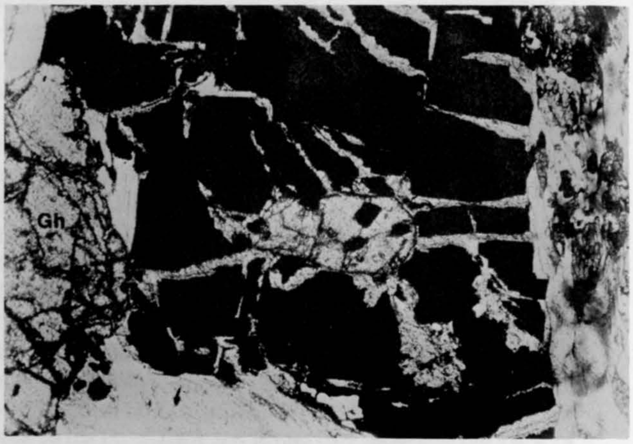

FIGURE 9. Cataclastically deformed magnetite (black) with gahnite (Gh). Wall rock, Little Bob deposit.

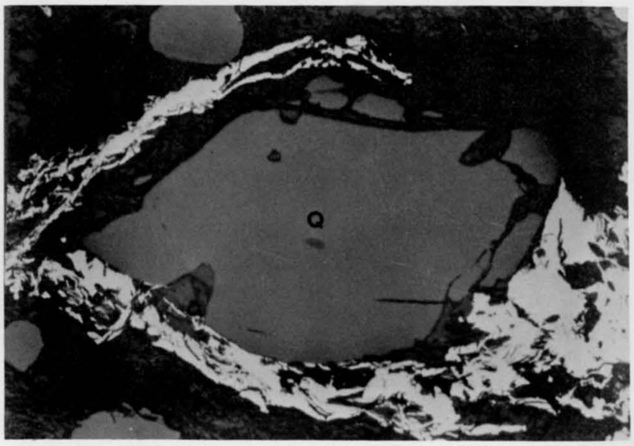

FIGURE 10. Clastic quartz (q) grain bounded by deformed pyrite; wall rock of Villa Rica deposit.

Villa Rica deposits. A second generation of magnetite is indicated texturally in ores of the Little Bob deposit.

Euhedral inclusions of early magnetite up to 2.0 mm in diameter occur in pyrite, sphalerite, and chalcopyrite. Similar individual crystals and aggregates are locally conspicuous in chloritic masses of gangue (Fig. 9). Linsoid masses of angular magnetite fragments occur in a matrix of granular pyrite and schistose pyrrhotite near the hanging wall of the major ore zone in the Little Bob deposit. These masses range in size up to approximately 21 by 13 by 6 mm and are apparently the result of post-ore deformation.

Magnetite is locally abundant at ore zone margins of the Villa Rica deposit where it forms patches of euhedral grains up to 2.8 mm in diameter. Angular magnetite fragments are commonly included within pyrrhotite aggregates of the more massive ore. Sheared zones within the intensely altered wall rocks locally contain streaks of brecciated magnetite in a matrix of quartz and biotite.

Late stage magnetite is common in the massive ores of the Little Bob deposit and is apparently contemporaneous with pyrrhotite deposition. Large irregular masses of magnetite locally embay and vein anhedral pyrite. Pyrrhotite and chalcopyrite are typically intergrown with this magnetite and exhibit mutual boundaries. Inclusions of hornblende and chlorite are corroded, embayed, and veined along cleavage planes by late stage magnetite.

Ilmenite - $FeTiO_3$. Euhedral ilmenite laths are common throughout gangue-rich portions of Little Bob and Swift mine ores. Fine lamellae of exsolved magnetite are conspicuous in most grains. The grains are evenly distributed and well aligned within foliated masses of ferromagnesian minerals. Ragged laths are commonly included within pyrite, pyrrhotite and garnet. The laths are uniform in size and are identical to altered ilmenite in hornblende-rich country rocks and chloritic wall rock. The wide distribution of ilmenite throughout the ore, wall rock, and country rock of these two deposits and its inclusion in pyrite and garnet suggest that it was an original constituent of the host rock and is genetically unrelated to ore deposition.

Gangue Minerals. Gangue of the Little Bob, Swift, and Villa Rica deposits exhibits similar mineralogic and textural characteristics, although local abundance of specific species is quite variable within deposits. The most abundant gangue minerals, including those comprising irregular masses of weakly mineralized, altered amphibolitic wall rock, are quartz, calcite, amphiboles, chlorite, biotite, albite, garnet, and gahnite ($ZnAl_2O_4$).

An initial period of silicification is reflected by angular quartz fragments included within pyrite grains and the local silicification of wall rocks. A later generation of anhedral, medium-grained quartz forms a matrix for granular aggregates of broken pyrite grains. This quartz is glassy and free of inclusions. Shattering is common in zones of deformation. Other sulfide veins, particularly sphalerite, cut second generation quartz.

Calcite occurs locally as an interstitial matrix for brecciated pyrite (Fig. 3). It is occasionally intergrown with quartz and was apparently deposited toward the end of the second period of silicification, but before the introduction of later sulfides. Calcite is typically replaced by sphalerite along cleavage and twin planes.

Irregular masses of untwinned, zoned albite up to 6.0 mm across occur in intensely altered fragments of wall rock consisting principally of chlorite and biotite. Refractive index ($N \alpha$) measurements of albite indicate a compositional range of An_2 to An_7. Selected samples are up to 12% albite. Most albite grains contain well aligned inclusions of chlorite biotite, amphibole, and ilmenite. Skeletal sphalerite crystals occur within or at the margins of larger albite grains (Figure 8).

Chloritic zones within massive ore and mineralized wall

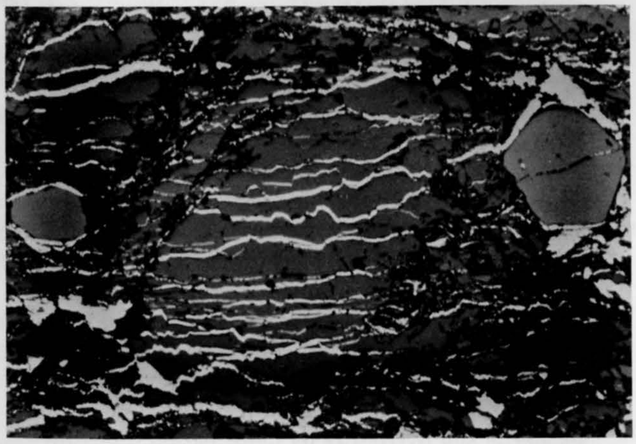

FIGURE 11. Clastic feldspar grain veined by pyrrhotite along cleavage planes; wall rock of Villa Rica deposit.

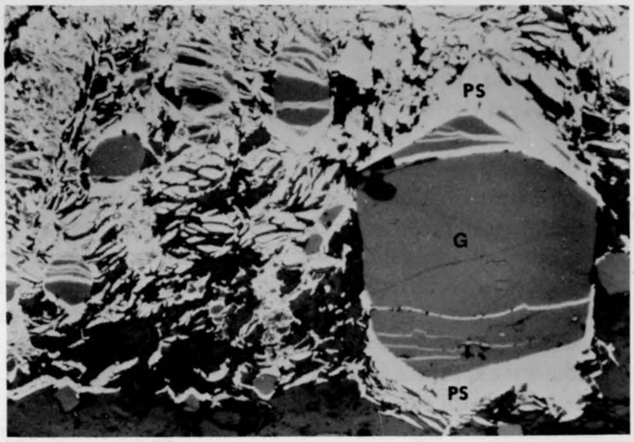

FIGURE 12. Extensive replacement of wall rock silicates by pyrrhotite. Note pressure shadows (PS) bounding euhedral garnet (G); Villa Rica deposit.

rock contain up to 4.4% poikilitic gahnite (Fig. 9). The mineral is blue-green and has a refractive index of 1.80. Semiquantitative spectrographic analyses indicate a Zn content of greater than 10%. Narrow zones within the Little Bob deposit contain euhedral gahnite in a matrix of granular quartz and sericite.

Chlorite exhibiting abnormal brown interference colors and multiple twinning on (001) is abundant in zones of mineralized wall rock and gangue-rich portions of the more massive ore bodies. It replaces hornblende, biotite, and garnet along grain margins and fractures, and occurs as isolated patches and clusters of grains. Refractive indices vary from $N_W = 1.63$ to $N_W = 1.65$.

The composition of ore zone, wall rock, and country rock garnets, as determined by cell edge, density, and refractive index measurements, were found to vary little within individual deposits. All garnets examined from the Little Bob, Swift, and Villa Rica deposits are members of the pyralspite family with a dominant almandine molecule.

The major gangue minerals of the Tallapoosa deposit are chlorite, quartz, and dolomite. Dolomite occurs in medium-grained, equigranular masses which are shattered along planes parallel to foliation. Sulfide mineralization is most intense in these deformed zones. Dolomite is extensively veined and embayed by sphalerite, chalcopyrite, and galena, and locally forms an interstitial matrix for aggregates of cataclastically deformed pyrite. Chlorite, actinolite, and quartz are abundant as inclusions within the sulfides and dolomite, and as major components of larger fragments of weakly mineralized country rock. Foliation is preserved in the larger masses although disorientation with respect to the regional foliation is common. Chlorite of the ore zones shows little optic variation from that of the altered wall rock and country rocks.

Deformation Effects: Ores of the Little Bob, Swift and Villa Rica massive sulfide deposits exhibit textures reflecting at least three stages of deformation temporally similar to those described for Ducktown, Tennessee ores (Carpenter, 1965). The deformational sequence determined for ores of these three deposits are as follows:

1. Pre-pyrite deformation.
2. Post-pyrite, pre-pyrrhotite-chalcopyrite-sphalerite deformation
3. Post-pyrrhotite-chalcopyrite-sphalerite deformation.

The first stage may represent several periods of superimposed cataclastic deformation as evidenced by inclusions of bent and broken silicate grains, particularly amphiboles, chlorite, and biotite, within undeformed pyrite. The second stage is characterized by broken, shattered and sheeted pyrite fragments enclosed within an apparently undeformed, unrecrystallized matrix of one or more of the other dominant sulfide minerals. The third stage of deformation is represented by shattered pyrite, chalcopyrite, and sphalerite, and deformation twinning and possible sub-grain formation in pyrrhotite and sphalerite. Only rarely may pyrite deformed in stage three be distinguished from pyrite deformed in stage two. Pyrite grains veined by later sulfides are locally dissected by fractures which transect both pyrite and the included sulfide veins. Such deformation may be classified as stage three.

The effects of several periods of deformation are preserved in ores of the Tallapoosa deposit. An initial period of post-metamorphic shearing is indicated by inclusions of broken, angular fragments of chlorite, actinolite, and quartz within pyrite crystals. Pyrite and dolomite grains have been shattered along definite cataclastic zones, indicating a second period of shearing after the crystallization of pyrite. The later ore minerals, sphalerite, chalcopyrite, and galena, show only local effects of post-ore shearing. These minerals are commonly fractured with minor dislocations of individual fragments.

CONCLUSIONS

Ore zones of the four deposits consist of lenses of both massive and disseminated sulfide mineralization which are situated within shear zones apparently conformable with the host rocks. Mineralogically, the ores of the Little Bob, Swift, and Villa Rica deposits are similar and of the Ducktown type, consisting of pyrite and pyrrhotite with varying amounts of chalcopyrite, sphalerite, galena, magnetite, and lesser other sulfides. Ore of the Tallapoosa deposit is significantly different from ore of the Little Bob, Swift, and Villa Rica mines due to the absence of pyrrhotite, abundance of gangue dolomite, and relatively low-iron sphalerite.

Ores of the Little Bob, Swift, and Villa Rica deposits appear to reflect a similar paragenesis. It is evident that pyrite was the initial sulfide to crystallize. Pyrite crystallization was followed by a period of cataclastic deformation which generally preceded the deposition of the other ore minerals. Detailed paragenetic interpretation cannot be attempted due to textural modifications produced during apparent post-ore annealing recrystallization and probable deformation sub-grain formation in pyrrhotite and sphalerite. On the other hand, ores of the Tallapoosa deposit appear to have undergone little or no post-depositional reorganization and texturally suggest the following paragenetic sequence: Pyrite→sphalerite→chalcopyrite→galena→tetradymite→gold. The initial period of pyrite deposition in the Tallapoosa deposit was separated from that of the other sulfides by a period of cataclastic deformation.

A complex post-depositional history is indicated texturally in ores of the Little Bob, Swift, and Villa Rica deposits. Pyrrhotite of the more massive ores has undergone at least partial annealing recrystallization, most probably as a normal response to one or more periods of regional metamorphism. Annealing recrystallization of chalcopyrite and galena is suspected but could not be confirmed in the current study. In addition, post-ore deformation has probably resulted in the minor recrystallization of sphalerite and pyrrhotite by deformation sub-grain formation at the ore zone margins and within gangue-rich portions of the ore.

The origin of the ore deposits described herein is open to considerable debate. The four massive sulfide occurrences exhibit unusual similarities which, coupled with their close proximity to one another, suggest a common origin. The intimate association of the Little Bob, Swift, and Villa Rica deposits with known metavolcanic rocks suggests that these occurrences are classical volcanogenic massive sulfide deposits and that many textural features now present can be ascribed to post-depositional shearing and metamorphic episodes, rather than to original deposition. On the other hand, it is difficult to attribute such an origin to the Tallapoosa deposit. This deposit gives every indication of being a simple shear zone replacement, post-dating or perhaps coinciding with the last major metamorphic event. Since the various textures that suggest such an origin for the Tallapoosa deposit are also present in ores of the other three deposits, it is likewise possible to ascribe all four occurrences to shear zone replacement by hydrothermal, sulfide-bearing fluids. Post-depositional recrystallization of Little Bob, Swift, and Villa Rica ores would similarly be attributed to later readjustment during a period of metamorphism that generally diminished in intensity to the northwest, becoming of insufficient magnitude to affect ores of the Tallapoosa occurrence.

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SHEAR ZONES IN THE CORBIN GNEISS OF GEORGIA

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ABSTRACT

The Corbin gneiss, a Precambrian (one billion year) orthogneiss located in the southwestern Blue Ridge of Georgia, contains cataclastic rocks previously mapped as Cambrian-age metasedimentary inclusions. The Corbin gneiss and the overlying, unconformable metasedimentary sequences are present in the southwestern core of the Salem Church anticlinorium. An investigation of the "inclusions" in the gneiss has revealed that some are mylonitized gneiss while others are rocks of variable graphitic content which suggest affinities with the overlying metasedimentary rocks. These metasedimentary inclusions are best explained as rocks that are either enclosed within synclines or that were included early in the history of the gneiss.

INTRODUCTION

This report presents preliminary information on a variety of rocks that occur within the unit which is generally referred to as the Corbin gneiss. Descriptions such as "crushed gneiss" and "metasedimentary inclusions" by previous workers indicate the controversial nature of these rocks. It is hoped that the information presented in this paper will clarify the relationship between these rocks and the adjacent gneiss.

LOCATION

The Corbin gneiss is exposed in Bartow and Cherokee Counties in the Cherokee Upland District of Georgia's Piedmont Province (Clark and Zisa, 1976). The Acworth, Allatoona Dam, South Canton, Waleska, and White East seven and one-half minute topographic quadrangles encompass the outcrop area of the Corbin gneiss (Fig. 1). However, the emphasis of this report will be on rocks exposed within the Allatoona Dam Quadrangle, where lakeshore outcrops provide almost continuous exposure of the Corbin gneiss and associated lithologies.

Good exposures of the Corbin gneiss are also found in roadcuts east of Cartersville along Georgia Highway 20 between Roland Springs and Sutallee.

PREVIOUS WORK

The Corbin gneiss was first described by Hayes (1897) who mapped the rock exposed near Corbin in Bartow County as a granite. Later work by Watson (1902) on the granites and gneisses of Georgia presents the first description of the effects of dynamic metamorphism on the Corbin gneiss. He described the rock as a granite that was altered to an augen gneiss by the effects of dynamothermal metamorphism. In addition, Watson mentioned that in places the metamorphism is of such intensity that the

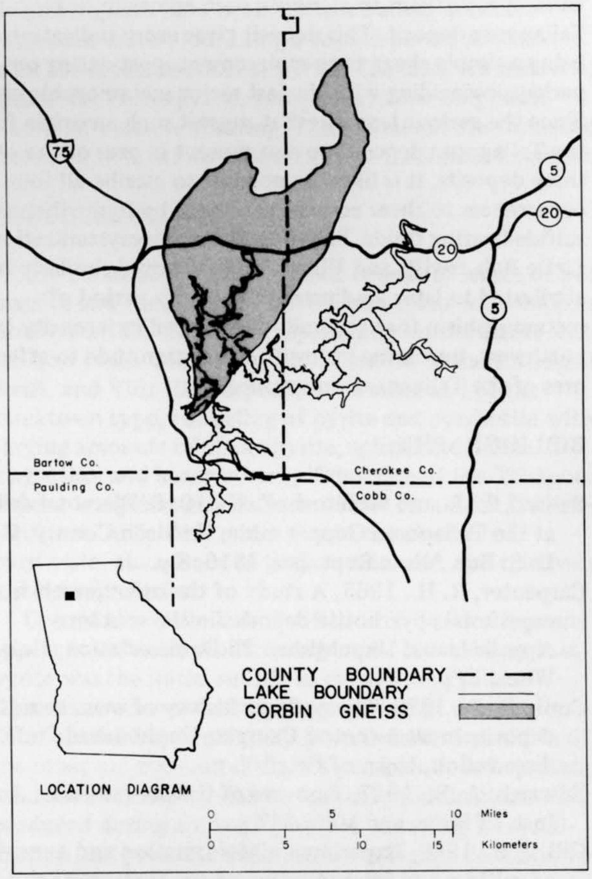

FIGURE 1. **Generalized areal extent of the Corbin gneiss.**

constituent minerals are "completely differentiated into parallel, alternating thin bands."

LaForge (Hull, et. al., 1919, P. 40) reporting on the Corbin gneiss, described the rock as ". . .typically a coarsely porphyritic augite-biotite granite, most of which has been so much sheared that it is now characteristically an augen gneiss." He adds, "In places it displays a fine-grained, non-porphyritic phase which has suffered less from crushing than the porphyritic rock and in other places it has been so greatly crushed that it is altered into biotite-sericite schist without a semblance of the original rock." It is interesting to note that observations made by both Hayes and LaForge allowed them to postulate that the Corbin gneiss was basement rock that furnished much of the sediment which now constitutes the surrounding metasedimentary rocks.

Outcrops exposed on the Etowah River four miles east of Cartersville (now beneath Lake Allatoona) led Crickmay (1936) to conclude that the Corbin Granite intruded the surrounding rocks. In addition, he reported that the granite had been locally micro-granulated yielding textures which range from coarse-grained augen gneiss to mylonite which

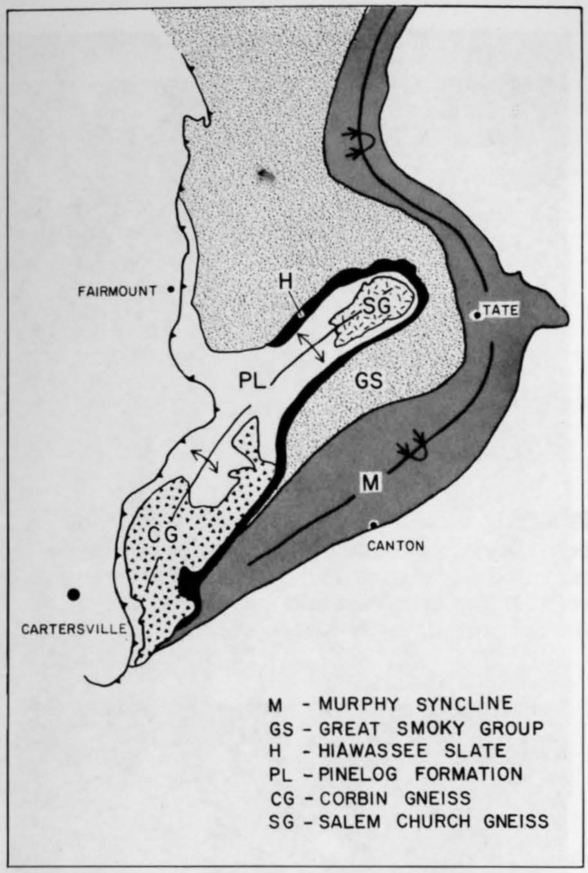

FIGURE 2. Generalized geologic map of the Salem Church Anticlinorium and Murphy Syncline

resembles fine-grained arkose.

Kesler (1950) mapped the Cartersville Mining District and defined an association of lithologies in the area of the Corbin gneiss that he interpreted to be caused by the alteration of Cambrian age sediments by igneous influence during late Carboniferous time. Kesler's lithologies include three gneisses: an oligoclase-mica gneiss, an andesine-augite gneiss, and a porphyroblastic gneiss -- the most abundant of the three, which he characterized by the presence of large orthoclase porphyroblasts. He further indicates that rocks of the Rome and Weisner Formations are included in the porphyroblastic gneiss.

Croft (1963), in a report on the geology and ground-water resources of Bartow County mapped a porphyritic granite gneiss. He concluded, based on the trace of the contact, that the Weisner Formation was intruded by the granite gneiss.

Morgan (1966) produced a model analysis of a sample of the Corbin gneiss taken at Cooper Branch on Lake Allatoona from which he concluded that the bulk chemical composition was far removed from the thermal minimum composition of a granite. He also indicated the presence of abundant rounded zircons. However, he did not state unequivocally that the Corbin gneiss originated from either sedimentary or igneous rock.

Uranium-lead analyses by Odom and others (1973) of zircons from the Corbin gneiss yield lead-lead ages in excess of one billion years. The authors further state that the Corbin gneiss does not intrude the Ocoee Series rocks but represents part of the pre-Ocoee basement.

Martin (1974) has described the Corbin gneiss as containing large light-grey microcline crystals up to 10 cm. in diameter, in a medium-grained ground mass consisting of blue quartz, plagioclase, biotite, garnet, and ilmenite, accessory zircon, sphene, and other minerals.

He concluded that the Corbin gneiss was emplaced as a magma and that it underwent differentiation as it cooled. Subsequently, it was subjected to two periods of metamorphism -- the former a granulite facies event and the latter of greenschist facies. He further concluded that following each of the metamorphic phases there was a period of uplift and erosion, the second of which has produced the present topographic surface.

REGIONAL GEOLOGY

The Corbin gneiss is exposed in the core of the Salem Church anticlinorium (Fig. 2). In contact with the gneiss are a variety of clastic metasedimentary rocks including conglomerates, quartzites, sericitic phyllites and graphitic phyllites.

The conglomerates, quartzites, and sericitic phyllites have been mapped as the Pinelog Formation by LaForge (Hull, et.al., 1919) and as the Weisner Formation by Kesler (1950). The more coarse-grained rocks of the sequence are, in places, feldspathic and locally contain layers of blue quartz clasts. In addition, the conglomerates and quartzites are responsible for much of the rugged topography of the Salem Church anticlinorium due to their relative resistance to weathering.

The graphitic phyllite is best exposed on the eastern margin of the gneiss. Fresh exposures of the graphitic phyllite are uncommon but the rock weathers to a distinctive black saprolite and soil which makes the lithology easily traceable. When seen in place, the rock is frequently interbedded with layers of blue quartz bearing meta-arkose. The graphitic phyllite trends northeastward around the nose of the Salem Church anticlinorium where Bayley (1928) mapped it is as the Hiawassee Slate. Less extensive exposures of graphitic phyllite have been located in areas well within the borders of the gneiss.

Structurally above the Hiawassee Slate is the Great Smoky Group which, in the area of study, is characteristically a uniform sequence of interbedded metagraywacke and phyllite.

The Salem Church anticlinorium is bounded on the west by the Great Smoky Fault and on the north and east by the rocks of the Murphy Syncline.

METHOD OF INVESTIGATION

Using Kesler's (1950) geologic map of the Cartersville Mining District, lithologic data were transferred to the 1968 1:24,000 scale Allatoona Dam Quadrangle. A reconnaissance sampling traverse of several of the specific exposures designated by Kesler as metasedimentary Rome and Weisner lithologies was undertaken. The sample localities are shown in Figure 3.

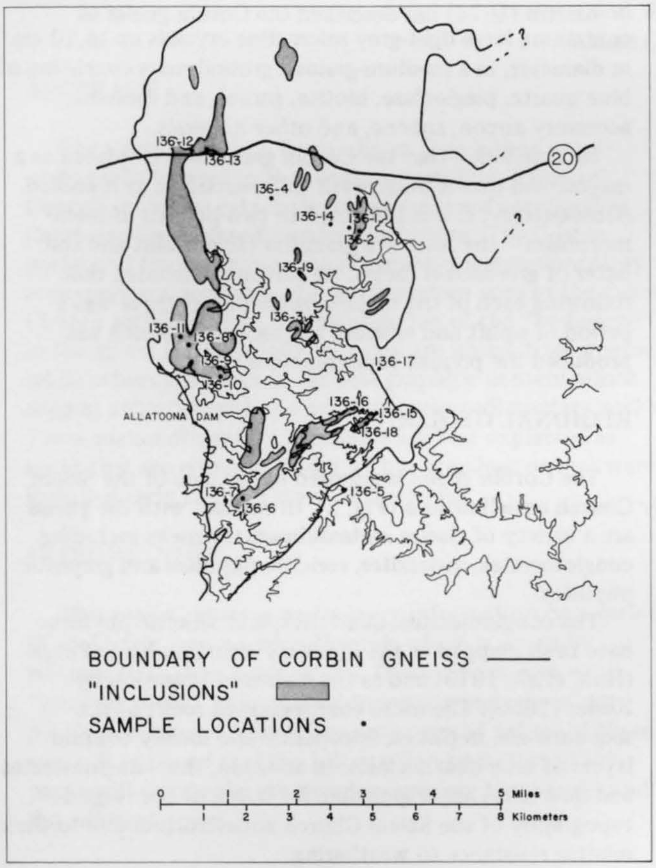

FIGURE 3. **Allatoona Dam Quadrangle showing the Corbin gneiss "inclusions" and sample locations (after Kesler, 1950)**

Contact and lithologic relationships were observed in outcrop. In addition, slabs and thin-sections were prepared from the collected samples. Portions of each sample were submitted for chemical analysis and the results of these analyses will appear in forthcoming reports.

GENERAL DESCRIPTION OF THE INCLUSIONS

An examination of the rocks within the areas delineated on Kesler's (1950) geologic map as metamorphosed Rome and Weisner Formations yields, upon casual observation, suites of rocks that are markedly different in texture from the adjacent porphyroblastic augen gneiss. In fact, the rocks appear to be conglomerates, quartzites, schists, and phyllites.

However, shear stresses have acted on the "inclusions" and the Corbin gneiss that surrounds them. The result has been that the rocks in most instances reveal cataclastic textures. In some cases, fluxion structure is visible to the unaided eye and even where flow is not readily apparent, the visible mineral constituents of the rocks are generally crushed and enclosed within pressure shadows (Plates I and II). In thin-section, the rocks display a variety of typical cataclastic textures (Plates III and IV).

There are several areas within the Corbin gneiss, particularly along the shore of Lake Allatoona, where unusually good exposure affords the opportunity to study outcrops where the augen gneiss has undergone a complete textural transformation as a result of shearing stresses. Careful inspection of these excellent exposures shows the increasing effects of shearing from the border of the shear zone to the central portion. Along the fringe or border sections of deformation the most easily observed effect of shearing is the crushing and stretching of the microcline porphyroblasts. The effect of the shearing increases until the rock is dominantly phyllonitic with random porphyroclasts of feldspar or quartz scattered through the fine-grained ground mass. Occasionally, stringers or layers of more vitreous mylonitic rock extend through the phyllonite

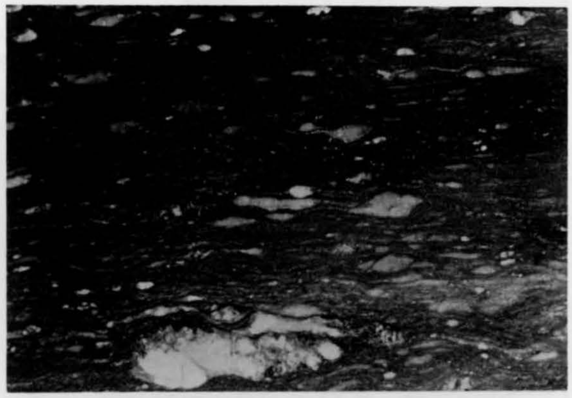

PLATE I. **Slab of sheared Corbin gneiss showing porphyroclasts of quartz (q) and microcline (m) couched in a cataclastic matrix. This rock was sampled from within one of the areas defined as an inclusion of Rome and Weisner Formations by Kesler (1950).**

PLATE II. **Outcrop of sheared Corbin gneiss showing the crushing effect of shearing on microcline (m) porphyroclasts. A quarter is used for scale. The exposure is located in an area mapped as Rome Formation by Kesler (1950).**

PLATE III. Photomicrograph of sheared Corbin gneiss showing two quartz porphyroclasts with "crush" recrystallization trails and recrystallized grain boundaries.

PLATE IV. Photomicrograph of sheared Corbin gneiss showing grain boundary recrystallization, pressure, shadows and crushing of porphyroclasts and microcrenulation of sericite (s).

(Plate V). Within this area, the effects of shearing are much like those described by LaForge (Hull, et. al. 1919). However, such dramatic exposures as those described above are rather rare.

Several localities were checked where the rocks present were definitely metasedimentary. The rocks in these areas ranged from meta-arkosic conglomerates to graphitic phyllites.

DISCUSSION AND CONCLUSIONS

The controversy over the origin of the rocks included within the Corbin gneiss cannot be solved by generalizing that they are all sheared gneiss. Because some of the inclusions are graphitic phyllite and other possible metasedimentary rocks, such a generalization is not valid. Work presently underway to delineate the sheared gneiss units and the sheared meta-arkosic rocks indicates that much of the rock that was mapped by Kesler (1950) as metasedimentary inclusions in the porphyroblastic gneiss is, in fact, sheared gneiss. Kesler's (1950, p. 38) description of the contact relationships between the porphyroblastic gneiss and the inclusions is given without any expressed regard to the cataclastic textures that are present. What Kesler describes as "secondary feldspar and quartz" in the "gneissoid" metashales near the gneiss-metashale contact is a description of what is, in fact, protomylonitic gneiss.

PLATE V. Outcrop of Corbin gneiss on Shut-in Creek showing the effects of cataclasis on grain size. Note the stretching of the microcline augen approaching the dark zones of the most intense deformation. A penny is used for scale.

That the Corbin gneiss is of igneous origin is quite possible, but not in the sense that is proposed by Kesler (1950). Kesler's interpretation of the origin of the gneisses was that they were derived in late-Carboniferous time by the static igneous alteration of Weisner and Rome Formation sedimentary rocks of Cambrian age. Igneous relationships have been seen by Martin (1974, p. 104) along the shore of Lake Allatoona where he reports xenoliths of basic rock in outcrops of Corbin gneiss. In addition, inclusions of fine grained gneissic rocks in the Corbin gneiss in Red Top Mountain State Park have been found (Gillon, personal commun., 1977 and Costello, unpublished data)(Plate VI). However, Martin (1974, p. 104-105) concludes the Corbin gneiss-metasediment contact is an unconformity. This is substantiated by field observations made in the Waleska and White East Quadrangles to the north where Corbin gneiss can be seen in direct contact with arkosic conglomerates (Plate VII). Therefore, as an alternative to Kesler's interpretation, it is proposed that the clastic metasedimentary rocks occupy the cores of synclines (Plate VIII) which resulted from the infolding of the unconformity.

It is evident that more detailed lithologic and structural mapping of the Corbin gneiss and associated lithologies is much needed to further the understanding of the role played by the zones of shearing in this complex area of the Georgia Piedmont.

PLATE VI. Igneous contact between megacrystic Corbin gneiss (below) and a finer grained gneissic rock (above).

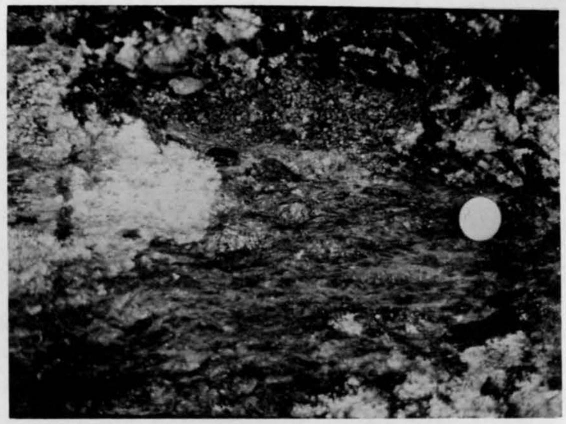

PLATE VII. Unconformable contact between porphyroclastic Corbin gneiss (top) and arkosic conglomerates of the Pinelog Formation. A dime is used for scale.

PLATE VIII. Syncline with a core of meta-arkosic rocks enclosed with Corbin gneiss.

ACKNOWLEDGEMENT

Grateful recognition is given to my colleagues at the Georgia Geological Survey for assisting me during preparation of this paper.

Thanks are also given to Michael Higgins of the United States Geological Survey who visited the study area with the writer and to Ken Gillon who provided valuable insights into certain aspects of the Corbin gneiss.

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THE GEOLOGY AND GROUND WATER OF THE GULF TROUGH

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ABSTRACT

The Gulf Trough is a subsurface geologic feature that affected the deposition of sediments from probably as early as upper Eocene, Jackson time through Miocene time. The trough can be found parallel to a hydrologic anomaly trending N53°E. from northern Thomas, Colquitt and southern Tift Counties and is hypothesized to extend into northern Effingham County. The Gulf Trough is distinguished from an earlier defined feature, the Suwannee Strait, by location, stratigraphic position, lithologic character and structure.

Ground-water availability is adversely affected by the presence of the Gulf Trough. Yields of 50 gallons per minute are typical, whereas in areas outside the trough's influence, 500-1000 gallons per minute are common. Lithologic parameters of the aquifer affect the overall porosity and it is postulated that they are responsible for the reduced yields. Samples taken from within the axial zone of the trough show a finer microcrystalline texture with fewer bioclasts than samples from the aquifer outside the trough. The fine texture can inhibit ground-water flow. Geophysical well logs provide further evidence of low porosity. Other possibilities such as reduced aquifer thickness and multi-aquifer wells may be considered partly responsible for the reduced yields. Faulting has been considered by other investigators as the cause for the reduced yields.

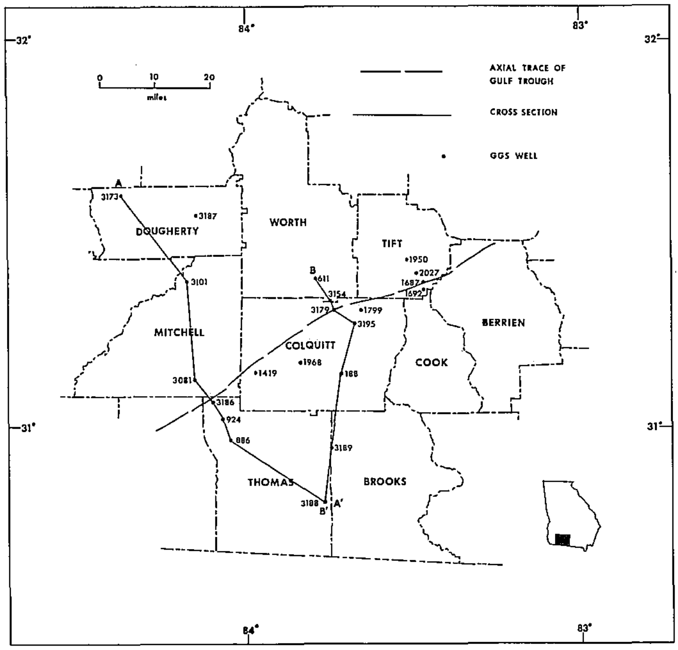


FIGURE 1 **Location map of GGS wells used for this report. Location of cross sections A-A′ and B-B′. Location of axial trace of the Gulf Trough.**

INTRODUCTION

The subsurface geologic feature known as the Gulf Trough is a relatively recent discovery. The name "Gulf Trough of Georgia" was first used by Herrick and Vorhis in 1963. (Hendry and Sproul, 1966, dropped the "of Georgia" and used "Gulf Trough" so that the feature wouldn't be restrictive to Georgia). The Gulf Trough is believed to be an extension of the Apalachicola Embayment located in the southwestern portion of Georgia and in the panhandle of Florida. The Apalachicola Embayment is recognized as an area of Late Mesozoic (Jurassic and Cretaceous) and Cenozoic sediment thickening. The greatest thickening occurs in the southwestern portion of the embayment; here the Tertiary is about 5,000 feet (1525 m) thick. The sediments thin in a northeasterly direction as the embayment narrows (Murray 1961). The trough extends in a northeasterly direction from the narrow end of the embayment into the Coastal Plain of Georgia (Fig. 1) and is postulated as being an area of concomitant Tertiary (possibly upper Eocene, Oligocene, Miocene and Pliocene) sediment thickening.

A geologic feature called the Suwannee Strait[1] was recognized by Coastal Plain investigators long before the Gulf Trough was first described. The Suwannee Strait was first described in Dall and Harris (1892) as a water passage extending from Savannah, Georgia to Tallahassee, Florida.

Since that time many investigators have delineated the subareal extent and geologic time span, speculated upon the cause and function and have given different names to the feature first called the Suwannee Strait. Some investigators have tried to show that the Suwannee Strait of Dall and others, and the Gulf Trough of Herrick and Vorhis and of this report are the same feature. After considering some of the data and descriptions of previous investigators (summarized below) as well as additional data acquired from the present study, the author believes that they are two independent subsurface geologic features with different geometries and geologic time spans. It is also believed that the geologic origin and function of these two features may indeed have been very different. The effect that the Gulf Trough exerts upon the present ground-water conditions of the principal artesian aquifer will be shown to be unique, while the Suwannee Strait has no apparent effect upon the ground-water conditions. The

[1] Original 1892 spelling was with one "n". Subsequent authors used two "n's" which is considered the correct spelling.

cause and effect relationship of the subsurface geology and the resulting ground-water behavior are responsible for the interest shown in this study of the Gulf Trough.

HISTORICAL BACKGROUND

The pertinent literature regarding the description of the Gulf Trough and its distinction from the Suwannee Strait is discussed below. The confusion of these features in recent literature has resulted from misinterpretation of the early literature and from postulates of origin and function based on insufficient data.

Dall and Harris (1892, pg. 121) proposed the name Suwannee Strait for a "passage between Florida and the mainland..." that "in Miocene time was a moderately deep body of water, the general trend of which did not differ much from that of a line drawn from Savannah to Tallahassee and which had a probable width of more than 50 miles." They state that this area is now occupied by the Okefenokee and Suwannee Swamps, and the trough of the Suwannee River, and is composed of deposits of clays and silicious material of Miocene age.

Applin and Applin (1964, pg. 1727) discussed a "major structural feature" that became evident from their stratigraphic study, but failed to completely identify the feature. They noted that a "channel or trough extended southwestward across Georgia through the Tallahassee area of Florida to the Gulf of Mexico". They point out that on the southeast side of the feature, from the top of the Oligocene down through the base of the late middle Eocene, the carbonate rocks from a continuous limestone sequence throughout southern Georgia into peninsular Florida. On the northwest side of the feature, from the middle Eocene down through the base of the Upper Cretaceous clastics predominate in southern Georgia and the Florida panhandle. The carbonates are laterally gradational within the "channel or trough" with the clastics, and are thinner in the gradational zone than they are on either side. They go on to state that with the passage of time the limestones of the peninsula overlapped western Florida and southern Georgia. They imply that the "structural feature" behaved as a natural barrier to carbonate sedimentation north of the channel beginning in the Upper Cretaceous and ceased to function after early middle Eocene time.

Jordan (1954) had a slightly different approach to this feature. She suggested that the Suwannee Strait is an erosional feature resulting from regional movement at the end of Cretaceous time, causing a channel to be cut between clastic rocks in Georgia and carbonate rocks in Florida. She stated that the location of the Suwannee Strait is between the Peninsular and the Chattahoochee "uplifts", but specifically drew the feature on a map extending from Tallahassee, Florida, through Lowndes County and the Okefenokee Swamp, Georgia.

Herrick and Vorhis (1963) proposed the term "Gulf Trough of Georgia" for a northeastern trending belt of anomolously thick sediments of Miocene age in northern Thomas, Colquitt and southeastern Tift Counties. These same investigators noted, in an unpublished report (1973), that the area of greatest thickening occurred along the same trend as an anomalous gradient phenomenon recorded on the potentiometric map of the principal artesian aquifer (Fig. 2). On the basis of this corroborating information the investigators extended the possible subareal extent of the Gulf Trough through Bulloch and Screven Counties. They lacked deep subsurface well data to further substantiate the coincidental trend.

Rainwater (1956) suggested that the Suwannee Strait existed along a line trending eastward from Jackson County, Florida, into Georgia. He placed the Suwannee Strait in the same geographic area as the Apalachicola Embayment and the Gulf Trough (as later defined). From his description of the strait it can be concluded that he borrowed the origin and function of the feature from the explanations of Applin and Applin (1944) and Jordan (1954), but shifted its position westward approximately forty miles (64 km). He used no geologic data to support this move. It seems evident that the confusion concerning the Gulf Trough and Suwannee Strait began here.

Chen (1965) further perpetuated the confusion when he showed a series of paleogeographic maps of southern Georgia and Florida and tried to demonstrate that the Suwannee Strait existed from at least Midwayan time through Jacksonian time, (early Paleocene through late Eocene), and migrated, during this time period to the northwest. In other words, it started out in the area delineated by Applin and Applin, and Jordan and it ended up where the Gulf Trough is now believed to exist. Chen supported the idea that the strait continued to act as a barrier and waterway separating carbonate and clastic depositional environments through late Eocene time. The barrier was previously postulated by Applin and Applin and Jordan only for sediments through middle Eocene time. Unfortunately, Chen used few data points to support his theory.

To emphasize the premise that the Suwannee Strait and the Gulf Trough are two separate features, the major conceptual differences noted prior to the present study are summarized below.

The Gulf Trough as described by Herrick and Vorhis (1963) existed in sediments of Miocene time. They noted a marked thickening of Miocene strata. They also implied an apparent thickening of strata deposited after middle Eocene time. The existence of the trough if surther supported by the potentiometric map of the principal artesian aquifer, especially along the line of Miocene sediment thickening.

The Suwannee Strait was recognized and defined by Applin and Applin (1944) and Jordan (1954) as being active from Late Cretaceous through middle Eocene time. It was presumed to have acted as a barrier to sedimentation, separating carbonates on the south side of the feature from terrigenous clastics on the north side. The well cuttings they used for their studies tend to support their hypothesis of a facies change across the zone where the axis of the Suwannee Strait is located. Unfortunately, the subsurface data relied upon by these investigators was inadequate for a thorough determination of the geometry, structure, and origin of the strait, therefore much liberty was taken by these early investigators regarding the nature of the feature.

The Suwannee Strait as defined by Dall and Harris (1892), Applin and Applin (1944) and Jordan (1954) was located

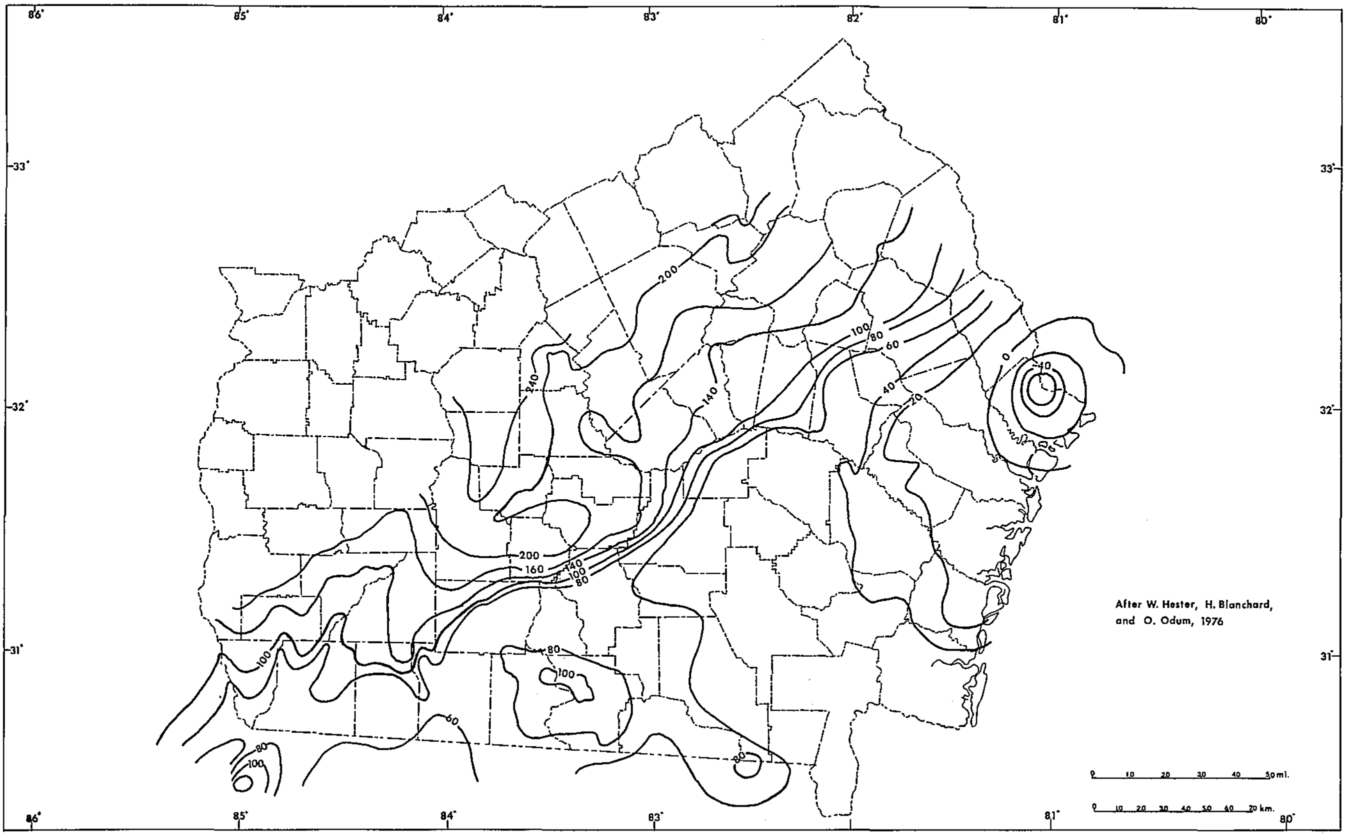


FIGURE 2 **Potentiometric map of the principal artesian aquifer, January-May 1976; simplified after Hester, Blanchard, and Odum, U.S.G.S.**

approximately forty miles (64 km) southeast of the position of the Gulf Trough as defined by Herrick and Vorhis (1963) and by this author. In addition, the Suwannee Strait is not known to be associated with any phenomenon affecting the hydraulic characteristics of the principal artesian aquifer.

ANALYTICAL PROCEDURES

A total of 21 wells were selected for this study. Their locations are shown on Figure 1 and they are listed on Table 1. Among these wells 15 are water wells, five GGS wells are cores, four with geophysical well logs. GGS #3154 is an oil test well with geophysical well logs. Almost all of these wells have complete sets of core or cuttings.

Samples were obtained from the sample library of the Georgia Geologic and Water Resources Division of the Department of Natural Resources. They were examined with a binocular microscope and in hand specimen where applicable. Samples were studied from the surface downward. In well cuttings and in core, formation boundaries were most often picked by the first appearance of diagnostic foraminifera or other fossil. Geophysical logs were used to help determine boundaries and relative porosities. Lithology was not generally considered a good criterion for determining the formation boundary between the Suwannee and Ocala Limestones.

The Suwannee Limestone is a bioclastic limestone with occasional dolomite layers. It is free of quartz sand, major clay layers and phosphate pellets. Its relative monomineralogy lithologically distinguishes the unit from the overlying Miocene strata which consists of major sand, clay and common dolomite but sparse limestone layers with occasional fossiliferous lenses and phosphate rich horizons. The Suwannee's textural and mineralogical similarity to the Ocala make it necessary to separate the two limestones by the first appearance of diagnostic foraminifera, however, in some core, lithologic differences were prominent and were used as well. *Paraotalia mexicana*, many species of *Lepidocyclina*, and/or *Dictyoconus sp.* distinguish the Suwannee Limestone from other formations. *Lepidocyclina ocalana*, *Nummulities (Operculina) marianensis*, *N. (Operculina) floridensis* or *N. (Operculina) ocalana*, *Pseudofragmina sp.* and *Asterocyclina sp.* are representative Ocala Limestone foraminifera.

The middle Eocene Lisbon Formation can be distinguished from the overlying Ocala Limestone by lithology as well as diagnostic fossils. The Lisbon Formation is commonly a fine grained, crystalline limestone that is arenaceous, argillaceous, glauconitic and micaceous.

The upper Paleocene formations, the lower Paleocene Clayton Limestone and the Cretaceous age formations were encountered in four wells. The units are disconformable and are distinguished by their lithologies.

Cross sections A-A′ (Fig. 3) and B-B′ (Fig. 4) are located on Figure 1 and were constructed to show the subsurface in a simplified manner. The fence diagram (Fig. 5) shows the relative thickness and subareal relationships of the strata in

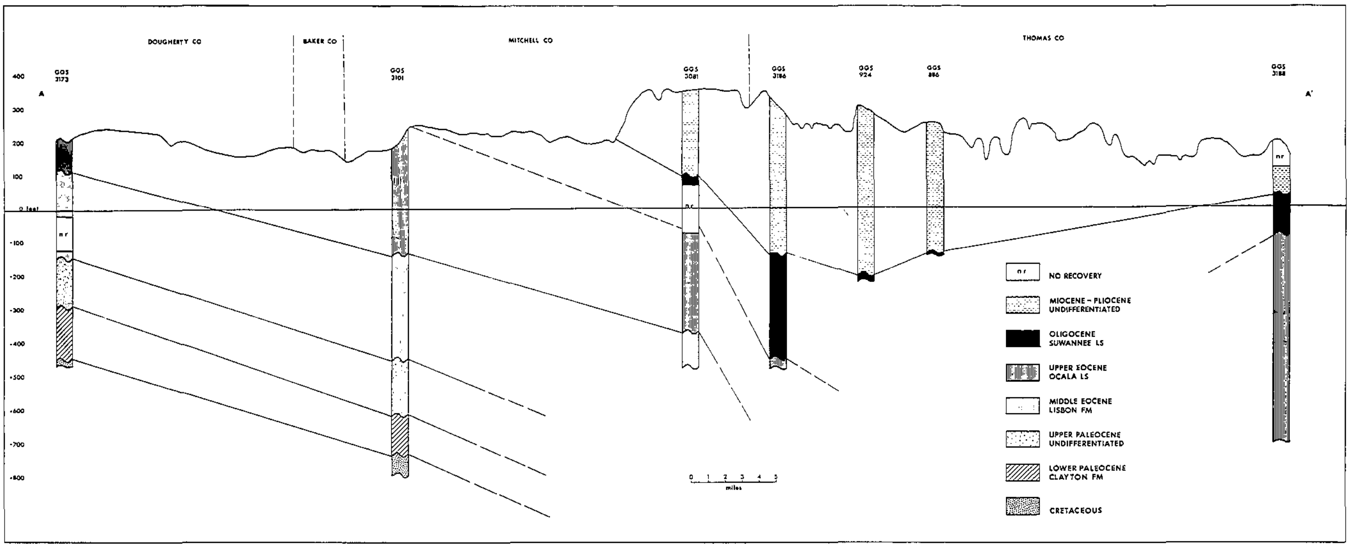


FIGURE 3 **Cross section A-A′ across trend of the Gulf Trough, through Dougherty, Mitchell and Thomas Counties.**

the study area. Solid unconformity lines are drawn where the the contacts were picked by diagnostic foraminifera, or by a prominant lithology change. Dashed lines are drawn where a contact is suspected because of the presence of poorly preserved fauna.

DESCRIPTION OF THE GULF TROUGH

In the present study the portion of the Gulf Trough that will be emphasized is that segment restricted to northern Thomas, Colquitt and southeastern Tift Counties. This portion of the Gulf Trough might be representative of the trough farther to the northeast, further study is in progress to extend the subsurface control in that direction.

Features of the Gulf Trough relating to ground water.

The existence and location of the trough is based partly on the potentiometric map pattern of the principal artesian aquifer (Fig. 2). The proposed extension of the trough into the northeastern section of the State is based solely on this map. In the central section of the Coastal Plain there is a series of closely spaced potentiometric contour lines representing an area of anomalous hydrologic characteristics in which there is a marked increase in gradient in the direction of flow. For example, north of the trough, in the segment from Tifton to Omega the gradient is 13 feet per mile (2.5 m per km) and measured from Omega to a point on a contour 2.5 miles (4 km) south is 28 feet per mile (5.3 m per km). In this zone of increased gradient, water wells have low yields compared to wells drilled outside of the trough. The Gulf Trough is hypothesized to have existed, and to continue to influence the geohydrology of the area located within this zone of closely spaced contours.

There are two characteristics of the trough evident on the map. First, the trough is a linear feature which appears to extend across the State. Although the trend of the axis of the trough has not been completely defined by geological methods, the anomalous potentiometric surface has a strike of N 53°E. The anomaly varies less than 10 to 12 miles (16 to 19.2 km) from a straight line extending from the southwestern corner of Grady County to the northern Effingham County line.

The second characteristic is that the feature is quite narrow. The trough widens where it joins the Apalachicola Embayment but in the central portion of the Georgia Coastal Plain it is only 8 to 14 miles (12.8 to 22.4 km) wide.

Features of the Gulf Trough relating to geology.

The zone of axis of the Gulf Trough is considered to lie within the hydraulic anomaly described above, with the axial trace approximating the mid-line of the anomaly. In the following paragraphs the subareal extent of the trough will be discussed using the data collected for this study.

An oil test well, GGS #3154, drilled near the Worth County - Colquitt County border penetrated 460 feet (140.3 m) of Miocene sediments before the Suwannee Limestone was reached. The Georgia Geologic and Water Resources Division drilled a core hole, GGS #3179, one mile (1.6 km) southeast of the oil test well to a depth of 705 feet (215 m) below land surface without reaching the Oligocene surface. Approximately five miles (8 km) to the southeast of the oil test well, or two miles (3.2 km) southeast of the city of Norman Park, a city water supply test well, GGS #3195, penetrated the Suwannee Limestone at a depth of 470 feet (143.4 m). The Ocala Limestone and Lisbon Formation were probably penetrated, but the fossil fauna was sparse and poorly preserved so that the boundaries are speculative. The three wells GGS #3154, 3179 and 3195 can be seen on the cross section B-B′ Fig. 4 and on the fence diagram Figure 5.

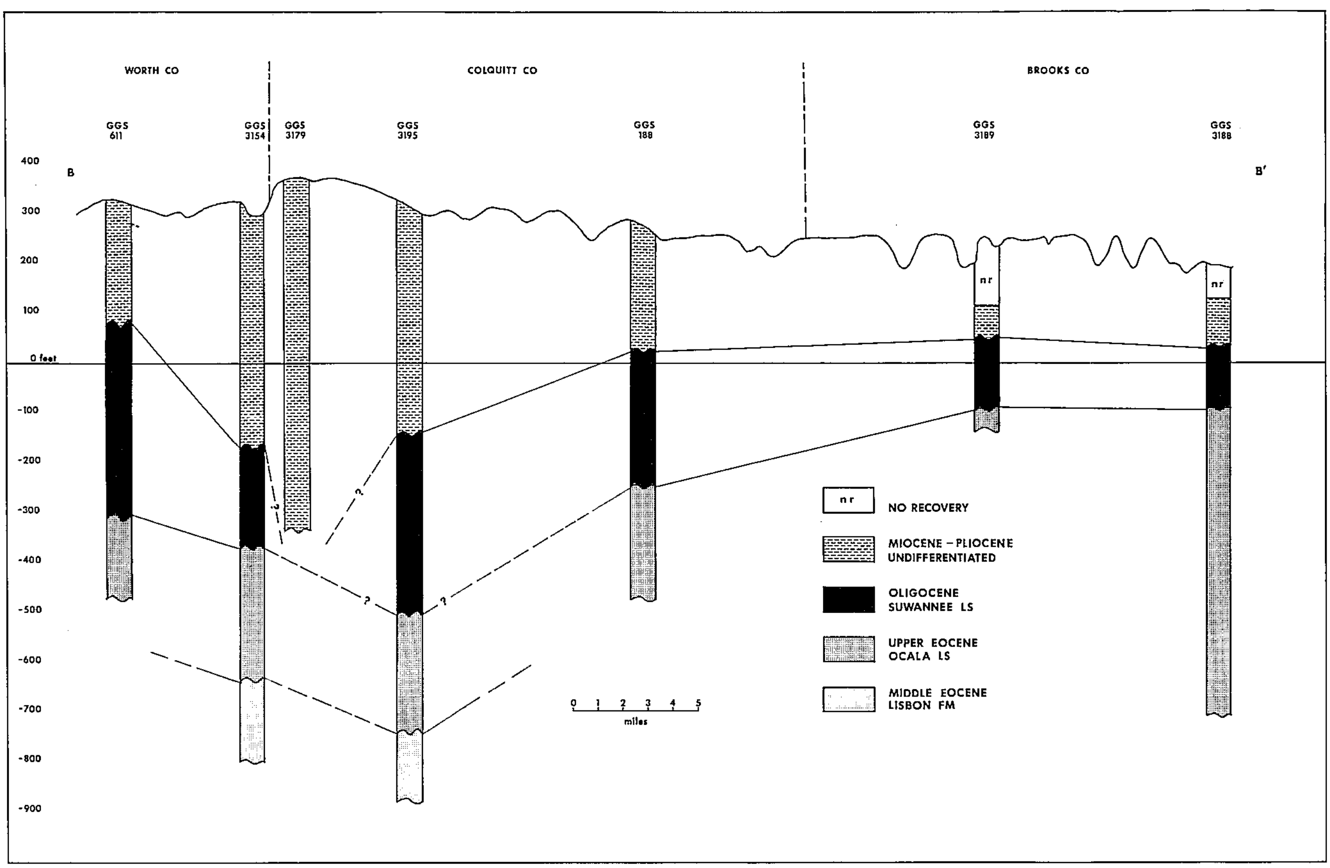
It it is assumed that the geologic data from these wells represent the typically disconformable Miocene - Oligocene boundary relationship and have complete formations, that GGS #3195 lies on the northwest flank of the trough, GGS #3179 was drilled in the zone of axis, and that GGS #3195 lies on the southeast flank, it can be concluded that within a distance of approximately 7 miles (11.2 km) the minimum subsurface dip on top of the Oligocene changes from approximately 90 feet per mile (17.2 m per km) to the southeast to approximately 40 feet per mile (7.6 m per km) to the northwest. This subsurface structure may indicate a local depocenter where Oligocene deposits are complete but thin due to a slow deposition rate, or are complete but thick and found deeper in the sediment basin. However, GGS #3179 does not reach the Oligocene surface, therefore the thickness of the Miocene and Oligocene is still unknown for this area of the trough. The possibility exists that the Miocene sediments filled in a locally eroded portion of the Oligocene, producing a local angular unconformity. If the attitudes were calculated on the complete Oligocene unit the dips would be less steep.

Several wells that lie in the trough have thick Miocene deposits. In southeastern Tift County GGS #1962 was drilled to a depth of 900 feet (274.5 m) and does not the decreased transmissivity such as multi-aqufer wells, structure and lithology, and any of them may be interrelated. These causes are discussed below.

Most wells drilled within the axis of the Gulf Trough never penetrate the principal artesian aquifer, however, a few very deep wells penetrate the top 20 to 50 feet (6.1 to 15.3 m). These deep wells have hundreds of feet of open hole above the Suwannee Limestone and the total yields are affected by the uncased portions of the overlying Miocene strata. The Miocene strata vary laterally in permeability and are not known to be a productive aquifer at any locality in the study area. The overall effect is a reduced yield for these multi-aquifer wells. In other words, only a small portion of the principal artesian aquifer is tapped, and its percentage contribution to the combined Miocene and Oligocene aquifer may actually be large. Some of the wells from northern Colquitt and southeastern Tift Counties fall into this category.

A second possible explanation of reduced transmissivity is a reduced thickness of Oligocene strata. In the northern part of Colquitt and southeastern Tift Counties there is a possibility of a thinner aquifer which may be due to erosion, nondeposition, or slow rate of deposition. A reduction in aquifer thickness, regardless of the cause may lead to both the anomalous thickness of the overlying Miocene strata and a concomitant thinness of Oligocene limestone resulting in the development of low yielding wells in a multi-aquifer. However, the low yields that occur in this area are similar to yields from wells found southwestward along the trough axis where the Miocene is about half as thick and the Oligocene seems to be a complete unit; therefore, there is some additional factor responsible for the low yields found in the zone of axis of the Gulf Trough.

FIGURE 4 Cross section B-B′ across trend of the Gulf Trough, through Worth, Colquitt, Brooks and Thomas Counties.



Faulting in a direction parallel to the trough axis could result in low permeability barriers and may reduce the thickness of the aquifer along the fault. This possibility has been proposed by several investigators (for example, see Sever 1966), however, their hypotheses will not be discussed here because of the lack of supporting geological data.

The last possibility is a change in lithology within the aquifer in the trough. The low yields from wells such as GGS #3195 and #3186 indicate that there is a possible lithologic control influencing the amount of water supplied to these wells from the principal artesian aquifer. Domenico (1972, pg. 168) states that when an increase in gradient occurs in the direction of flow across potentiometric contour lines, reduced permeability within the aquifer is indicated. Textural and mineralogical changes can drastically alter the permeability of an aquifer. Cemented, fine grained limestones transmit less water at a slower rate than partially cemented, bioclastic granular limestones. Saccharoidal dolomites have different permeabilities than fine grained dolomites and limestones. Murray (1960) shows that in dolomite sequences the early stages of dolomite growth are accompanied by decreased porosity and reduced pore size. Porosity and pore size increase as complete dolomitization is approached. The cuttings and cores used for this study show considerable variation in the parameters that affect a rock's ability to transmit water. The textural and mineralogical characteristics of the limestones within the trough vary noticeably, on a macroscopic scale, from the textures and mineralogies of the limestones on either side. Adequate petrographic analysis of the cuttings and core is beyond the scope of this study, therefore, only general lithologic trends and spatial relationships will be discussed.

LITHOLOGIC COMPONENTS GULF TROUGH

The Ocala Limestone affords the best example of textural and mineralogical change across the Gulf Trough on a macroscopic scale. The Suwannee Limestone also has textural and mineralogical differences analagous to some of the changes within the Ocala, but they are not as pronounced as those that occur in the Ocala.

On the northwestern side of the trough, particularly in Dougherty and Mitchell Counties, the Ocala is a grain-supported, framework limestone, with a great amount of visible porosity, both primary and secondary. The limestone has a similar lithology and texture throughout: bioclastic granular with coralline algae, bryozoan debris, foraminiferal remains and other kinds of bioclasts. Cementing varies haphazardly in amount and in zones of placement. Dolomitization is minor with dolomite occurring as euhedral rhombs, sparsely scattered throughout the section. Since the limestone is extensively recrystallized the foraminiferal control necessary to identify faunal zone boundaries within the Ocala is lacking.

In oil test well GGS #3154 from Worth County (Fig. 1) the cuttings show that the Ocala Limestone has a visibly porous and not very bioclastic. The low porosity found in the cuttings is supported by the geophysical logs and by the low sustained yield mentioned earlier.

On the southeast side of the trough the Ocala Limestone is texturally and mineralogically more diverse than on the northwest side. The Ocala tends to have thin discrete lithologic beds within the whole. Some of these beds tend to be finer grained and less bioclastic, almost a lithographic limestone. The cuttings from the lower portions of some wells tend to be fine grained, well cemented, with sparse, small, delicate *Lepidocyclinas sp.* Dolomite beds occur in layers 50 (15.25 m) or more feet thick. The dolomite is most commonly saccharoidal, with euhedral crystals. This type of dolomitization usually obliterates the original texture and is generally very porous. These dolomite units cannot be traced laterally with existing controls. Most limestone layers are bioclastic granular, similar to the Ocala on the northwest side of the trough.

The most outstanding difference in the Ocala Limestone on the southeast side of the trough is the presence of gympsum in the lower portion of the limestone. The gypsum occurs in lenses and in intergranular pore spaces. The gypsum is found in close association with both the limestone and dolomite from cuttings in GGS #188 (Fig. 1).located near Moultrie, Colquitt County, and from core in GGS # 3188 located in Thomas County.

In a test hole drilled by the Army Corps of Engineers near Valdosta, Lowndes County, gypsum was encountered in vugs and in intergranular pore spaces in the Ocala Limestone within a zone 900-1000 feet (274.5-305 m) below penetrate the Oligocene surface. One mile (1.6 km) to the northwest of this well, GGS #1687 was drilled to a depth of 700 feet (213.5 m) with the bottom 20 feet (6.1 m) lying within the Oligocene (Fig. 5). In the northern part of Colquitt and southern Tift Counties it is evident that the Miocene sediments are found to be in excess of 900 feet (274.5 m). Indeed, local water well drillers have reported encountering Miocene-type sediments in wells drilled to more than 1000 feet (305 m) deep in the Omega area. GGS well #1419 can be found along the strike of the trough axis in the southwest portion of Colquitt County. In this well the cuttings show that 820 feet (250.1 m) of sediments were drilled and the Oligocene surface was reached at 640 feet (195.2 m).

In the northwest corner of Thomas County, the Miocene units are thinner than they are to the northeast along the trough axis. The Meigs city test well, GGS #3186, appears to be situated close to the trough axis (Fig. 3). In this well the Miocene is 470 feet (143.4 m) thick. GGS wells #3081 and #3101, drilled on the northwestern flank of the trough, show that the Miocene unit thins and disappears, exposing the Suwannee Limestone along the Pelham Escarpment near the town of Baconton, in Mitchell County. On the southeastern flank of the trough the Miocene thins to approximately 150 feet (45.75 m) thick in some areas of central Thomas County. For example, in GGS well #3188, the Miocene is 160 feet (48.8 m) thick.

The dip calculated on the Oligocene surface from GGS #3081 to #3186 is approximately 45 feet per mile (8.57 m per km) to the southeast. The dip calculated from GGS # 886 to #924 is approximately 15 feet per mile (2.88 m per

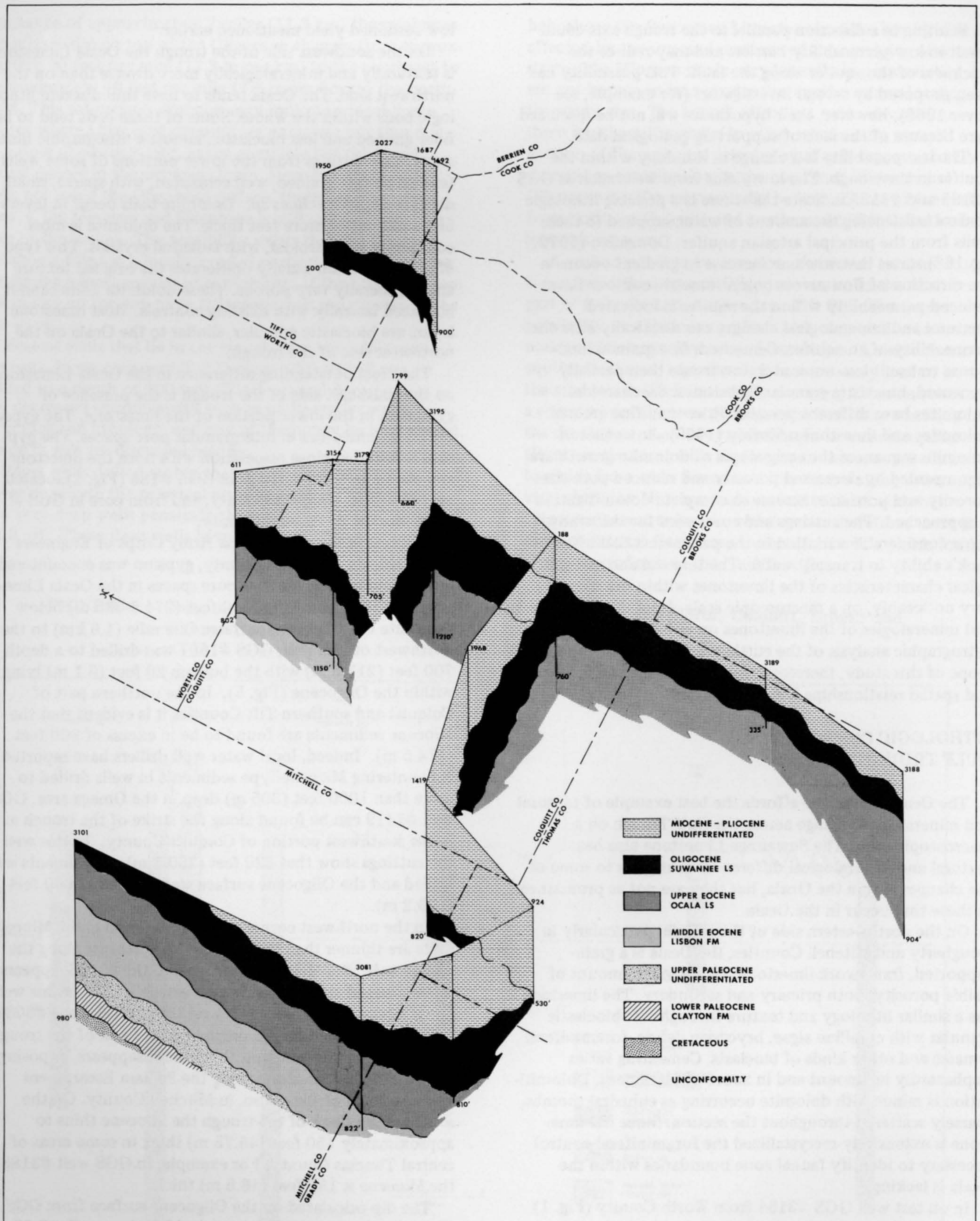


FIGURE 5 Fence diagram of the southwestern portion of the Gulf Trough.

km) to the northwest. The regional dip of the Oligocene surface for the study area outside of the trough is approximately 10 feet per mile (1.9 m per km) to the southeast.

In GGS #3186, at Meigs, the thickness of the Suwannee Limestone is 310 feet (97.6 m) and is assumed to be a complete unit in a disconformable relationship with the overlying Miocene sediments. The Oligocene is somewhat thicker near the trough axis than it is in GGS #3081, northwest of the trough, near the town of Pelham, where the Suwannee is approximately 200 feet (61 m) thick. In the northern Colquitt County area the Oligocene is approximately 200 feet (61 m) thick at the site of GGS #3154, it is 350 feet (106.75 m) at #3195, near Norman Park and 300 feet (91.5 m) at the site of #188 southeast of Moultrie. The Oligocene sediments are thickest along the trough axis and they thin to approximately 130 feet (39.65 m) in southern Thomas County towards the site of GGS well # 3188. If the excessive thickness of the Miocene sediments in the northern Colquitt County area is considered to occur locally then the two areas (northern Colquitt and northwestern Thomas Counties) reflect similar information regarding attitudes and thickness of the Oligocene.

In the above discussion of the subareal extent of the trough, a mental picture of the change in thickness of the various units becomes clear. Along the strike of the trough axis, from southeastern Tift to southwestern Colquitt Counties, the Miocene is thickest. It gets progressively thinner to the southwest. The Oligocene is generally thicker in the trough. Both of these units thin out on the adjacent sides of the trough.

Generally, the Ocala Limestone thickens just southeastward of the area of thickening of the Miocene and Oligocene units. However, some of the top contacts used for the Ocala Limestone in the zone of the trough are estimated due to insufficient faunal data, and most wells never penetrate the underlying formations, therefore, the actual thicknesses still remain to be accurately defined.

INFLUENCE OF THE GULF TROUGH UPON GROUND WATER

The Gulf Trough has a profound effect upon the groundwater availability in south Georgia. It has been previously pointed out that there is a steep increase in gradient of the potentiometric surface in the direction of flow across the axis of the Gulf Trough. This steep gradient indicates that the transmissivity of the principal artesian aquifer has been inhibited.

The unique ground-water conditions within this narrow zone have a negative influence on the potential for local industrial development and community growth. Properly constructed wells in the major portion of Tift, and most of Colquitt, and Thomas Counties, adjacent to the trough, can yield as much as 1000 gallons per minute, whereas within the zone of axis of the trough, wells constructed in excess of 1000 feet (305 m) deep may yield water at a rate of only 50 gallons per minute. Some of these wells are constructed with as much as 500 feet (152.5 m) of open hole below the casing and penetrate only the upper portion of the principal artesian aquifer. The cause of the low yield is not related to the method of well construction but probably to the lithologic character of the subsurface. The expense of drilling wells in excess of 1000 feet (305 m) is prohibitive to potential community and industrial investors, especially if adequate quantities of ground water can be obtained in the areas adjacent to the trough at lesser depths.

Two city water supply test wells drilled near the axis of the trough, GGS #3186 near Meigs, and GGS #3195 near Norman Park, proved to be a surprise in the quantities of water yielded. GGS #3195 was cased to the Miocene-Oligocene contact at 470 feet (143.3 m), therefore water withdrawal is exclusively from the principal artesian aquifer. The well was pumped at the rate of 250 gallons per minute, which was considerably more than the 50 g.p.m. supplied to other wells in that area, but much less than wells outside the influence of the trough.

The city of Meigs water supply well GGS # 3186 was test pumped at 200 gallons per minute. The well was cased to 177 feet (54 m) (within the Miocene) and remained an open hole to a depth of 810 feet (247 m). The yield was much less than the well driller and consulting engineer anticipated, since they planned a principal artesian aquifer well that commonly supplies 1000 gallons per minute in the south part of Thomas County. The yield was quite ample considering that the well is located in the zone of axis of the trough.

DISCUSSION: CAUSES OF LOW YIELDS

The low yielding wells, discussed above, result from decreased transmissivity. There are many possible causes for land surface. The presence of gypsum with no other evaporite species present was determined by X-ray diffraction analyses of this core.

Gypsum has an adverse effect upon the ground water quality on the southeast side of the trough. Stiff diagrams drawn by Zimmerman (1977) show anomalously high sulfate concentrations in the area southeast of the trough and normal readings of sulfate north of and within the trough.

Lithologic characteristics such as texture, mineralogy and biota are ultimately determined by the geochemistry of the depositional environment. The geochemistry is affected by many physical parameters including water temperature, water depth, energy, currents, photic penetration levels and latitude. The prevailing conditions in the water mass in the area of the trough compared to conditions on either side during sediment deposition are responsible for the genetic differences found in the limestones discussed above. The marine depositional environment within the trough may have acted as a partial boundary or separation and the water may not have been free to completely mix across the entire depositional environment.

The original lithology however, is only partly responsible for the diagenetic changes. The flow rates, residence time and flow direction of ground water within the subareally exposed limestones influence diagenesis as well. Equilibrium conditions at the fresh water – salt water interface are also believed to influence the deposition of dolomite and perhaps gypsum as well (Folk and Land 1975).

The lithologic differences that have been found to occur in the aquifer rocks across the trough suggest possible conditions in the depositional environment. One lithologic factor of the limestones found within the trough is the higher micrite content with correspondingly lower amount of biota. This factor may indicate less active currents or other parameter affecting precipitation of microcrystalline calcite. However, the micrite may be diagenetic and in that case the postdepositional environment would be the causitive factor.

Another factor is the thickening of the Oligocene and especially Miocene strata formed in the trough. Assuming that the Miocene and Oligocene are disconformable in all areas of the trough in the study area, then it is possible that there may have been a faster rate of deposition within the trough due to subsidence. The thicker Miocene unit within the trough in Colquitt County may be evidence of more pronounced subsidence in this area of the trough producing local basins of deposition within the length of the trough.

The presence of gypsum on the southeast side of the trough is direct evidence that there is a difference in the postdepositional environments within and on either side of the trough. The limestones deposited within the trough are texturally finer grained and less bioclastic and porous than limestones on the southeast side, and may have acted as a barrier and restricted the postdepositional emplacement of the gypsum to the southeast side of the trough.

In the trough therefore, carbonate deposition may have taken place at a different rate and under dissimilar equilibrium conditions than the adjacent areas of carbonate deposition. These adjacent areas may have developed as a carbonate platform similar to that of the Bahama Banks today and the Gulf Trough may have acted as an area of incomplete separation of that platform with its own unique depositional environment.

SUMMARY AND CONCLUSIONS

The Gulf Trough is a subsurface geologic feature represented by a thickening of Miocene strata and possibly Oligocene and upper Eocene as well. The trough is located along a line trending in a northeast direction from northern Thomas County, Colquitt County, and southern Tift County. The Gulf Trough is distinguished from the feature called the Suwannee Strait by the age of the formations encountered, the lithologic characteristics of these formations and the location of the feature.

The presence of the Gulf Trough has an adverse effect upon the ground-water yields of the principal artesian aquifer. Low water yields may be due to:

1. The principal artesian aquifer being located much deeper in the subsurface than elsewhere so that few wells are drilled deep enough to benefit from the more permeable zones.
2. Multi-aquifer wells that tap the entire Miocene and only the top 20-50 feet (6.1-15.25 m) of the aquifer.
3. The principal artesian aquifer being thinner in the Norman Park area due to erosion, non deposition, or slow rate of deposition of Oligocene sediments.
4. Faulting in a direction parallel to the trough, which may result in a low permeability barrier, such as a reduced aquifer thickness across a fault plane.
5. A carbonate facies change across the feature altering the permeability of the aquifer limestones.

The origin of the thicker deposits within the Gulf Trough may be associated with subsidence related to the formation of the Apalachicola Embayment. The subsidence caused differences in the chemical equilibrium and physical conditions within the water mass across the trough at the time it was active. The geochemical and physical parameters influenced carbonate deposition which in turn affected the postdepositional diagenetic environment, altered the original limestone, caused dolomitization, and controlled the emplacement of gypsum. The resulting limestone types have affected the ground-water conditions that are unique to the central portion of the Coastal Plain of Georgia.

ACKNOWLEDGMENTS

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TABLE 1 List of GGS wells selected for this study. Unless otherwise stated, well samples are cuttings.

Brooks County

GGS 3189* (core)
Location: 30° 56′ 26″ N
83° 44′ 06″ W
Altitude: 200′ (61m)
T. D. (Total Depth): 84-335′ (25.6 - 102.2m)
Depth to Suwannee Lm.: 148′ (45.1m)
Depth to Ocala Lm.: 292′ (89m)

Colquitt County

GGS 188
Location: 31° 08′ 15″ N
83° 42′ 30″ W
Altitude: 288′ (87.8m) (above sea level)
T. D.: 760′ (231.8m) (below land surface)
Depth to Suwannee Lm.: 245′ (74.7m) (below land surface)
Depth to Ocala Lm.: 545′ (166.2m) (below land surface)

GGS 1419
Location: 31° 08′ 15″ N
83° 57′ 30″ W
Altitude: 307′ (93.6m)
T. D.: 820′ (in Miocene) (250m)

GGS 1799
Location: 31° 18′ 00″ N
83° 38′ 45″ W
Altitude: 285′ (86.9m)
T. D.: 660′ (in Miocene) (201.3m)

GGS 1968
Location: 31° 09′ 33″ N
83° 49′ 55″ W
Altitude: 320′ (97.6m)
T. D.: 800′ (244m)
Depth to Suwannee Lm.: 480′ (146.4m)
Depth to Ocala Lm.: 670′ (204.4m) (estimated)

GGS 3195*
Location: 31° 15′ 13″ N
83° 40′ 22″ W
Altitude: 330′ (100.6m)
T. D.: 1200′ (366m)
Depth to Suwannee Lm.: 470′ (143.4m)
Depth to Ocala Lm.: 450′ (137.3m)
Depth to Lisbon Fm.: 1080′ (329.4m)

GGS 3179 (core)
Location: 31° 17′ 33″ N
83° 43′ 24″ W
Altitude: 370′ (112.8m)
T. D.: 705′ (in Miocene) (215m)

Dougherty County

GGS 3173* (core)
Location: 31° 35′ 29″ N
84° 20′ 24″ W
Altitude: 210′ (64m)
T. D.: 675′ (205.8m)
Depth to Lisbon Fm.: 94′ (28.6m)
Depth to upper Paleocene: 351′ (107m)
Depth to lower Paleocene: 500′ (152.5m)
Depth to Cretaceous: 661′ (201.6m)

GGS 3187 (core)
Location: 30°31′ 05″ N
84° 06′ 44″ W
Altitude: 195′ (59.5m)
T. D.: 1515′ (462m)
Depth of well: 79.3-1401.3 (24.2-427.9m) at 79.3′ (24.1m) in Ocala Lm.
Depth to Lisbon: 243.8′ (74.4m)
Depth to upper Paleocene: 417.1 (127.2)
Depth to lower Paleocene: 709.8′ (216.5m)
Depth to Cretaceous: 928′ (283m)

Mitchell County

GGS 3101
Location: 31° 22′ 40″ N
84° 09′ 52″ W
Altitude: 176′ (53.7m)
T. D.: 973′ (296.7m)
Depth to Ocala Lm.: 50′ (15.25m)
Depth to Lisbon Fm.: 308′ (93.9m)
Depth to upper Paleocene: 664′ (202.5m)
Depth to Cretqceous: 912′ (278.2m)

GGS 3081
Location: 31° 07′ 11″ N
84° 08′ 39″ W
Altitude: 348′ (106.2m)
T. D.: 822′ (250.7m). No samples 274-422′ (83.6-128.7m) in Ocala Lm. at 422 (128.7m)
Depth to Lisbon Fm.: 722′ (220.2m)

Thomas County

GGS 886
Location: 30° 58′ 00″ N
84° 02′ 35″ W
Altitude: 255′ (77.8m)
T. D.: 530′ (161.7m)
Depth to Suwannee Lm.: 395′ (120.5m)

GGS 924
Location: 31° 01′ 25″ N
84° 03′ 40″ W
Altitude: 205′ (62.5m)
T. D.: 530′ (161.7m)
Depth to Suwannee Lm.: 500′ (152.5m)

Thomas County (Cont'd)

GGS 3186*
Location: 31° 03′ 53″ N
84° 05′ 12″ W
Altitude: 330′ (100.6m)
Depth: 810′ (247m)
Depth to Suwannee Lm.: 470′ (143.4m)
Depth to Ocala Lm.: 780′ (238m) (estimate)

GGS 3188* (core)
Location: 30° 48′ 39″ N
83° 45′ 23″ W
Altitude: 200′ (61m)
Depth: 70-904′ (21.4-275.7m)
Depth to Suwannee Lm.: 162′ (49.4m)
Depth to Ocala Lm.: 289′ (88.1m)

Tift County

GGS 1692
Location: 31° 20′ 55″
83° 27′ 17″ W
Altitude: 329′ (100.3m)
T. D.: 900′ (274.5m) (in Miocene)

GGS 1687
Location: 31° 22′ 10″ N
83° 27′ 15″ W
Altitude: 324′ (98.8m)
T. D.: 700′ (213.5m)
Depth to Suwannee Lm.: 640′ (195.2m)

GGS 2027
Location: 31° 23′ 40″ N
83° 27′ 50″ W
Altitude: 328′ (100m)
T. D.: 605′ (184.5m)
Depth to Suwannee Lm.: 575′ (175.4m)

GGS 1950
Location: 31° 25′ 10″ N
83° 30′ 00″ W
Altitude: 335′ (102.2m)
T. D.: 500′ (152.5m)
Depth to Suwannee Lm.: 390′ (118.95m)

Worth County

GGS 611
Location: 31° 23′ 40″ N
83° 49′ 00″ W
Altitude: 330′ (100.6m)
Depth: 55-802′ (16.8-244.6m)
Depth to Suwannee Lm.: 252′ (76.86m)
Depth to Ocala Lm.: 638′ (194.6m)

Worth County (Cont'd)

GGS 3154*
Location: 31° 19′ 03″ N
83° 44′ 15″ W
Altitude: 325′ (99.1m)
T. D.: 5568′ (1698m). Depth completed 1780′ (542.9m) - Cretaceous
Depth to Suwannee Lm. surface: 480′ (146.4m)
Depth to Ocala Lm.: 690′ (210.45m)
Depth to Lisbon: 1060′ (323.3m)
Depth to upper Paleocene: 1190′ (362.95m)

*Geophysical logs are available for these wells.

THE SANDY SPRINGS GROUP AND RELATED ROCKS IN THE GEORGIA PIEDMONT — NOMENCLATURE AND STRATIGRAPHY

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ABSTRACT

The Sandy Springs Group is named for the town of Sandy Springs, Georgia. Four formations comprise the group (listed from probable oldest to youngest): the Powers Ferry Formation, which includes the Mableton Amphibolite Member; the Chattahoochee Palisades Quartzite; the Factory Shoals Formation; and the Rottenwood Creek Quartzite. These units extend from Alabama to South Carolina and can be correlated with units of other studies along strike.

Two massive mica gneiss units are present along the southeastern boundary of the Sandy Springs Group. These units are termed the Long Island Creek Gneiss and Yellow Dirt Gneiss. Although they border the Sandy Springs Group for long distances, these gneissic units are bounded by faults and their stratigraphic position is unknown.

INTRODUCTION

Northwest of the Brevard Zone, in the Georgia Piedmont, is a series of long, linear, northeast-trending ridges generally held up by quartzites. The valleys and slopes between the ridges are underlain by schists, gneisses, and amphibolites. The topography of this area is greatly different from the area to the southeast and to the northwest (Clark and Zisa, 1976; Fig. 1, this paper). The rocks of this area form a distinct structural and stratigraphic sequence. Higgins (1966, 1968) used the informal name "Sandy Springs Sequence" for these rocks. Unfortunately, he capitalized Sequence erroneously implying a formal name. Since then, other workers in the area have used the name "Sandy Springs Sequence" in a more or less formal way. Sequence is not a recognized stratigraphic name (American Commission on Stratigraphic Nomenclature, 1970) and should be abandoned.

The purpose of this paper is to formalize the name Sandy Springs Group, to define the group and its constituent formations and members of formations, and to propose correlations with other units along strike.

SANDY SPRINGS GROUP

The Sandy Springs Group is here formally named for the town of Sandy Springs, Georgia (see U. S. Geological Survey, Sandy Springs, Georgia, 7.5 min. topographic quadrangle), where the name "Sandy Springs Sequence" was used

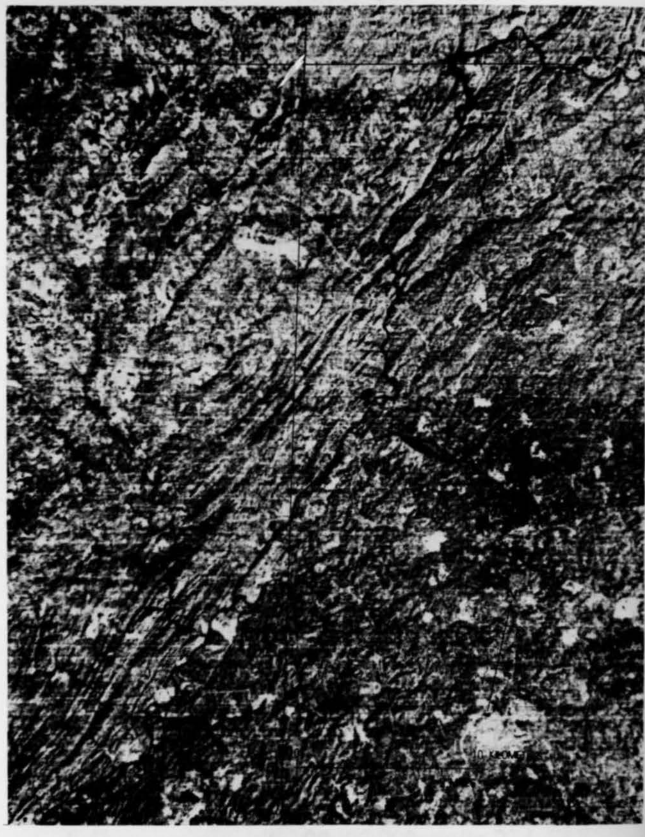


FIGURE 1 Landsat image showing the topographic expression of the Sandy Springs Group.

TABLE 1 Stratigraphic units of the Sandy Springs Group

Rottenwood Creek Quartzite

Factory Shoals Formation

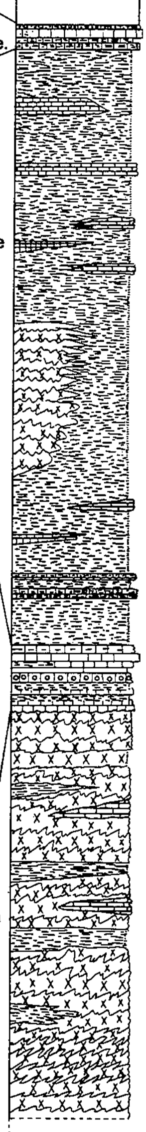
Chattahoochee Palisades Quartzite

Powers Ferry Formation

Mableton Amphibolite Member

FIGURE 2 Diagrammatic stratigraphic column of the Sandy Springs Group

FORMATION	THICKNESS, IN METERS	DESCRIPTION	LITHOLOGY
Rottenwood Creek Quartzite	50	Micaceous quartzite and massive quartzite.	
Factory Shoals Formation	1000–2000	Kyanite and staurolite schist with interbedded micaceous quartzites, amphiboles, and quartz-muscovite-biotite-plagioclase gneiss.	
Chattahoochee Palisades Quartzite	100 unconformity(?)	Micaceous quartzite, muscovite schist, massive quartzite. Ocherous layer, feldspathic quartzite. Micaceous quartzite with garnets alternating with muscovite schist.	
Powers Ferry Formation including Mableton Amphibolite Member	More than 1000	Biotite-oligoclase-microcline gneiss with intercalated muscovite-biotite schists, amphibolites, and biotite amphibolites.	



(Higgins, 1966, 1968), and where most of the formations that comprise the group are fairly well exposed. The group consists of the formations listed in Table 1. The age of the group is unknown, but is probably late Precambrian and (or) early Paleozoic. The thickness of the group (apparent thickness because of folding) and its constituent formations and the apparent stratigraphic relations, are shown in Figure 2. Rocks of the Sandy Springs Group have now been

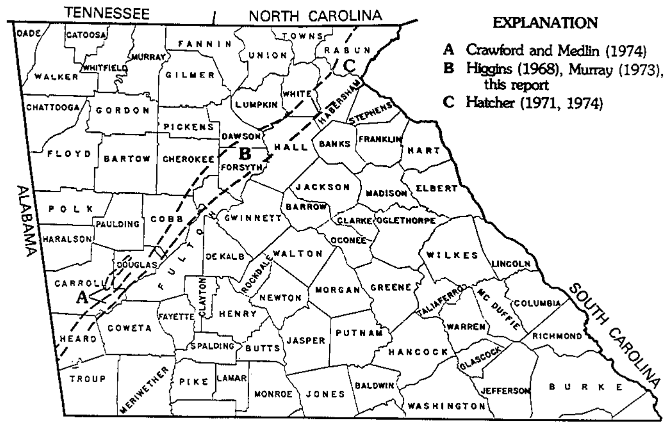


FIGURE 3 County outline map of Georgia showing known outcrop area (inside dashed lines) of the Sandy Springs Group. A-Crawford and Medlin (1974); B-Higgins (1968), Murray (1973), this report; C-Hatcher (1971).

mapped as far northeast as South Carolina (Fig. 3; Murray, 1973, and unpub. data; Hatcher, 1971, 1974) and as far southwest as Alabama (Crawford and Medlin, 1974). Across strike, they extend from the Brevard Zone as far as 30 km (19 mi.) to the northwest (Fig. 3). Measured sections through parts of the group are given in an open file report (Higgins and McConnell, 1978).

Crawford and Medlin (1974) proposed many informal names for units they mapped in the "Sandy Springs Sequence." These names are, for the most part, not adopted in this paper because they were informal and were published only in a field trip guide. We prefer to use group names for geographic localities as near as possible to the type area.

Powers Ferry Formation

The Powers Ferry Formation is here named for Powers Ferry (now called Powers Ferry Landing, where Powers Ferry once crossed the Chattahoochee River; U. S. Geological Survey, Sandy Springs, Georgia, 7.5 min. topographic quadrangle) where the rocks are well exposed in new cuts in the Powers Ferry Landing shopping center and office park (Fig. 4). The Powers Ferry Formation consists of the rocks that Higgins (1966, 1968) called "the gneiss-schist-amphibolite unit," and of other lithologies now mapped along strike by Hatcher (1971, 1973), Murray (1973), Crawford and Medlin (1974), and McConnell (unpub. data: also see McConnell and Abrams, in same publication).

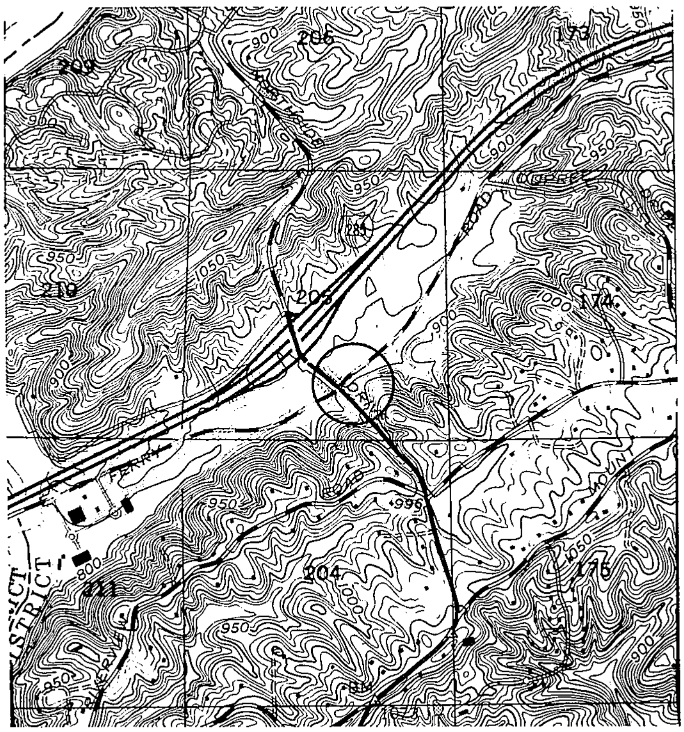


FIGURE 4 The type locality for the Powers Ferry Formation (U. S. Geological Survey, Sandy Springs, Georgia, 7.5 min. topographic quadrangle).

As mentioned above, the Powers Ferry Formation consists mainly of intercalated gneiss, schist, and amphibolite in decreasing order of abundance. One large mappable amphibolite unit is named below.

The gneiss in the Powers Ferry Formation is generally a medium-grained gray, discontinuously striped muscovite-biotite-microcline-quartz-oligoclase (An_{12}) gneiss, locally with pink microcline augen as much as 7 cm (2.75 inches) long. In places it contains small amounts of epidote, opaque minerals, and garnet. Concordant quartz-microcline pegmatites are fairly common, and pods and discontinuous layers of muscovite-biotite-quartz schist, amphibolite, and hornblende schist are common and locally comprise most of the formation. Parts of the gneiss resemble massive quartzite. Some biotite-quartz-plagioclase gneiss (probably metagraywacke) occurs within the formation, and specularite quartzite and layered amphibolite are mappable for short distances.

The Powers Ferry Formation is apparently the oldest formation in the Sandy Springs Group. The base of the formation has not been seen because the Powers Ferry Formation is present only in the cores of antiforms. The formation is in sharp contact, perhaps unconformably, with the overlying Chattahoochee Palisades Quartzite (Fig. 2). This contact is well exposed in cuts of the Louisville and Nashville Railroad on the flanks of the Vinings antiform (Vinings anticline of Higgins, 1966, 1968) near Vinings, Georgia (U. S. Geological Survey, Northwest Atlanta, Georgia 7.5 min. topographic quadrangle). The actual thickness of the Powers Ferry Formation is unknown because the base is not exposed. Based on map distribution the formation is estimated to be more than 1000 m (300 ft.) thick.

Mableton Amphibolite Member

Layered hornblende-plagioclase amphibolites within the Powers Ferry Formation are here named the Mableton Amphibolite Member (Fig. 5) for exposures along and north of Fountain Road just northeast of Mableton (U. S. Geological Survey, Mableton, Georgia, 7.5 min. topographic quadrangle; McConnell, unpub. data). Higgins (1968) showed two belts of disconnected layered amphibolite within the Powers Ferry Formation (gneiss-schist-amphibolite unit as mapped by Higgins), but McConnell's recent work has shown that they are connected. The amphibolite appears to be 500-700 m (1650-2300 ft.) thick based on map distribution.

Chattahoochee Palisades Quartzite

The Chattahoochee Palisades Quartzite (Fig. 6) is here named for the Palisades of the Chattahoochee River in Chattahoochee Palisades State Park (U. S. Geological Survey, Sandy Springs, Georgia, 7.5 min. topographic quadrangle) where it forms cliff outcrops (see Higgins, 1968, lower quartzite unit). It is commonly a massive white,

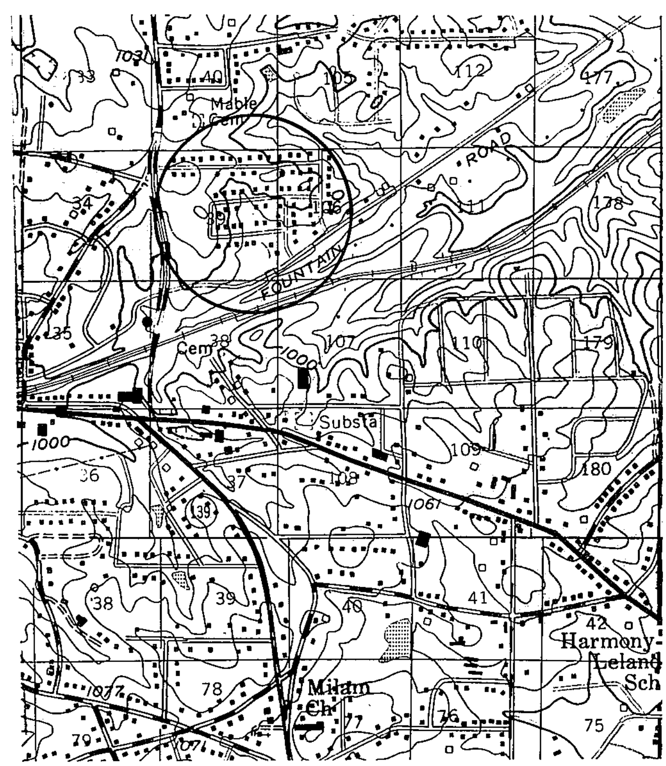


FIGURE 5 The type locality for the Mableton Amphibolite Member (U. S. Geological Survey, Mableton, Georgia, 7.5 min. topographic quadrangle).

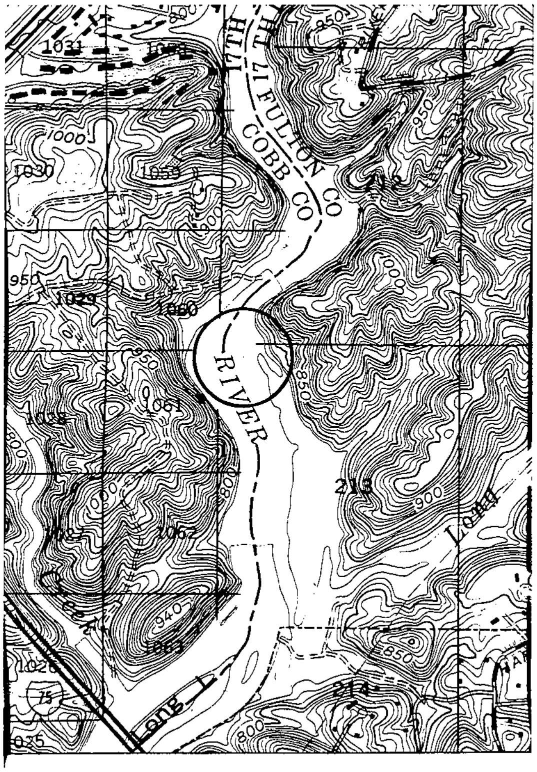


FIGURE 6 The type locality for the Chattahoochee Palisades Quartzite (U. S. Geological Survey, Sandy Springs, Georgia, 7.5 min. topographic quadrangle).

yellowish, or bluish, sugary to vitreous quartzite containing accessory mica and elongated garnets. Locally it displays graded bedding, but it is more common to find beds 0.3 - 1.2 m (1-4 ft.) thick that are interlayered with feldspathic quartzite and muscovite schist. Along strike, variations to quartzose mica schist also occur due to sedimentary facies changes. The bottom of the quartzite is in sharp, possibly unconformable, contact with the underlying Powers Ferry Formation. The top grades over an interval of a few meters into rocks of the Factory Shoals Formation (see below). Based on map distribution, the quartzite appears to be about 100 m (329 ft.) thick, but is locally absent perhaps due to nondeposition.

Factory Shoals Formation

The Factory Shoals Formation (Fig. 7) is here named for exposures along Sweetwater Creek at Factory Shoals (U. S. Geological Survey, Ben Hill, Georgia, 7.5 min. topographic quadrangle; see McConnell and Abrams, in same publication). It includes the rocks mapped by Higgins (1968) as the aluminous schist unit. The Factory Shoals Formation is commonly a light-gray lustrous garnet-biotite-oligoclase-muscovite-quartz schist locally containing kyanite or staurolite. It also contains layers of quartz muscovite schist, thinly bedded red micaceous quartzite, muscovite-biotite-plagioclase metagraywacke, and graphitic muscovite-quartz schist. In some localities metagraywacke predominates. It is

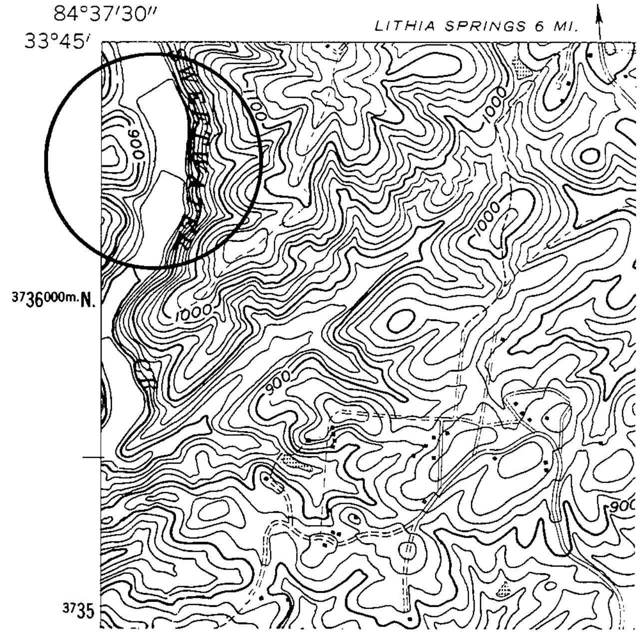


FIGURE 7 The type locality of the Factory Shoals Formation (U. S. Geological Survey, Ben Hill, Georgia, 7.5 min. topographic quadrangle).

in gradational contact (see above) with the Chattahoochee Palisades Quartzite at its base, and grades over a short interval into the Rottenwood Creek Quartzite at its top (Fig. 2). Based on map distribution (Higgins, 1968; McConnell, unpub. data) the formation appears to be 1000-2000 m (3290-6570 ft.) thick.

Rottenwood Creek Quartzite

The Rottenwood Creek Quartzite (Fig. 8) is here named for excellent exposures along Rottenwood Creek about 1000 m (3290 ft.) southeast of Akers Mill Road (U. S. Geological Survey, Sandy Springs, Georgia, 7.5 min. topographic quadrangle). It is commonly a massive muscovite-bearing quartzite containing small amounts of plagioclase, epidote, and opaque minerals (Higgins, 1966). It is the uppermost unit of the Sandy Springs Group. At the base it grades over a short interval into the Factory Shoals Formation. It crops out in the cores of synforms, and is absent in many places due to erosion or perhaps nondeposition. Based on map distribution (Higgins, 1968), it is approximately 50 m (164 ft.) thick.

RELATED ROCKS

Long Island Creek Gneiss

The Long Island Creek Gneiss (Fig. 9) is here named for exposures on the southeast side of Long Island Creek near and along Roswell Road (U. S. Geological Survey, Sandy Springs, Georgia, 7.5 min. topographic quadrangle; Higgins, 1968). It is in fault contact with all other adjacent rocks so its stratigraphic position is unknown. It is named here

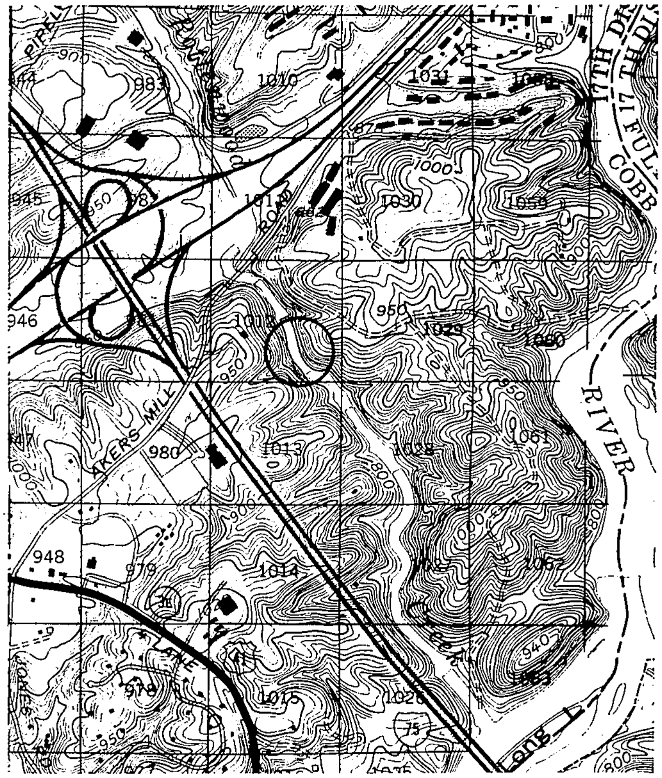


FIGURE 8 **The type locality of the Rottenwood Creek Quartzite (U. S. Geological Survey, Sandy Springs, Georgia, 7.5 min. topographic quadrangle).**

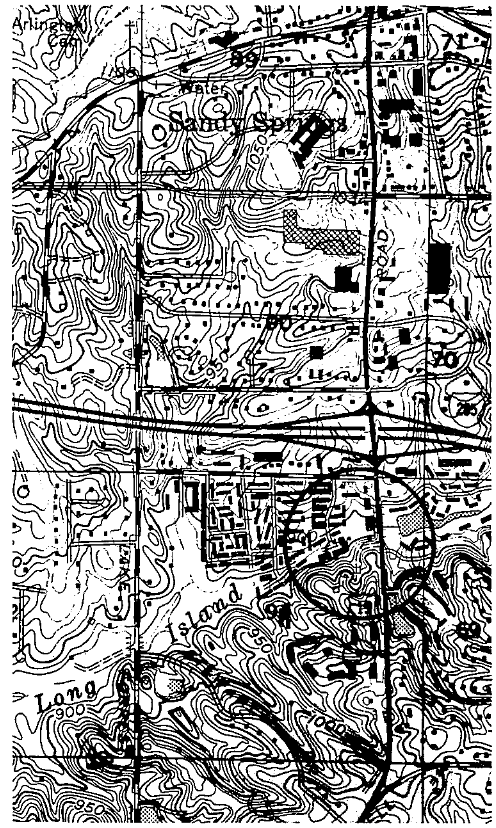


FIGURE 9 The type locality of the Long Island Creek Gneiss (U. S. Geological Survey, Sandy Springs, Georgia, 7.5 min. topographic quadrangle).

because it forms a mappable boundary of the Sandy Springs Group. The Long Island Creek Gneiss is typically a dark gray epidote-biotite-plagioclase gneiss. It contains euhedral crystals of sphene and a few tiny garnets. It weathers to a massive yellow saprolite. Good exposures of typical Long Island Creek Gneiss are seen in cuts for driveways at the new I.B.M. facility on U. S. Highway 41 just southeast of the Chattahoochee River (U. S. Geological Survey, Northwest Atlanta, Georgia, 7.5 min. topographic quadrangle). The thickness of the Long Island Creek Gneiss is unknown.

Yellow Dirt Gneiss

Crawford and Medlin (1974, p. 11) informally named the "Yellow Dirt gneiss" for the community of Yellow Dirt, Heard County, Georgia (U. S. Geological Survey, Lowell, Georgia 7.5 min. topographic quadrangle). The gneiss is commonly a fine- to medium-grained, biotite-epidote-muscovite-quartz-plagioclase-microcline rock that has strong cataclastic textures (blastomylonite and mylonite gneiss, see Higgins, 1971). It bounds the Sandy Springs Group for long distances, and is here proposed as a formal formation with the type locality as proposed by Crawford and Medlin (1974; p. 11). The gneiss is estimated to be 75-240 m (250-800 ft.) thick (see Crawford and Medlin, 1974). Because it is bounded by faults, its stratigraphic position is unknown.

CORRELATIONS

Table 2 gives the correlation of stratigraphies proposed by Crawford and Medlin (1974), Higgins (1966, 1968) and the authors, as proposed in this paper. To the northeast, the rocks of the Sandy Springs Group are probably correlative, as indicated, with some of the rocks in the Tallulah Falls area (Higgins, 1966; Hatcher, 1974).

ACKNOWLEDGMENTS

We would like to thank S. M. Pickering, Jr., R. D. Hatcher, Jr., and D. J. Milton for constructive reviews.

Table 2.—Correlation chart of Sandy Springs Group with other informal units in northwestern Georgia.

SW NE

<table>
<tr><th>"Sandy Springs Sequence"
Crawford & Medlin (1974)</th><th>"Sandy Springs Sequence"
Higgins (1966,1968)</th><th>Sandy Springs Group
This paper</th><th>Tallulah Falls Formation
Hatcher (1974)</th></tr>
<tr><td>Mt. Olive Church (schist)</td><td>Not present</td><td>Not present</td><td>Not present</td></tr>
<tr><td>Adamson quartzite</td><td>Upper quartzite</td><td>Rottenwood Creek Quartzite</td><td>Quartzite-schist member and graywacke schist member(?)</td></tr>
<tr><td>Backbone schist</td><td rowspan="3">Aluminous schist</td><td rowspan="3">Factory Shoals Formation</td><td rowspan="3">Garnet-aluminous-schist member</td></tr>
<tr><td>Anneewakee graphitic schist-quartzite</td></tr>
<tr><td>Sparks Reservoir (schist and gneiss)</td></tr>
<tr><td>Dry Creek quartzite</td><td>Lower quartzite</td><td>Chattahoochee Palisades Quartzite</td><td>Not defined</td></tr>
<tr><td>Chapel Hill Church (gneiss and schist)</td><td rowspan="3">Gneiss-schist-amphibolite</td><td rowspan="3">Powers Ferry Formation including Mableton Amphibolite Member</td><td rowspan="3">Graywacke-schist amphibolite member</td></tr>
<tr><td>Mt. Vernon Church graphitic schist-quartzite</td></tr>
<tr><td>Mt. Vernon Church schist</td></tr>
</table>

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STRATIGRAPHY OF THE TOBACCO ROAD SAND – A NEW FORMATION

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ABSTRACT

The Tobacco Road Sand is a belt of coastal marine sands of late Eocene age. This belt of coastal sands of probable sound or lagoon origin lies adjacent to and parallel with the Fall Line in eastern Georgia and western South Carolina. It grades downdip to the south into carbonate deposits; the Cooper Marl of South Carolina and its equivalent in Georgia. It extends from the vicinity of Houston and Pulaski Counties, Georgia, in the west to Aiken County, South Carolina, in the east; a distance of approximately 130 miles (209 km). The width of the existing outcrop and subcrop belt varies from approximately 30 miles (48 km) in the west to 40 miles (64 km) in the east. The known thickness of the formation ranges from 9.5 feet (2.9 m) to 30 feet (9.1 m) but averages about 20 feet (6.1 m). The original volume of the deposit, therefore, was approximately 17 cubic miles (71 cu. km).

Regional uplift to the west in Alabama has caused the outcrop belt of the Tobacco Road Sand to grade westwardly into its offshore equivalents, the "Cooper Marl" and the Ocala Limestone. Similarly, because of the uplift along the Cape Fear Arch in North Carolina, the Tobacco Road Sand passes eastward into the more offshore Cooper Marl between the Savannah and Santee Rivers in South Carolina.

The Tobacco Road Sand overlies, in some places gradationally, in other places with apparent discontinuity, either the Irwinton Sand or the Twiggs Clay. The Sandersville Limestone, known only from the Sandersville area of Washington County, Georgia, represents the discontinuous, basal carbonate facies of the Tobacco Road Sand that in other areas has either been completely leached or silicified.

Physical and paleontological correlation are consistent with a late Eocene age for the Tobacco Road Sand.

INTRODUCTION

The Tobacco Road Sand is a deposit of coastal marine sands that is probably of beach, sound or lagoon, and possibly in places, of estuarine origin. In the past, these sands have invariably been included in the Barnwell Formation. LaMoreaux (1946a, 1946b) was the first to recognize these sands as a subdivision of the Barnwell Formation and he referred to them as the upper sand member of the Barnwell Formation. He identified the upper sand member in Twiggs, Wilkinson, and Washington Counties, Georgia. Although other authors have extended the known area of occurrence of the upper sand member of LaMoreaux, its recognition and stratigraphic definition have been incomplete and inconsistent. The standard Barnwell lithology is based on a bland, massive, deep red, residuum, and is by definition late Eocene in age (Cooke and Shearer, 1918, p. 49, 52; Cooke, 1936, p. 89; Cooke, 1943). The subdivisions within the Barnwell residuum can not be traced any distance with confidence. It is therefore, the purpose of this paper to propose that the upper sand member of the Barnwell as a new formation, the Tobacco Road Sand. The authors will describe its range of lithologies as thoroughly as current data permits, and will define the formation in terms of its stratigraphic relationships, age, origin, and areal distribution.

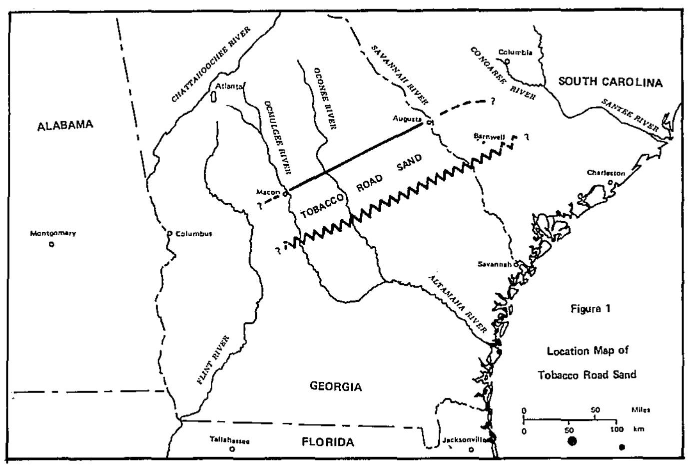


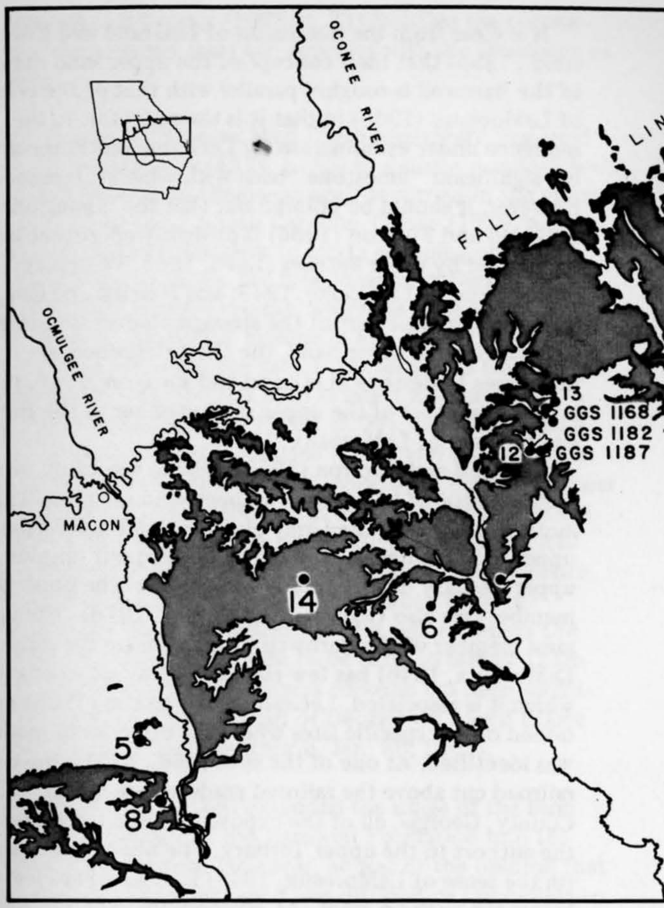
Figure 1

Location Map of Tobacco Road Sand

HISTORICAL REVIEW

LaMoreaux (1946a and 1946b) first recognized and differentiated the lithostratigraphic unit, referred to here as the Tobacco Road Sand, as the upper sand member of the Barnwell Formation. His description of the unit (1946b, p. 63, 64), is as follows:

> In the southern and central half of the area of outcrop of upper Eocene deposits, a coarse red sand overlies the thin clay bed at the top of the Irwinton sand member of the Barnwell formation. Characteristic of this bed in the area of the report are the flat polished beach pebbles scattered along the base of the bed. These pebbles are believed to be derived from flat



OUTCROP BELT JACKSONIAN DEPOSITS WITH LOCATIONS OF CITED OUTCROPS AND CORES

FIGURE 2

fragments of the many resistant quartz veins in the weathered schist and gneiss of the Piedmont north of the Coastal Plain in eastern Georgia. The pebbles range from 1/4 inch to 2 inches in diameter, and were probably rounded by wave action along a beach.

The upper coarse pebbly sand is nearly everywhere highly weathered, and seldom exceeds 20 feet in thickness, although in its southern area of outcrop near the contact of the deposits of Jackson age with the undifferentiated Oligocene and Miocene residuum, it may thicken to a maximum of 25 to 30 feet. The exact thickness of this bed is hard to determine because of the beds above. The member may have originally been an upper Eocene limestone which is now represented only by a weathered coarse red angular quartz sand. Because its lithology is similar to the typical Barnwell formation of eastern Georgia and South Carolina, it is believed that this bed probably represents a thin upper member of the Barnwell formation in east-central Georgia.

LaMoreaux (1946a, p. 21) also states that:

> The upper sand member appears to lie unconformably on the Irwinton sand member and...... It may represent a thin upper member of the Barnwell or possibly a thin residual bed of Oligocene or Miocene age. In the report it is considered the upper member of the Barnwell formation and was mapped as such.

Additionally, it is clear from his description of measured sections (ibid., p. 64, 96, 97, 120, 148) of the upper sand member that his concept of the unit was a massive-bedded, coarse sand that contains quartz granules. The quartz grains are subangular to subrounded, and thin clay stringers may be locally evident. Cross-bedding is present locally. Polished, rounded, flat quartz pebbles or gravel are common in the Twiggs, Wilkinson, and Washington Counties area at the base of the unit.

LeGrand and Furcron (1956) recognized the upper sand member of LaMoreaux (1946) east of Washington County in Georgia. Their general discussion (ibid., p. 37) is as follows:

> The upper sand member of the Barnwell formation in eastern Georgia consists of coarse red sand and mottled red sandy clay. It is typical of the undifferentiated Barnwell formation of Barnwell County, S.C. This upper sand member is generally massive and commonly mottled with streaks of gray clay.
>
> The upper sand member includes at least one limestone bed; fragments of limestone at several horizons indicated that several beds of limestone were present before being removed in solution by circulating ground water. The Sandersville limestone member of

the Barnwell formation described by Cooke (1943, p. 62) may be represented in the upper 30 feet of the Barnwell formation. Fossils are not known to be present in the massive red sand beds, although the various limestone beds are everywhere fossiliferous. The sporadic occurrence of limestone and the mottled character of the sandy clay indicates that intrastratal solution has occurred in the bed (or beds) and that this solution has resulted in subsidence of the overlying sand and clay. The subsidence has dissarranged the original bedding into a nondescript mass of red sand and mottled sandy clay.

> The thickness of the upper sand member is generally indeterminable because its contact with the underlying Irwinton sand member is indistinct. Also, the overlying residue of the Suwannee limestone closely resembles the upper sand member of the Barnwell formation, making this contact difficult to map. It is believed that the upper sand member of the Barnwell formation is of Jackson age, although a part of this sand in southeast Burke County may represent material weathered from limy deposits of Oligocene age.

They (ibid., p. 50, 51) go on to add:

> The members of the Barnwell described farther west are hardly distinguishable in Burke County. The formation here is composed of red sands in which one or more limestone and chert beds occur. One fossiliferous limestone bed is present within the upper 40 feet of the Barnwell formation. This bed, ranging in thickness from 2 to 20 feet, may be equivalent to the Sandersville limestone in Washington County. The Barnwell is approximately 100 feet thick at Greens Cut, 200 feet at Waynesboro, and 220 feet at Midville in the southern part of the County.

On page 69 (ibid.):

> The upper sand member of the Barnwell formation is exposed east and north of Gibson. It is usually characterized by brilliant red sands and gray mottled sand clay. Thin limestone beds present in the upper sand member in Jefferson and Burke Counties were not noticed in Glascock County.

And (ibid., p. 75):

> The upper sand member of the Barnwell formation is extensively exposed on the Louisville plateau north of Louisville and along the lowlands in the southern part of the county (Jefferson). A relatively thin limestone bed within this upper sand member, largely dissolved away, is probably the reason for the many sinks on the upland.

It is clear from the discussion of LeGrand and Furcron (ibid., 1956) that their concept of the upper sand member of the Barnwell is roughly parallel with that of the concept of LaMoreaux (1946) in that it is the only unit in the sequence under examination by LeGrand and Furcron that has significant "limestone" beds with which it is associated. However, it should be pointed out that the "limestone" of LeGrand and Furcron (1956) is probably chert that has been recognized by other authors (Lyell, 1845; Veatch and Stephenson, 1911; Cooke, 1943; and Herrick and Counts, 1969) in the same part of the stratigraphic section in eastern Georgia. On the other hand, the chert described as Suwannee Limestone (LeGrand and Furcron, 1956, p. 51) is the basal chert of the upper sand member of the Barnwell (in the sense of LaMoreaux, 1946).

LeGrand and Furcron (1956) include too many units in their concept of LaMoreaux's upper sand member. The mottled, sandy clay and gray clay repreatedly referred to as upper sand member of the Barnwell is, in our opinion, an upper Tertiary deposit and is not a part of the upper sand member (see also Georgia Geol. Survey, 1976). The upper sand member of the Barnwell Formation (in the sense of LaMoreaux, 1946) has few clay and sandy clay beds with which it is associated. LeGrand and Furcron (1956) mentioned only 3 specific sites where the upper sand member was identified. At one of the sites (ibid., p. 51), the deep railroad cut above the railroad grade at Greens Cut, Burke County, Georgia, all of the exposed section is referred by the authors to the upper Tertiary. The upper sand member (in the sense of LaMoreaux, 1946) is exposed at elevations below the railroad grade. At Waynesboro in Burke County, another of the localities mentioned by LeGrand and Furcron (ibid., p. 51), upper Tertiary sands and gravels (high river terrace?) overlie the upper sand member in the vicinity of Brier Creek.

Herrick and Counts (1968) recognized the upper sand member of the Barnwell Formation in Burke County, Georgia, but referred to the unit under both the terms unnamed upper sand and sands of Oligocene(?) undifferentiated age:

> In this report these upper sands in eastern Georgia, particularly in the subsurface where their relationship with other units is more evident, are regarded as the probable up-dip equivalent of the Flint River - Suwannee Formations and thus are considered to be of Oligocene age. For the sake of convenience the upper sands, together with the Irwinton Sand Member, are in instances, necessarily included under the term, Oligocene(?) undifferentiated. (Herrick and Counts, 1968, p. 39).

In addition, they (ibid., p. 41) suggest that the basal Jacksonian, Griffins Landing beds are correlative with the "sands of Oligocene(?) undifferentiated age" and with the "Cooper Marl"[1] (ibid., p. 61 (Stop 5); also see Huddlestun, Marsalis, and Pickering, 1974, p. 2-9).

[1] We believe that the Cooper Marl, as it has been recognized in Georgia, is a distinct lithostratigraphic unit and should be differentiated from the Cooper Marl of South Carolina. Therefore, in this paper, our references to the unit that has been called Cooper Marl in Georgia will be accompanied by quotation marks. The references to the typical South Carolina Cooper Marl will not be accompanied by quotation marks.

Herrick and Counts (1968, p. 31) justified the correlation of these upper Barnwell deposits with the Oligocene as follows:

> The term, "Cooper Marl," was apparently first used by Stephenson (1914, p. 85) to designate certain beds of "Eocene or Oligocene age" occurring in the vicinity of Charleston, South Carolina. Cooke (1936, p. 82) first applied this term to strata cropping out along the Cooper River, the type locality in Berkeley County, South Carolina, which he regarded at that time as uppermost Eocene in age. Later Cooke (1943, p. 74) extended the use of this term into Georgia but he still regarded this marl as latest Eocene in age. Cooke and MacNeil (1952, p. 27), on the basis of a mollusk, *Chlamys cocoana* (Dall), and a foraminifer *Bolivinella rugosa* (Howe), concluded that the Cooper Marl in South Carolina was earliest Oligocene in age and equivalent to the Red Bluff Formation of Alabama and Mississippi. On the basis of field evidence, more particularly Foraminifera taken from the Cooper Marl, the authors of this report believe the Cooper Marl in Georgia to be early to middle Oligocene in age. Also, *Chlamys cocoana* has been identified in the Cooper Marl of Georgia by Lloyd Glawe (Personal Communication, Jan. 30, 1965).

Carver (1972, p. 168) extended the usage of the term upper sand member of the Barnwell Formation into the Houston County area. However, he was of the opinion that in most cases, the upper sand was difficult or impossible to recognize in the field:

> The deep-red color and apparent lack of primary structure in most outcrops of the upper sand member apparently led Cooke and MacNeil (1952) to the conclusion that the unit was the residuum of a sandy limestone, an idea which has persisted for many years and has been reinforced by the supposition that either the "Cooper Marl" or Sandersville Limestone should occur at about that horizon.
>
> As a result of rapid iron-oxide cementation, sedimentary structures and even grain size, are difficult, or impossible to recognize in the upper 10 to 25 feet of most Barnwell Formation sections. As a result, the upper sand member has not been sufficiently well recognized to receive a formal name. Exposures at Section 2, however, firmly established its existence as coarse, gravelly sand in down-dip sections of the Barnwell Formation. It appears to be an extensive, regressive lithology of the Jackson Group, possibly a fluviatile sediment. (op. cit.)

BARNWELL FORMATION

The upper sand member of the Barnwell Formation as originally described by LaMoreaux (1946) is herein named the Tobacco Road Sand. One point, however, needs further clarification and that is the viability of the name Barnwell Formation. It is this formation that the "upper sand" has been a member of in the past.

The authors hesitate to use the name Barnwell until the lithostratigraphic unit is redefined for the following reasons. (1) The exposures in the railroad cut, 0.5 miles east of the station at Barnwell, Barnwell County, South Carolina, have been commonly considered to be a reference locality for the formation (Cooke , 1936, p. 91, 92). Cooke and MacNeil (1952) later concluded that the sediments in that exposure were of the Miocene, Hawthorne Formation. Doering (1961), with some uncertainty, mapped the same deposits as Citronelle Formation which he believed to be early Pleistocene in age. Connell (1968) again referred to this exposure as the type locality of the upper Eocene, Barnwell Formation. It is the opinion of the authors that the formation exposed in the railroad cut east of Barnwell is the same upper Tertiary formation that covers northern Screven County, Georgia, adjacent to Barnwell County, South Carolina. This upper Tertiary formation overlies the Tobacco Road Sand in southern Burke County. The authors are inclined to regard these sands, gravels, and clays as Hawthorne, Miocene age because field and core evidence in southern Screven County at this time suggests correlation with the Hawthorne Formation.

(2) Field examination by the authors and by William H. Abbott and Allan-Jon Zupan of the South Carolina Geological Survey have been unable to locate any useful exposures of correlatable upper Eocene sands in Barnwell County that can be established as a reference section for the Barnwell Formation. All exposures of deposits of probable late Eocene age in Barnwell County are either deep red, tough, almost hard, sand residuum or are so close to residuum[1] that practically all traces of primary sedimentary or biogenic structures have been eliminated, leaving only a few very vague traces in some exposures. As a result, there can be no certain correlation between residuum of proposed upper Eocene deposits in Barnwell County and the less severely weathered and more correlatable upper Eocene sand sections in Georgia or in Aiken County, South Carolina.

(3) Residuum of the "typical" Barnwell lithology can have no useful stratigraphic value because the deep red, residuum sands are associated with many Coastal Plain sand deposits of diverse ages and origins from Cretaceous to Pleistocene

Until the name Barnwell can be placed in a proper and useful stratigraphic context, and the stratigraphic terminology associated with the name is clarified, the name must be considered stratigraphically a *nomen dubium.*

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[1]In this paper saprolite is used to describe partially oxidized and leached sediments that still contain traces of primary and/or biogenic sedimentary structures. Residuum describes sediment that has been so thoroughly leached and oxidized that no primary and/or biogenic sedimentary structures have been preserved.

Section of Tobacco Road Sand exposed at type locality in Richmond County, Georgia

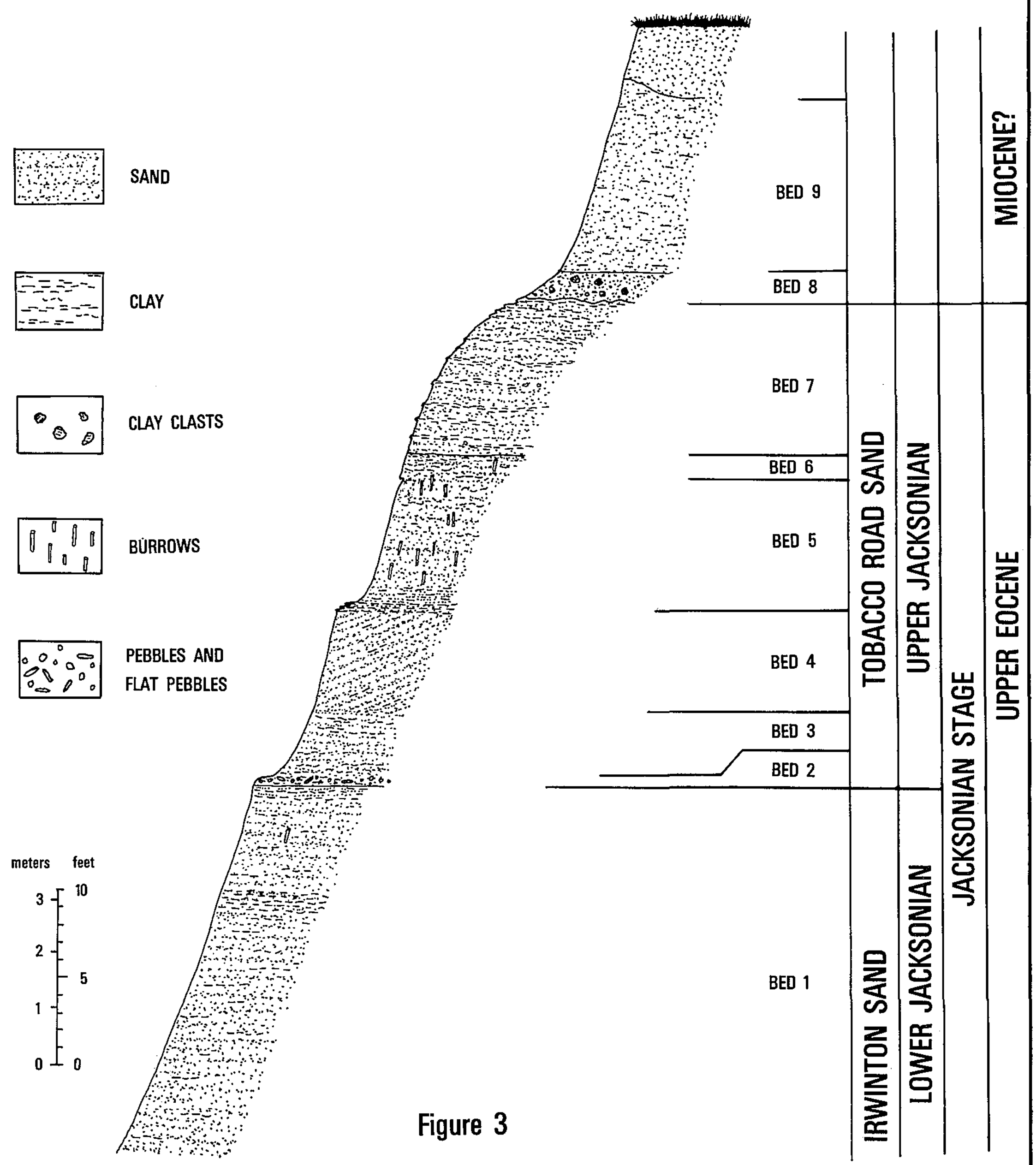


Figure 3

THE TOBACCO ROAD SAND

The type locality of the Tobacco Road Sand is located on the east side of Morgan Road, 0.35 miles north of the junction of Morgan Road and Tobacco Road, Richmond County, Georgia (Fig. 3). The following is a lithologic description of the type locality of the Tobacco Road Sand (see Fig. 4):

Formation	Bed	Description	Thickness
Neogene undifferentiated (Hawthorne?)	Bed 10	Sand (surficial sand of the soil zone), medium-grained, loose, pale tan color.	4.5 feet (1.4 m)
	Bed 9	Residuum, fine sand, argillaceous; massive; soft but tough when moist, hard and brittle when dry, resistant; deep-red color.	10 feet (3 m)
	Bed 8	Sand, very poorly sorted, medium- to coarse- to pebbly, quartz pebbles up to 2.5 cm in size, rounded to subangular, some pebbles crumble under pressure; clayey, with irregular-shaped clay clasts up to 3 cm; massive-bedded, no apparent sedimentary structures; very abrupt and irregular basal contact.	1.5 feet (0.5 m)
Tobacco Road Sand	Bed 7	Sand, fine to medium-grained, mainly medium-grained and moderately poorly sorted in lower 1 or 2 feet (0.3-0.6 m), rest of sand in bed 7 is fine-grained and well sorted; bedding very distinct and undulatory, some small scale crossbedding in the sand layers; conspicuous thin clay layers range in thickness from 0.5 mm to 1 cm, a few clay clasts present, some clay layers disrupted and distorted, more closely spaced clay layers in upper 2 feet (0.6 m); where exposed sand is fresh, it is soft, slightly coherent, and pale yellow to cream in color, where the sand is weathered and case-hardened, it is tough, resistant, and deep-red in color; lower contact of bed slightly undulatory and very abrupt, varying from a knife-edge contact to gradational over approximately 2 mm.	9.0 feet (2.75 m)
	Bed 6	Sand, medium-grained, somewhat argillaceous, micaceous, scattered quartz granules and thin clay chips; bedding varies from being distinct to being indistinct and disrupted, in some places the bed is bioturbated but the sediments are incompletely mixed, in other places bioturbation is not evident and the bedding is conspicuous, each layer varying from a few milimeters to less than a centimeter in thickness; moderately well sorted; soft and coherent when moist and freshly exposed, weathers to a tough, resistant ledge former; a few scattered burrows present; weathers gray to brownish-yellow; grades downward into:	1.5 feet (0.5 m)
	Bed 5	Sand, medium- to coarse grained, some very coarse sand grains and quartz granules present, micaceous, discontinuous clay laminae and and clayey sand layers, a few clay clasts present; layering present but indistinct and subtle, clay laminae slightly undulatory; scattered *Callianassa* burrows present throughout; base of bed consists of 6 inches (15 cm) of finely interlayered sand and clay; sorting moderate to poor: grades downward into:	7.5 feet (2.3 m)
	Bed 4	Sand, medium-grained, some coarse quartz sand grains scattered throughout, some heavy minerals present, very slightly argillaceous, quartz grains subrounded to rounded and frosted; fairly well sorted; very soft and incoherent where fresh, slightly coherent on weathered surface; very well bedded, consists of gently dipping (to the southeast) crossbed sets, individual layers vary from a few milimeters to 2 cm in thickness.	6 feet (1.8 m)

Formation	Bed	Description	Thickness
	Bed 3	Sand, fine-grained, slightly argillaceous, micaceous, heavy minerals common; scattered clay laminae and thin layers up to 2 millimeters thick, clay layers become closer spaced toward top of bed; sand well sorted; soft but slightly coherent.	3.5 feet (1 m)
	Bed 2	Gravelly sand with common flat quartz pebbles, pebbles well rounded to subrounded, up to 2.5 inches (6 cm), orientation of flat pebbles not always horizontal; sand medium- to coarse-grained scattered clay layers throughout bed, up to 1 cm in thickness; sorting is poor to moderate; gravelly sand soft and lacking coherence; clay confined to clay layers and not present as matrix of sand; lower contact sharp but conformable.	0.5 feet
Irwinton Sand	Bed 1	Sand, medium-grained, only very slightly coherent, very slightly argillaceous, discontinuous clay laminae and thin layers scattered throughout bed, becoming closer spaced toward top of bed, some scattered clay clasts, quartz granules and pebbles present but scattered throughout, pebbles up to 1 cm in size, a few clayey sand layers up to 5 cm thick are present; bedding subtle but distinctly horizontal, bedding surfaces are short, discontinuous; moderately well sorted.	21^{+} feet (6.4^{+} m)
			64.5 feet (19.8 m)
		Total Tobacco Road Sand thickness	**28 feet (8.5 m)**

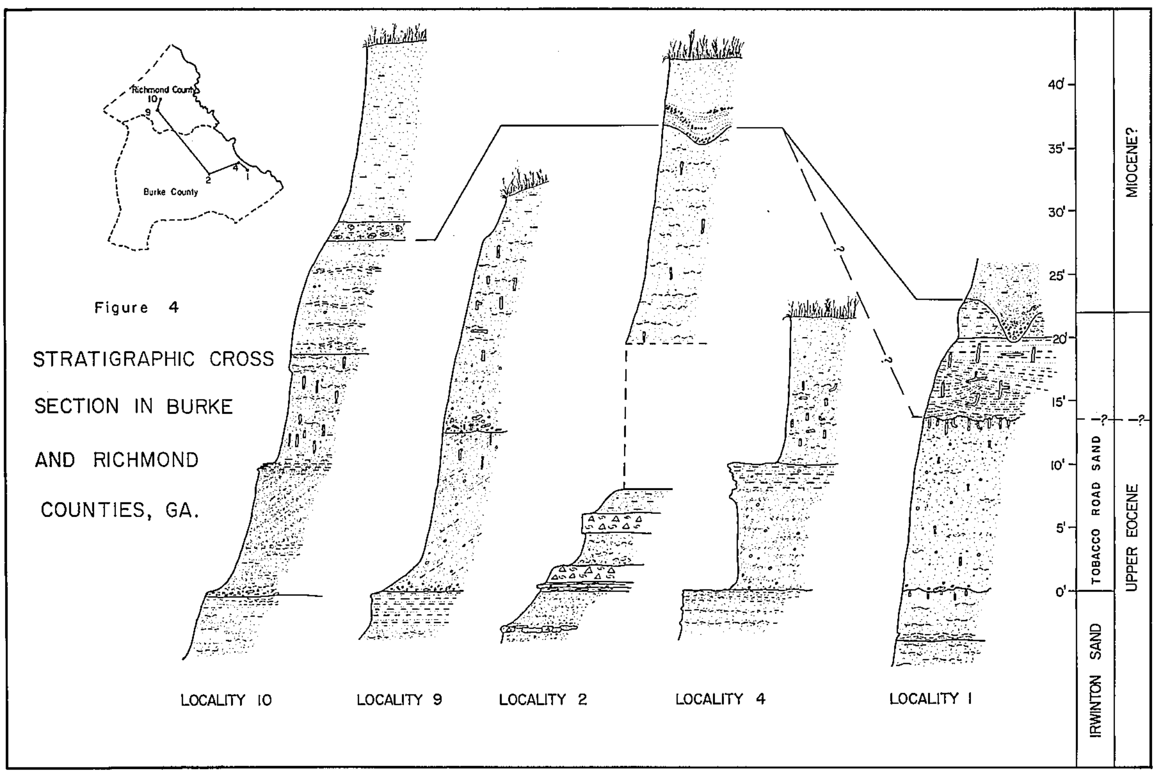


Figure 4

STRATIGRAPHIC CROSS SECTION IN BURKE AND RICHMOND COUNTIES, GA.

LITHOLOGY

The Tobacco Road Sand is dominately a medium to coarse quartz sand with subordinate amounts of clay minerals, mica, chert, calcite, and rarely glauconite. Iron and very rarely, manganese oxides are present in the weathered and leached sections. The Tobacco Road Sand contains only a small amount of clay occurring as the matrix of poorly sorted sands and in discontinuous clayey laminae and thin layers. Only locally does chert and limestone (Sandersville Limestone) form a significant proportion of the Tobacco Road section.

There are several differing facies within the Tobacco Road Sand that are related to the proximity of the ancient shoreline. The facies that was closest to the shoreline is now exposed in far updip areas in the vicinity of the Fall Line. For convenience in this discussion, this is called the nearshore facies. Another facies is typically exposed in a farther offshore direction, away from the Fall Line and to the south. For convenience this facies is referred to as the offshore facies.

The most characteristic and widespread of the two facies of the Tobacco Road Sand is the offshore facies and is typically exposed in Burke, Washington, northern Laurens, and southern Wilkinson Counties. This offshore facies is not exposed at the type locality of the formation near the Fall Line (see Fig. 5). Typically, the offshore facies of the Tobacco Road Sand is a massive bedded, bioturbated, incompletely mixed, fine to medium, or medium to coarse sand with subordinate clay and mica. The clay component of the offshore facies is unevenly distributed in the sand and occurs as wispy concentrations or as very wavy, disrupted laminae or thin layers. The almost invariable saprolite condition of the Tobacco Road Sand enhances the color contrast between the red sands and the light colored kaolinized clay laminae and swirls, thereby imparting to the bioturbated sands a distinctive appearance.

In all exposures of the typical sandy phase of the Tobacco Road Sand examined by the authors, the formation is invariably oxidized and leached to some degree. Fortunately, exposures are not rare where it is only mildly oxidized and leached. Only two exposures are known, however, where unweathered sediments of the offshore, bioturbated sand facies of the Tobacco Road are still intact, and these occur only as lenses within the mass of oxidized sand (Localities 12 and 14). At both of these localities the unweathered Tobacco Road Sand consists of bioturbated, noncalcareous, nonfossiliferous, massive-bedded, argillaceous, glauconitic, greenish-gray, medium-grained quartz sand. The base of the Tobacco Road Sand at Locality 12 consists of an unconsolidated, argillaceous, sandy limestone that contains a sparse suite of foraminifera and ostrocods. On the basis of these observations, it can be concluded with some degree of confidence that the offshore facies of the Tobacco Road Sand, before it was oxidized and leached, consisted mainly of nonfossiliferous, massive-bedded, bioturbated, slightly argillaceous, glauconitic, slightly micaceous, greenish-gray sand. The basal bed of the formation was sporadically calcereous and fossiliferous. In some areas the calcareous beds (now chert) appear to be interlayered with sand beds throughout the entire Tobacco Road section. Commonly the calcerous beds were limestone or sandy limestone.

Closely associated with the offshore, bioturbated sand facies are beds of fossiliferous or nonfossiliferous chert (replaced from limestone) that most commonly occurs near the base of the formation but also may occur as beds scattered throughout the formation. The chert varies from being relatively pure to being sandy, or as occurring as matrix between sand grains, forming chert cemented sandstone (for example, Locality 6). The chert or cherty sand may be massive and completely devoid of primary sedimentary structures, or may be thin bedded, thick bedded, or rarely cross-bedded.

The occurrence of the chert in the offshore facies is patchy. It is most likely to be encountered in Burke and northern Jenkins Counties and in Washington and northern Laurens Counties. It is very rare or absent in Twiggs, Wilkinson, and most of Jefferson Counties (except in the Ogeechee River area). In the areas where it is most commonly encountered in outcrop, chert beds may disappear within distances of a few tenths of a mile. Based on the spotty distribution of the chert, it appears that the chert may account for less than one percent of the total areal distribution of the Tobacco Road Sand. Present data is insufficient to indicate any precise pattern in the distribution of the chert, and it is not possible at this time to predict the precise lithologic nature of the chert at any site or in any area except it is generally sandy or occurs in a sandstone in those exposures nearest the Fall Line.

In the Sandersville area of Washington County, Georgia, the Sandersville Limestone represents the unaltered limestone of the offshore facies of the Tobacco Road Sand and consequently is recognized here as a member of the Tobacco Road Sand. The Sandersville Limestone may not have been silicified because of its compact, impermeable nature, or because of the impermeability of the overlying upper Tertiary clays and sandy clays.

During late Neogene time, incision of the drainage basins to elevations below the base of the soft and permeable Tobacco Road Sand exposed the formation to vadose conditions and, consequently, to oxidation and chemical leaching. As a result of this weathering, the patchy calcareous beds were replaced by chert and the glauconite in the sand was oxidized and leached, being replaced by hydrated ferric oxides, resulting in the deep-red color and sporadic chert that is typical of the severely leached Tobacco Road Sand.

The nearshore facies of the Tobacco Road Sand is exposed in a narrow band near to and parallel with the Fall Line. In general, there are more discreet thin to thick beds in this facies and the contrast in sand lithologies between various beds within the formation is more pronounced than it is in the offshore facies, as at the type locality. In some beds the sands are very poorly sorted, pebbly, devoid of primary sedimentary structures, and massive-bedded. In other beds the sands may be very well sorted, fine to coarse grained with distinct, small to large scale, horizontal-bedded or cross-bedded layers. The amplitude of the cross-bed sets ranges from a few inches (or centimeters) to as much as

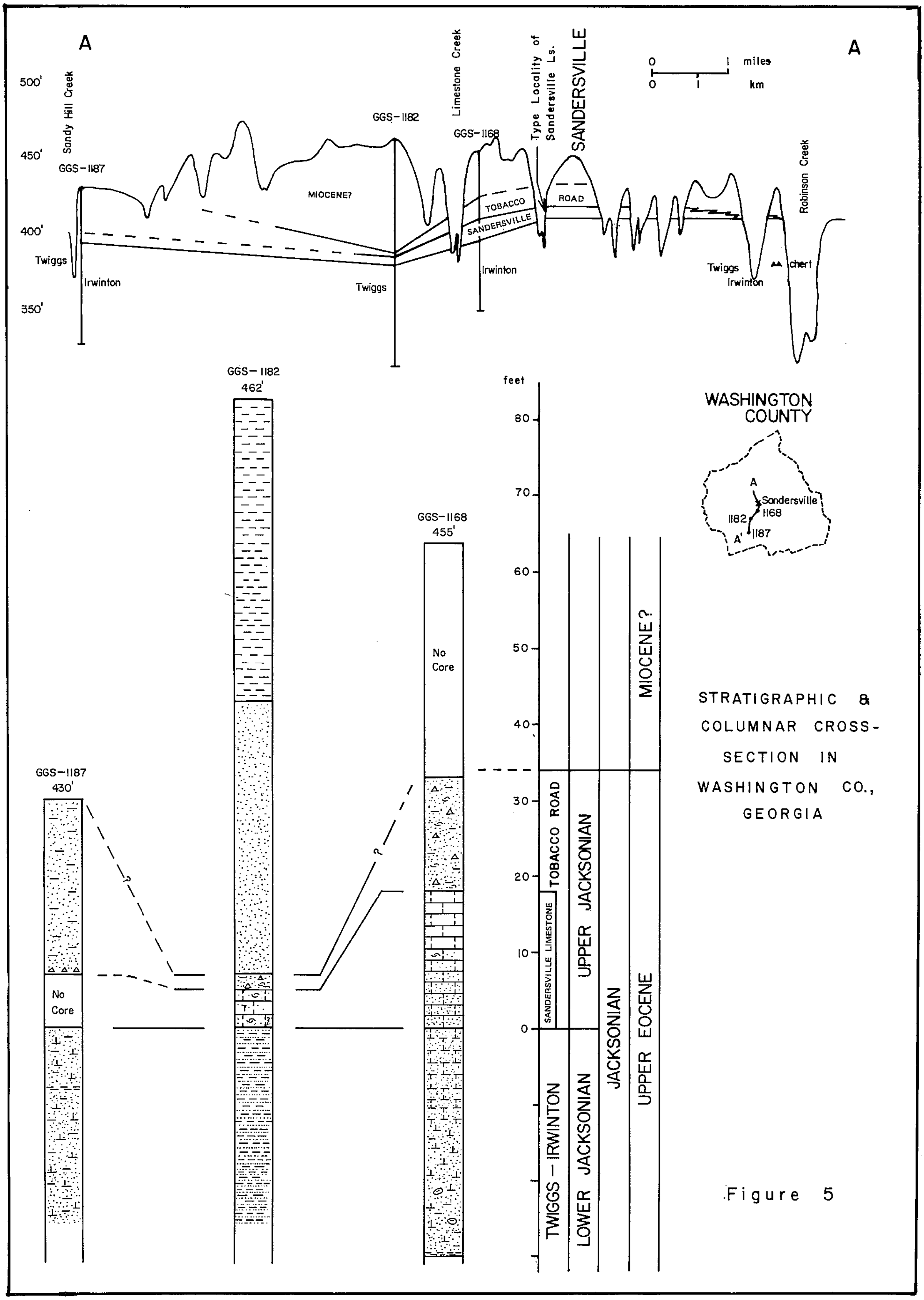

A
A
500'
450'
400'
350'
Sandy Hill Creek
GGS-1187
Twiggs
Irwinton
GGS-1182
MIOCENE?
Twiggs
Limestone Creek
GGS-1168
TOBACCO
SANDERSVILLE
Irwinton
Type Locality of Sandersville Ls.
SANDERSVILLE
ROAD
0 1 miles
0 1 km
Robinson Creek
Twiggs
Irwinton
chert
GGS-1182
462'
GGS-1168
455'
No Core
GGS-1187
430'
No Core
feet
80
70
60
50
40
30
20
10
0
TOBACCO ROAD
SANDERSVILLE LIMESTONE
TWIGGS – IRWINTON
UPPER JACKSONIAN
LOWER JACKSONIAN
JACKSONIAN
MIOCENE?
UPPER EOCENE
WASHINGTON COUNTY
A
Sandersville
1168
1182
A'
1187
STRATIGRAPHIC & COLUMNAR CROSS-SECTION IN WASHINGTON CO., GEORGIA

Figure 5

6 feet (1.8 m). At the other extreme, beds may be very thinly bedded and the fine sands and clays may be laminated. Beds of bioturbated sand occur rarely in any section of the nearshore facies. Evidence of biologic activity is not common in the nearshore facies and in the vicinity of the Fall Line even the occurrence of burrows, except in Richmond County, is rare.

The basal bed of the Tobacco Road Sand in the nearshore facies commonly consists of a coarse, poorly sorted, pebbly sand that contains flattened beach pebbles that may be as much as 2 inches (5 cm) in diameter. Locally, in the vicinity of the Fall Line, clay pebbles may be found in basal Tobacco Road sediments.

The basal poorly sorted, pebbly sand and the basal cherts of the Tobacco Road are almost mutually exclusive, i.e., they very rarely occur together in the same outcrop (only one instance is known) and no intergradation is known between the two basal lithologies (no chert or silicified shells are known to occur in the poorly sorted, flat pebble bed, and no flat pebbles are known to occur in the basal cherts). Essentially then, the poorly sorted, flat pebble bed is characteristic of and restricted to the nearshore facies of the Tobacco Road Sand whereas the basal cherts are characteristic of the offshore facies but also occur rarely as chert cemented sandstone in the nearshore facies.

It is postulated in this study that the Tobacco Road Sand represents an open sound or lagoon facies of the uppermost Eocene deposits in Georgia and South Carolina. If this is correct, then the latest Eocene sound must have been separated from the open ocean by barrier islands or at least some obstruction to free watermass interchange. Remnants of possible barrier-like islands are exposed along Ga. 247 in northern Pulaski and southern Houston Counties, Georgia, where very clean, loose, well-sorted sands are exposed in scattered road cuts. Certainly this type of sand is an unusual lithologic component of the Tobacco Road Sand, especially in the offshore facies. The geographic location of these sands suggest a barrier island-like origin. "Cooper Marl" is found immediately to the south (seaward) and only Tobacco Road Sand is found to the north (landward).

TOBACCO ROAD - IRWINTON/TWIGGS BOUNDARY RELATIONSHIPS

The base of the Tobacco Road Sand commonly makes a very distinctive and traceable interval across the outcrop belt. As was pointed out by LaMoreaux (1946b, p. 11), the top of the Irwinton Sand in Twiggs, Wilkinson, and Washington Counties commonly contains a clay bed. This clay bed may be as much as 6 feet (1.8 m) thick and is a useful marker for locating the contact between the Tobacco Road Sand and Irwinton Sand in this area. In places this clay bed thins to a 6 inch (15 cm) thick bed composed of alternating thin laminae of gray clay and fine to coarse grained sand. To the east of Washington County the clay bed becomes increasingly more discontinuous in its distribution and the basal sand of the Tobacco Road commonly lies directly on sands of the Irwinton.

In the nearshore facies of the Tobacco Road, the basal bed of the formation differs from the underlying Irwinton Sand in being composed of coarse, poorly sorted sand that may be pebbly and contains flattened beach pebbles (Localities 10 and 11). Where the marker clay bed is not present, the sands of the underlying Irwinton are well sorted, commonly fine to medium grained but may be coarser near the Fall Line. The Irwinton Sands are soft and almost incoherent, and vaguely to conspicuously bedded (especially as exposed by the action of running water in gullies and steepheads).

In the Sandersville area of Washington County, cores taken by the Georgia Geological Survey (GGS 1168, 1182, 1187) show other aspects of the Tobacco Road/Irwinton contact (Fig. 6). The upper part of the Irwinton Sand in the GGS 1168 progressively becomes more calcareous and less sandy upward in the section. Only in the upper 10 feet (3 m) of the calcareous section in the GGS 1168 core is typical Sandersville Limestone lithology present. Elsewhere in Washington County (GGS 1182), the Sandersville Limestone is represented by only 5 feet (1.5 m) of limestone that abruptly overlies very sandy Twiggs Clay and in turn is overlain by 3 feet (0.9 m) of cherty Tobacco Road Sand. Finally, in the GGS 1187 core, no Sandersville Limestone is present, possibly due to poor core recovery. The basal Tobacco Road Sand is cherty and it overlies calcareous Irwinton Sand.

Typical Sandersville Limestone is present only locally in Washington County and where it is present, it consistently occurs between the underlying Irwinton Sand or Twiggs Clay and the overlying Tobacco Road Sand, or it may occur in place of the sand lithology of the Tobacco Road. In some places there is a sharp contact between the Sandersville Limestone and the underlying Irwinton Sand or Twiggs Clay (Figs. 5 and 6). In other places there is no contact as such, instead a gradual change in lithology exists.

Two localities in Burke County present a similar situation to that present in Washington County. At Hatchers Mill on Brier Creek (Locality 2, Fig. 5) and at Newberry Creek (Locality 3), the lowest cherty layers are sandy and are interlayered with Irwinton Sand. At these two localities the thickest chert layers occur in the upper part of the chert-bearing intervals and directly underlie or are interlayered with the Tobacco Road Sand. At Hatchers Mill the upper, thicker cherts contain the index fossil *Periarchus quinquefarius* which is found only in the Tobacco Sand and its facies equivalents the Sandersville Limestone, and the "Cooper Marl". The basal chert beds of the Tobacco Road Sand and the upper sandy chert beds of the Irwinton are a part of the same general facies as the Sandersville Limestone. The main difference is that the other carbonate beds have been silicified and the Sandersville Limestone has not. Due to the sporadic occurrence of the chert and its intimate interlayering with typical Tobacco Road sediments, the authors prefer to regard the cherts as a particular lithologic phase of the Tobacco Road Sand, thereby restricting the use of the name Sandersville Limestone to the limestone beds in the immediate Sandersville area.

LITHOLOGIC DISTINCTION BETWEEN TOBACCO ROAD SAND AND UNDERLYING UNITS

The Tobacco Road Sand is consistently coarser grained and is generally more poorly sorted than the underlying

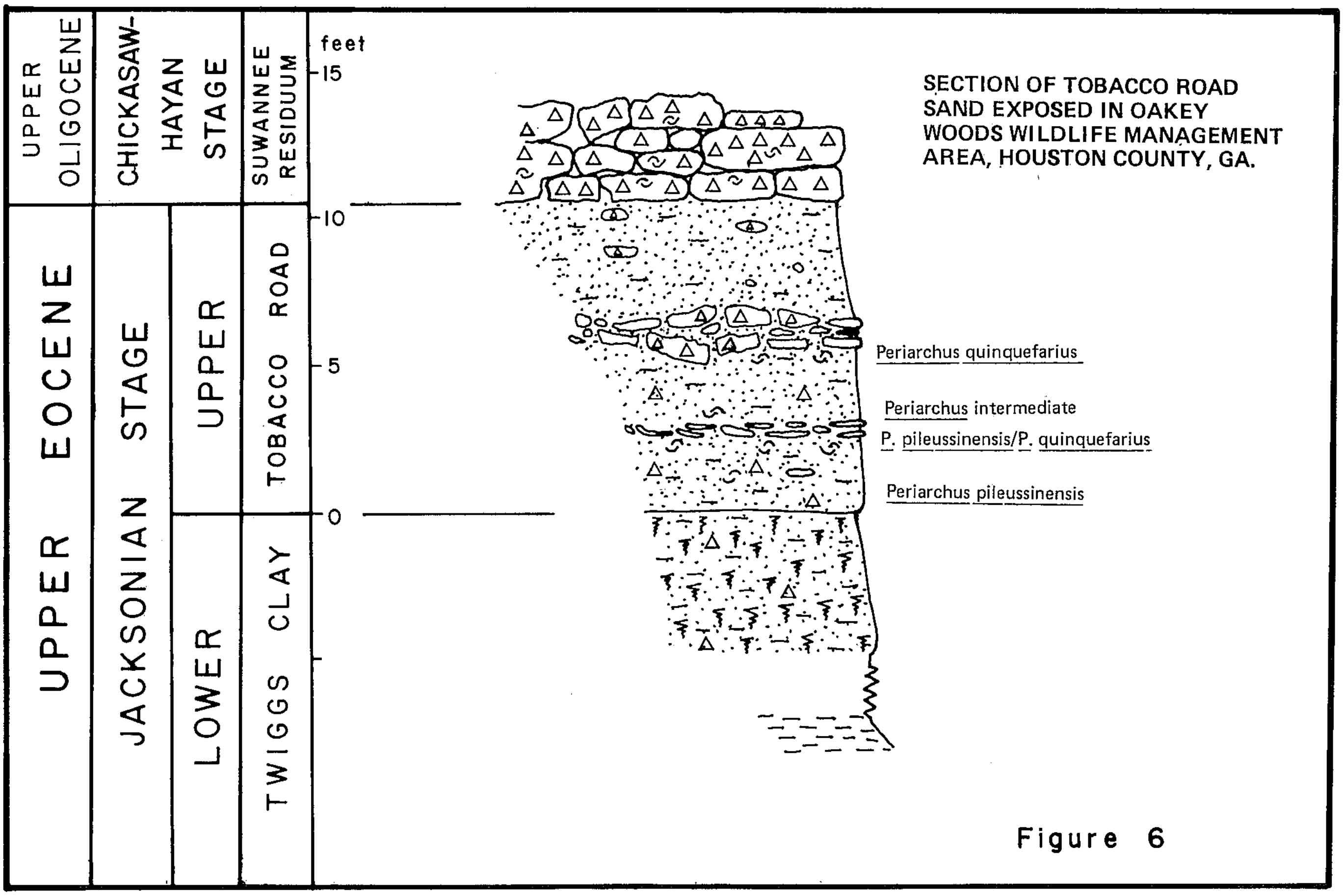


Figure 6

Irwinton Sand. Only in the vicinity of the Fall Line have coarse-grained, rather poorly sorted sand beds been observed in the Irwinton Sand that are similar lithologically to the Tobacco Road Sand (Locality 11). These beds in the Irwinton Sand contain only small (commonly 0.5 inch or 1 cm), subrounded pebbles, not the large (up to 2 inches or 5 cm), well-rounded pebbles and flat pebbles found in the Tobacco Road Sand.

In the Irwinton Sand bioturbation does exist locally and in thin beds. In the Tobacco Road Sand; however, bioturbation commonly occurs on a massive scale, involving much or all of the unit in any one exposure.

The Irwinton Sand is consistently a well bedded sand that may be very thinly and intricately horizontally bedded, or less commonly, crossbedded. Except in the vicinity of the Fall Line, crossbedding with amplitudes greater than one or two feet is very rare in the Irwinton. In contrast, the Tobacco Road Sand is more commonly very thick bedded or even massive, the original primary bedding structures having been obliterated by subsequent bioturbation and mixing of the sediments. Crossbed sets are known to reach at least six feet (1.8 m) in the Tobacco Road Sand.

In general, the earlier Jacksonian deposits (Clinchfield Sand, Twiggs Clay, and Irwinton Sand) contain a more prominent clay component than does the Tobacco Road Sand. In the earliest Jacksonian Clinchfield Sand and in the younger Irwinton Sand, there are laminae, layers, thin beds, thick beds, and lenses of relatively pure clay. The Twiggs Clay, a relatively pure clay, makes up most of the terrigenous clastic lower Jacksonian deposits in the downdip areas of most of eastern Georgia.

LITHOLOGIC DISTINCTION OF TOBACCO ROAD SAND WITH OVERLYING UNITS

The Tobacco Road Sand is overlain by several different units in its outcrop belt. In southeastern Houston, northern Pulaski, and northern Bleckley Counties, the Tobacco Road is overlain by chert and clay residuum of late Oligocene age. Also in northern Pulaski and Bleckley Counties, and in southernmost Burke County the Tobacco Road Sand is overlain by an unnamed, thinly bedded, fine sand and clay unit that may be a nearshore facies of the upper Oligocene Suwannee Limestone. There are other scattered occurrences in southern Twiggs County and in Wilkinson County of heavily saprolitized, thin-bedded sands and clay that may represent the same unnamed unit.

In southern Twiggs and Wilkinson Counties, in northern Laurens County, southern Washington County, southern Richmond County, and in Burke County, the Tobacco Road Sand is overlain disconformably in places by crossbedded feldspathic sands and gravels, and poorly sorted clayey, pebbly sands that are probably a marginal marine to lower delta floodplain facies of the Hawthorne Formation. These deposits are typically even more poorly sorted than the Tobacco Road Sand. Locally these clayey sands are cross-bedded or are indurated into a poorly sorted sandstone or claystone (Altamaha grit of Veatch and Stephenson, 1911).

In some areas near the Fall Line, and in some places near major streams, the Tobacco Road Sand appears to be overlain disconformably by undifferentiated crossbedded sands and gravels that are probably Plio-Pleistocene high river terrace deposits. Such deposits can be seen in a sand pit just above the floodplain of Brier Creek near Ga. 23 between Girard and Sardis, Burke County. There are several exposures near Dublin where similar channels of crossbedded gravels cut down into the Tobacco Road Sand (e.g., Locality 7). In the Midville area of Burke County, the Tobacco Road Sand is missing in GGS-2136 and high terrace sands and clays lie directly on the Irwinton Sand. In the Savannah River area in Aiken County, South Carolina, the Tobacco Road Sand is overlain by an ironstone cemented conglomerate that is reminiscent of the high river terrace gravels in Alabama and along the Chattahoochee River in Georgia.

These high terrace deposits are much coarser grained than would be expected for any beds or facies of the Tobacco Road Sand in the same area relative to the Fall Line.

FACIES RELATIONSHIPS OF THE TOBACCO ROAD SAND WITH ADJACENT UNITS

Unlike the underlying Irwinton Sand, the Tobacco Road Sand does not grade downdip or seaward, by gradual facies change, into thick clay deposits; and farther downdip into carbonates. The quartz sands of the Tobacco Road appear to pass directly, with only minor, intermediate clay beds, into the bioclastic, microfossiliferous, calcarenitic "Cooper Marl". The facies change is by intergrading or interfingering, probably through an intermediate barrier island-like sand.

AREAL DISTRIBUTION

The westernmost, currently recognized exposures of the Tobacco Road Sand are in southeastern Houston County at Localities 5 and 8. The easternmost known exposures are in Aiken County, South Carolina. Tobacco Road Sand residuum is probably present in Barnwell County, South Carolina, because the pre-Neogene surficial residuum in Barnwell County is very similar to the Tobacco Road residuum in Burke County, Georgia.

To the north the Tobacco Road Sand occurs at high elevations in the vicinity of the Fall Line (e.g., Locality 11). Its original northern limit probably extended some small but undeterminable distance north of the present Fall Line.

To the south, in a downdip or seaward direction, the Tobacco Road Sand passes into the "Cooper Marl" of Georgia by facies change.

The length of the existing belt of Tobacco Road sediments from southeastern Houston County, Georgia, to Aiken County, South Carolina, is about 130 miles (209 km). In the east, across the strike of the belt of Tobacco Road Sand, the formation is approximately 40 miles (64 km) wide from Augusta, Richmond County to southern Burke County, Georgia. In the west, the belt of Tobacco Road Sand is approximately 30 miles (48 km) wide from Jones to Pulaski Counties, Georgia. The areal extent of the Tobacco Road Sand is, therefore, roughtly 4500 square miles (12,000 km^2).

The original western limit of the Tobacco Road Sand is unknown due to epeirogenic uplift in the Mississippi Embayment and its eastern margins in Georgia. However, it is possible that the formation originally extended as far west as Alabama. Because of the regional, epeirogenic uplift in western Georgia and Alabama, the structural strike of the upper Eocene outcrop belt in central Georgia begins to diverge from the depositional strike. The structural strike swings to the southwest in the vicinity of the Ocmulgee River and as a result, the outcrop belt crosses facies boundaries that are progressively farther offshore. Therefore the Tobacco Road Sand grades laterally in a westward direction into "Cooper Marl", and the "Cooper Marl" grades laterally, to the southwest into Ocala Limestone.

Similarly, because of the uplift along the Cape Fear Arch in North Carolina, the structural strike of the Tobacco Road begins to diverge from the depositional strike in South Carolina. As a result, the outcrop belt swings eastward and crosses facies boundaries that are progressively farther offshore. The Tobacco Road Sand in South Carolina passes eastward, laterally into the basal part of the Cooper Marl between the Savannah and Santee Rivers in South Carolina.

THICKNESS AND GEOMETRY

Where reasonably complete sections of the Tobacco Road Sand are exposed, it averages between 15 (4.5 m) and 20 (6.1 m) feet thick. The greatest known thickness of the Tobacco Road Sand is present at the Albion Kaolin mine at Hephzibah in Richmond County, Georgia (Locality 10), where 31 feet (9.5 m) is exposed. The least total thickness that is currently known is 10.5 feet (3.1 m) at Locality 5 in Houston County (Fig. 7).

In general, the Tobacco Road Sand is a nearshore blanket that ranges in thickness from approximately 10 feet (3 m) to 28 feet (8.5 m). The original volume of Tobacco Road sediments in its area of known occurrence is approximately 17 cubic miles (71 km^3).

PHYSIOGRAPHIC EXPRESSION

The Tobacco Road Sand is a relatively tough, resistant unit of which vertical faces are commonly exposed, even where mildly oxidized. At places where it is thoroughly leached and case-hardened with iron oxide cement the formation is very tough and brick-like. There is commonly at least 10 feet (3 m) of this very resistant residuum where the soil profile is developed on the Tobacco Road Sand.

The underlying Irwinton Sand, Twiggs Clay, and, at lower elevations, kaolin and kaolinitic sands of early Tertiary and Late Cretaceous age are very susceptible to physical erosion. Where these units are thick and the Tobacco Road Sand lies high above the valley bottoms, as in Twiggs and Wilkinson Counties, stream dissection is deep and there is relatively great relief for the Coastal Plain (Fall

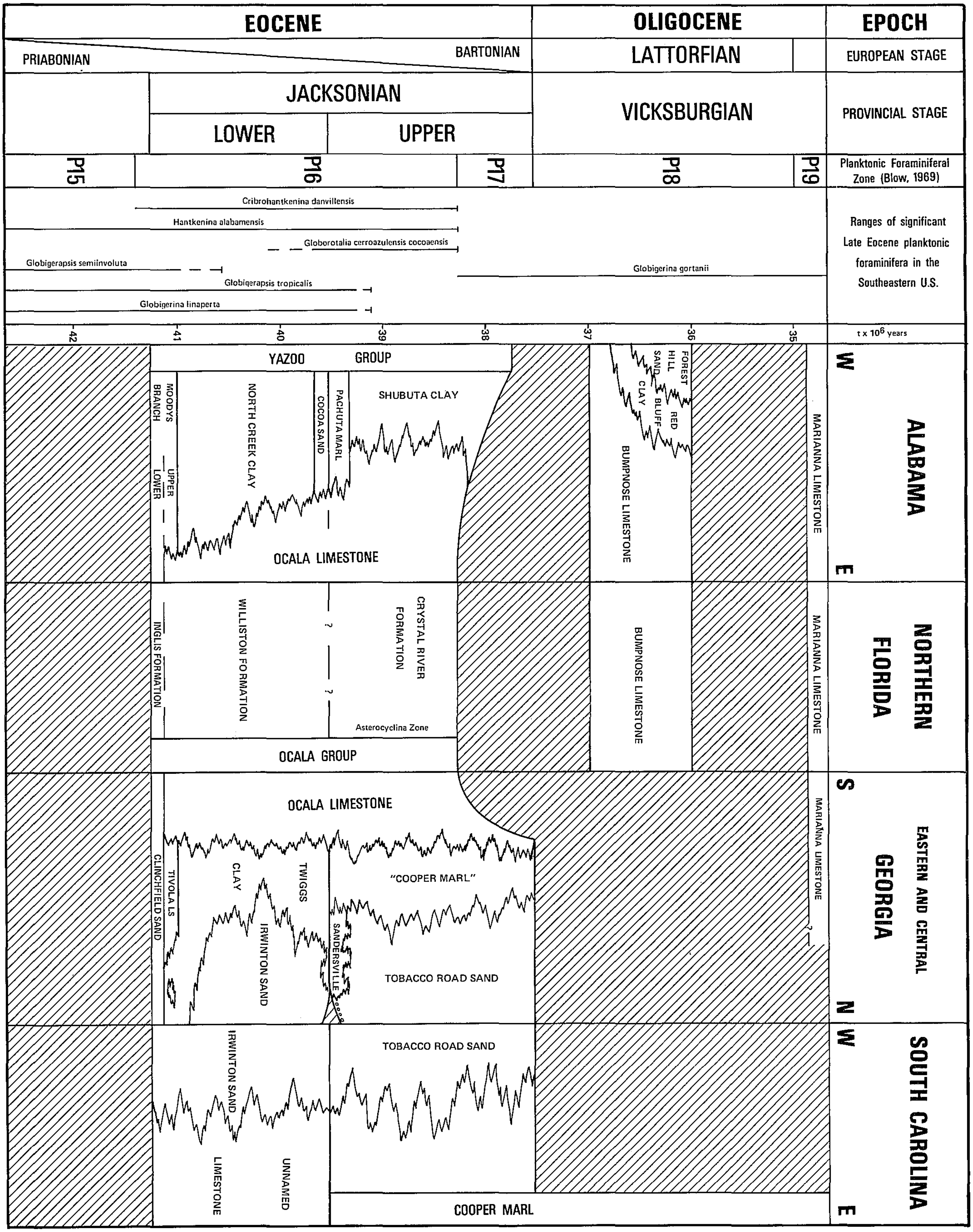

EOCENE
OLIGOCENE
EPOCH
PRIABONIAN
BARTONIAN
LATTORFIAN
EUROPEAN STAGE
JACKSONIAN
LOWER
UPPER
VICKSBURGIAN
PROVINCIAL STAGE
P15
P16
P17
P18
P19
Planktonic Foraminiferal Zone (Blow, 1969)
Cribrohantkenina danvillensis
Hantkenina alabamensis
Globorotalia cerroazulensis cocoaensis
Globigerapsis semiinvoluta
Globigerina gortanii
Globigerapsis tropicalis
Globigerina linaperta
Ranges of significant Late Eocene planktonic foraminifera in the Southeastern U.S.
42
41
40
39
38
37
36
35
t x 10^6 years
YAZOO GROUP
MOODYS BRANCH
UPPER
LOWER
NORTH CREEK CLAY
COCOA SAND
PACHUTA MARL
SHUBUTA CLAY
OCALA LIMESTONE
FOREST HILL SAND
RED BLUFF CLAY
BUMPNOSE LIMESTONE
MARIANNA LIMESTONE
W
E
ALABAMA
INGLIS FORMATION
WILLISTON FORMATION
CRYSTAL RIVER FORMATION
Asterocyclina Zone
OCALA GROUP
BUMPNOSE LIMESTONE
MARIANNA LIMESTONE
NORTHERN FLORIDA
S
N
OCALA LIMESTONE
CLINCHFIELD SAND
TIVOLA LS
CLAY
TWIGGS
IRWINTON SAND
SANDERSVILLE
"COOPER MARL"
TOBACCO ROAD SAND
MARIANNA LIMESTONE
EASTERN AND CENTRAL GEORGIA
W
E
TOBACCO ROAD SAND
IRWINTON SAND
LIMESTONE
UNNAMED
COOPER MARL
SOUTH CAROLINA

Figure 7

Line Hills). The large-scale removal of these less competent underlying deposits causes massive undercutting of the resistant Tobacco Road Sand. As a result, the Tobacco Road Sand has been removed from much of these areas, leaving only remnants capping high ridges and outliers which may extend almost to the Fall Line. These ridges commonly are sites of main highways such as the high ridge that US 80 follows east of Macon or the ridge that GA 57 follows between Gordon and Irwinton. These ridges capped by the Tobacco Road are also the sites of some larger communities such as Gordon, Jeffersonville, and Irwinton. The Tobacco Road, for which the formation is named, also follows a ridge capped by Tobacco Road Sand.

Farther south, where the Tobacco Road Sand lies closer to the valley bottoms and is not so extensively undercut, as in Burke, southern Richmond, and much of Jefferson Counties, there are broad, rolling upland areas with widely separated, deeply dissected stream valleys (Louisville Plateau of Cooke, 1925). In this area, too, the gravels, clayey sands and sandy clays of the updip feather edge of the Neogene (Hawthorne) deposits directly overlie the Tobacco Road Sand. The Neogene deposits also make up a tough, resistant unit and have similar topographic expression, therefore adding to the volume of resistant deposits. Finally, where the soft Irwinton Sand or Twiggs Clay dips under the stream valley floors and the Neogene thickens to over 100 feet, the plateau topography gives way to the more closely and evenly spaced, dendritic drainage pattern of the Vidalia Upland (Clarke and Zisa, 1976).

BIOSTRATIGRAPHY

The only macro-fossils that have been identified from the Tobacco Road Sand are *Pariarchus pileussinensis*, which occurs only near the base of the unit at a few localities (Localities 5 and 12), *Periarchus quinquefarius*, and *Crassostrea gigantissima* (Sandersville Limestone). *Callianassa major* burrows are present in the Tobacco Road sediments at several sites near the Fall Line including the type locality.

Some of the cherts of the offshore facies of the Tobacco Road contain what appear to be rich suites of mollusks but these are generally preserved only as molds and impressions. In a few rare places the aragonitic shells of pelecypods and gastropods have been replaced by silica.

A scant assemblage of poorly preserved foraminifera and ostrocods have been observed from assorted samples from the Sandersville Limestone and from the basal calcareous bed at Locality 12 in Washington County.

In Georgia the Tobacco Road Sand grades downdip and seaward into the upper Eocene "Cooper Marl". The one macrofossil that is known to be restricted to these two units (including the Sandersville Limestone) is *Periarchus quinquefarius. Periarchus quinquefarius* is descended from *P. pileusseninsis*, a species present in lower Jacksonian, and in basal upper Jacksonian deposits. Based on observations by the authors at Localities 5 (Fig. 7) and 12, the evolutionary development appears to have taken place during the early part of the Tobacco Road time. This is consistent with the presence of only *P. pileussinensis* in the basal upper Jacksonian Pachuta Marl of Alabama and Mississippi (Cooke, 1959; Huddlestun, 1965; Toulmin, 1977).

The "Cooper Marl" in Georgia contains a P16 planktonic foraminiferal assemblage of Blow (1969) (see Fig. 8) based on the co-occurrence of the planktonic foraminifera *Cribrohantkenina danvillensis, Hantkenina alabamensis*, and *Globorotalia cerroazulensis.* Because these species of foraminifera became extinct within the "Cooper Marl", the upper part of the "Cooper" is probably contained within Zone P17. Both P16 and P17 are latest Eocene in age, and by extrapolation, the Tobacco Road Sand is also latest Eocene in age.

The Tobacco Road Sand is time-stratigraphically equivalent to the Ocala Limestone that crops out along the Flint River (Lake Blackshear) west of Cordele, Crisp County, Georgia. In this area the Ocala contains a P16 planktonic foraminiferal assemblage but lithologically is reminiscent of the "Cooper Marl". Likewise, the Tobacco Road Sand is equivalent to the *Asterocyclina* Zone of the Ocala Limestone (Crystal River Formation of Puri, 1957) in Jackson County, Florida; to the Ocala Limestone in Alabama (Huddlestun, 1965; Huddlestun and Toulmin, 1965; Toulmin, 1977); to the Pachuta Marl and Shubuta Clay in western Alabama and in Mississippi (Deboo, 1965); and to the lowest Cooper Marl of South Carolina. All these units contain a P16 planktonic foraminiferal assemblage, and some contain a P17 assemblage.

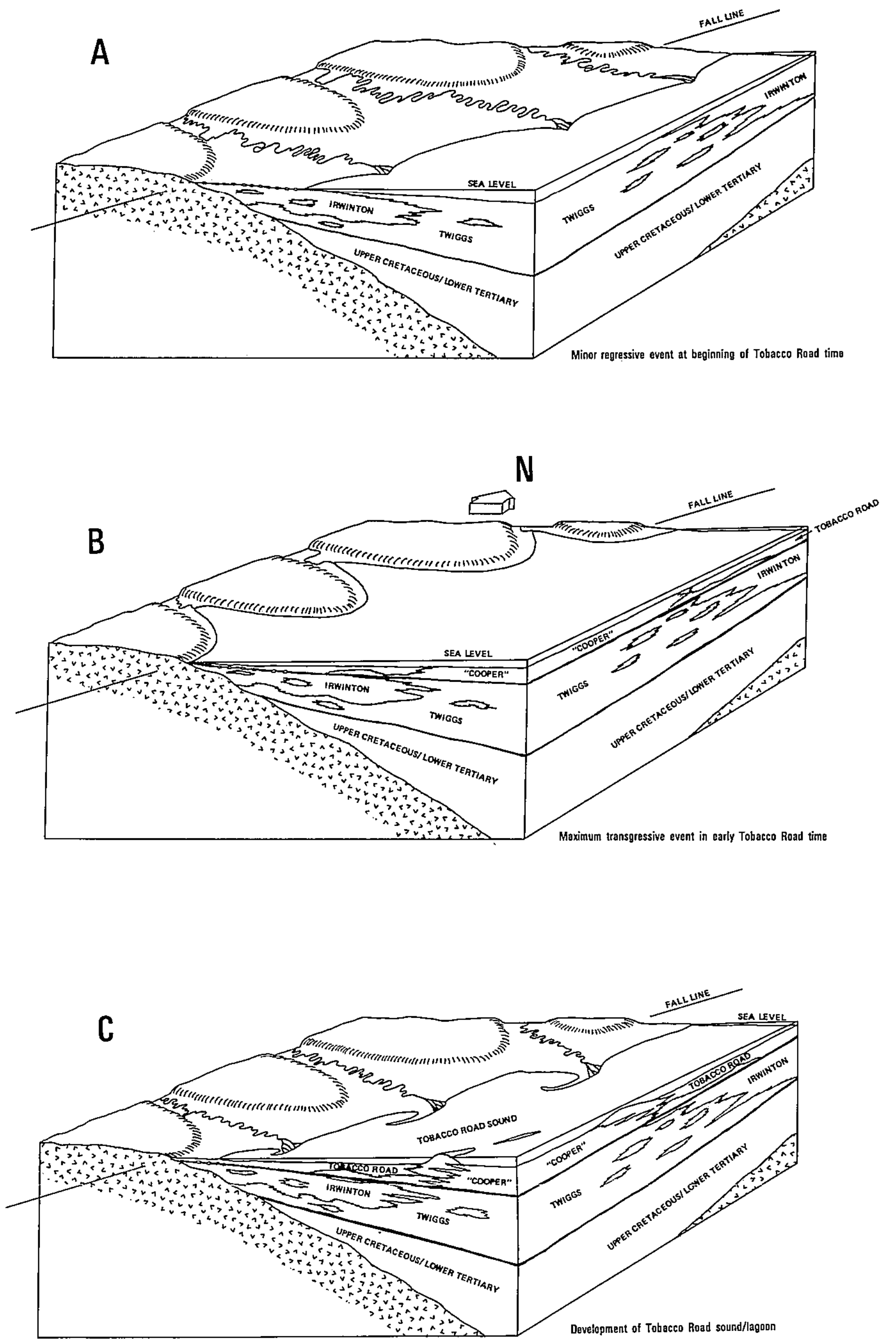
The Tobacco Road Sand is late Eocene in age and is in the Priabonian/Bartonian Stage of Europe, and the provincial Jacksonian Stage of the Coastal Plain of eastern North America.

RATE OF SEDIMENTATION

Examination of Figure 8 will show that the time duration for the Jacksonian of the southeastern United States is approximately 4 million years. This estimate for the time duration of Jacksonian deposition is based on the occurrence of an earliest P16 planktonic foraminiferal assemblage (concurrence of *Globigerapsis mexicana* and *Cribrohantkenina danvillensis*) (Fig. 8) in the basal Jacksonian Moodys Branch Formation at its type locality in Jackson, Mississippi, and a P17 planktonic foraminiferal assemblage in the upper part of the Shubuta Clay in Mississippi and Alabama, and in the "Cooper Marl" of Georgia.

The Tobacco Road Sand is restricted to the upper part of the Jacksonian. Although there is a change in the planktonic foraminiferal assemblages between the lower and upper Jacksonian in the southeastern United States, this change is not reflected in the established zonations (Bolli, 1957; Blow, 1969; Stainforth et al., 1975). Therefore for convenience of discussion, it is assumed that the lower/upper Jacksonian boundary is in the middle of the Jacksonian, and that the duration of the upper Jacksonian is 2 million years. This assumption, as will be observed, will not seriously influence the calculated sedimentation rates.

The thickest known section of the Tobacco Road Sand is 28 feet (8.5 m) at the type locality. If the 8.5 m of Tobacco Road Sand exposed at the type locality were deposited over a time interval of 2 million years, the average rate of sedimentation would be approximately 0.5 cm/1000



BLOCK DIAGRAM SHOWING INFERRED DEVELOPMENT OF TOBACCO ROAD SEDIMENTATION AND PALEOGEOGRAPHY

Figure 8

years. This is very comparable to the estimated sedimentation rates of blue clays, red clays, and foraminiferal oozes in the ocean basins (Svedrup, 1953). Inspection of Tobacco Road Sand exposures suggest that such slow sedimentation rates appear to be unrealistically low. We therefore conclude:

1. The value of 0.5 cm/1000 years is more representative of the average rate of sedimentation of the formation as a whole, and is indicative of the rate of relative subsidence of the coastal area.
2. Sedimentation at any given site was discontinuous and probably varied from nondepositional (over periods of years or centuries) to extremely rapid (e.g., sedimentation resulting from a violent storm).

ENVIRONMENTAL PARAMETERS

Evidence for marine origin. The following evidence supports a marine, nearshore, origin for the Tobacco Road Sand:

1. There are abundant examples of bioturbation of the sediment.
2. Nondescript burrows are of common occurrence whereas *Callianassa* burrows are more rare in occurrence.
3. *Crassostrea gigantissima* is present in the Sandersville Limestone, a limestone facies of the Tobacco Road Sand.
4. Mollusks are present locally within the cherts or cherty sands.
5. The echinoid genus *Periarchus* is present locally, both in the cherts and in the sands (Echinodermata is restricted to marine habitats).
6. Shark and ray teeth are present in Tobacco Road Sands in the Albion Mine, Hephzibah, Richmond County, Georgia.

The following evidence supports a coastal environment of variable marine conditions for the Tobacco Road Sand:

1. Larger foraminifera, typically abundant in upper Eocene deposits in Georgia (Ocala Limestone), are absent in the cherts of the Tobacco Road Sand and in the Sandersville Limestone.
2. Diverse echinoid faunas which reach their peak in numbers and diversity in the upper Eocene limestones in the Southeast, are absent in the Tobacco Road Sand and in the Sandersville Limestone.
3. Crossbed sets, ranging in amplitude from small scale to large scale (up to 6 feet (1.8 m) in amplitude), are locally present in the Tobacco Road Sand.
4. Characteristically the sands of the Tobacco Road are poorly sorted, generally with an argillaceous matrix and with small pebbles scattered throughout.
5. Flat beach pebbles interpreted as being of beach origin, occur commonly at the base in the nearshore facies of the Tobacco Road.

Evidence for sound/lagoon origin. Although the evidence supporting a sound/lagoon origin for the Tobacco Road Sand is not compelling, still there is some evidence to support this interpretation. The reasons for proposing this origin are as follows:

1. *Periarchus* is the most common identifiable fossil in the Tobacco Road. The senior author has observed that the various species of the genus *Periarchus* are associated with foraminiferal populations with relatively low diversity and high species dominance (Waller, 1964) in the Pachuta Marl, Moodys Branch Formation, Wenona Sand, Clinchfield Sand, and Tivola Limestone (also see Deboo, 1965; Bandy, 1949). *Periarchus* is associated with relatively rich molluskan faunas which are accepted as being of nearshore origin. *Periarchus* is <u>not</u> associated with abundant and diverse upper Eocene echinoid faunas; conversely when highly diverse and rich echinoid faunas or foraminiferal faunas are present, (Deboo, 1965; Bandy, 1949) *Periarchus* is either rare or absent. Finally, in the rich "Scutella" bed of Mississippi, Alabama (Moodys Branch Formation) and in Georgia (Clinchfield Sand) there are low concentrations of foraminifera and few species present.
2. *Crossostrea gigantissima* is present in the Sandersville Limestone. Invariably where it is present in Coastal Plain deposits (McBean Formation, Clinchfield Sand, Tobacco Road Sand, and Trent Marl) it is associated with a foraminiferal fauna with low diversity and high species dominance. This is consistent with the premise that *C. gigantissima* is the ancestor of the living *C. virginica* and that both species inhabited similar niches, prefering brackish waters.
3. Burrows of *Callianassa major (Ophiomorpha nodosa*; Weimer and Hoyt, 1964) are locally common in the nearshore facies of the Tobacco Road Sand. The early Tertiary *Callianassa* burrows are essentially identical to those that the present *Callianassa major* inhabit in the present coastal areas. These burrowing shrimp are especially abundant in the intertidal zone (Weimer, R. J., and Hoyt, J. H., 1964) along the Atlantic and Gulf coasts and are found also in very shallow water, subtidal areas, and in front of modern beaches. The occurrence of *Callianassa* burrows in near association with flat beach pebbles is consistent with at least part of nearshore facies of the Tobacco Road Sand being a beach deposit or an extremely shallow water deposit.
4. The long, narrow shape of the Tobacco Road deposit and the abrupt change in lithology and type of sedimentary structures between it and its downdip correlative ("Cooper Marl") are best explained by a physical barrier. The barrier (possibly barrier island) would presumably restrict exchange of marine and sound or lagoon waters. The belt of facies change from the Tobacco Road Sand to the "Cooper Marl" is relatively narrow and is best exemplified in outcrop by the exposures along Ga. 247 in the northwestern corner of Pulaski County, Georgia (Locality 8). There the red residual sands of the Tobacco Road are interbedded with cherty sandstone layers and lenses, and this lithology is associated with clean, well-sorted, loose sands that may be remnants of barrier islands or sand bars of Tobacco Road age. The same stratigraphic interval can be examined 2.5 miles (4 km) north of

Locality 8 in a steephead in the Oakey Hills Game Preserve (Locality 5). There again, the Tobacco Road red residual sands contain several layers of chert that contain silicified *Periarchus pileussinensis* at the base, and *P. quinquefarius* in the middle to upper part of the exposure (Fig. 6). This narrow belt of facies change between the Tobacco Road Sand and the "Cooper Marl" appears to be no more than three miles wide in this area.

Changes of Environment between Twiggs/Irwinton time and Tobacco Road time. Channels of tidal origin have not been observed in the Tobacco Road Sand although channels and channel-fill deposits of tidal origin are commonplace in the underlying Irwinton Sand and Twiggs Clay. It is possible that the Twiggs and Irwinton represent more of a tidal to subtidalmud and sandflat environment whereas the Tobacco Road was deposited in a slightly deeper sound or lagoon environment. During Twiggs-Irwinton time, tidal access may have been impeded by very broad tracts of extremely shallow water or exposed tidal flats. Tidal currents consequently may have scoured out channels to facilitate tidal exchange as in the present marsh system in Coastal Georgia. In contrast, during Tobacco Road time, if the water depth were slightly greater, freer tidal exchange and a consequent lessening in tidal channel scouring by tidal currents would be facilitated.

Oceanographic change. There is evidence of broad oceanographic change that commenced with the transgression during early Tobacco Road time. The presence of sediments rich in planktonic foraminifera ("Cooper Marl" and equivalents) in the Cordele, Hawkinsville, Millen, and Sylvania (shallow subsurface) areas, far from the normal habitate of these organisms in the deep oceanic environment, is probably due to changes in marine currents. At least part of the change reflected in the sandy, east Georgia Jacksonian deposits may be due to an acceleration in the velocities of oceanic and continental shelf currents (which itself is closely related to wind velocities and wind directions). The acceleration of the ocean currents could result in upwelling along the continental margins and transport of chemically and physically more stable, nutrient-rich water masses on the shelf. These more open-ocean, nutrient-rich water masses would allow proliferation of the shelf biota as well as the introduction of open-ocean plankton into the shelf waters. This may also contribute to the sudden increase in the evidence of biologic activity in the nearshore, Tobacco Road deposits as compared with the underlying Jacksonian deposits.

Climatic change. Upwelling waters are generally cooler than surface waters in that they are derived from greater depths of the ocean. As a result, the shelf water mass during late Jacksonian time should have been somewhat cooler than the shelf waters during early Jacksonian time. This should have resulted in a change in climatic conditions in the coastal and inland areas during late Jacksonian time in that cool coastal waters in subtopical to warm temperate regions of the earth produce an excess of evaporation over precipitation, resulting in more arid climates.

Change in nature of terrigenous clastic supply. Associated with the transgression after the beginning of late Jacksonian time, there was a change, relative to the earlier Jacksonian, in the nature of the terrigenous detritus supplied to the nearshore waters. Three aspects of this were:

1. a change in volume of sediments,
2. a change in the lithologic character and components,
3. therefore a possible change of the erosion and transport conditions in the source area.

The volume of early Jacksonian terrigenous clastic deposits originally in the Jacksonian belt in Georgia is very roughly 100 cubic miles (410 km^3) as compared with approximately 17 cubic miles (71 km^3) for the Tobacco Road Sand. Not only are the early Jacksonian, terrigenous clastic deposits thicker but they also extend farther downdip, or offshore than does the Tobacco Road Sand (for example see Huddlestun, Marsalis, and Pickering; 1974, Fig. 14, p. 27).

It is at first surprising that there is so little clay in the Tobacco Road Sand and its offshore equivalents, the "Cooper Marl" (10-20% clay; Pickering, 1970, p. 50, 51) and the Ocala Limestone. It would seem unlikely that the Piedmont of Georgia, which contains large amounts of schists and phyllites, should supply so little clay mineral to the continental shelf. There appear to be two alternative explanations for this deficiency: one is that the deficiency in supply is real, that the Piedmont supplied very little clay mineral to the shelf, or, the Piedmont did supply normal amounts of clays to the shelf waters, but the clays bypassed the shelf. The preferable explanation appears to be the latter in that it is difficult to construct a reasonable, convincing model in which the weathering of granites, gneisses, schists, phyllites, and mafic rocks preferentially supplies quartz to the sea but very little clay.

The most likely explanation for the bypassing of the shelf by the clay is their having been removed from the shelf in suspension. Strong currents on the shelf and strong winds in the coastal areas could conceivably produce enough turbulence and flow in the shelf waters, sound and open shelf to keep the clay particles in suspension, thereby removing much of the clay from the continental shelf and scattering the rest across the shelf as minor or trace components in the limestones.

The bypassing of the shelf by the clay may also account for much of the thinness of the Tobacco Road Sand as compared to the rest of the Jacksonian, and for the overlap of the upper Jacksonian carbonates over the lower Jacksonian clastics. If the upper Jacksonian clays had been deposit ed on the shelf rather than being bypassed, the thickness of the Tobacco Road could have been greatly increased and there would have been deposited an offshore clay facies belt between the nearshore sound/lagoon sands and the offshore, shelf limestones.

The large size and roundness of the pebbles in addition to the coarseness of the sand of the Tobacco Road requires the Tobacco Road clastics to have had a slightly different history than the underlying Twiggs and Irwinton clastics. Furthermore, the coarseness of the Tobacco Road precludes the Irwinton, and probably the underlying kaolinitic sands of early Tertiary and Late Cretaceous age, from being a source for the Tobacco Road. This suggests that the source of at least the coarser fraction of the Tobacco Road was direct erosion of Piedmont rocks and saprolite. Both the relatively large size of the Tobacco Road pebbles and the less rounded and smaller pebbles of the Irwinton near the Fall Line would suggest nearby sources for both units.

Basal Tobacco Road Geologic Event. There are several lines of evidence that suggest that a minor geologic event, tectonic and (or) climatic/oceanographic in origin, took place toward the end of Irwinton/Twiggs time (early Jacksonian) and the beginning of Tobacco Road time (late Jacksonian).

(1) Decline in clastic input. Clastic input onto the continental shelf of eastern Georgia began to diminish near the end of the early Jacksonian and the percentage of biogenic debris increased. This decline in clastic input reached a peak at the end of the early Jacksonian and the beginning of the late Jacksonian. Evidence for this decline in terrigenous clastic input consists of the following: (a) common occurrence of cherty or calcareous beds at the top of the Irwinton Sand and the base of the Tobacco Road Sand, (b) a concomitant diminishing sand content upward through this cherty or calcareous sequence, (c) widespread, very thinly laminated clay and fissile, silty clay beds at the top of the Irwinton Sand in east central Georgia (also LaMoreaux, 1946) indicating a diminishing grain size of the terrigenous clastics near the end of the early Jacksonian and (d) glauconite (green sand) beds at the top of the Twiggs Clay in the area west of the Ocmulgee River (Huddlestun, Marsalis, and Pickering, 1974; also see Fig. 6) suggesting a very low rate of sedimentation.

(2) Sea level fluctuation. In addition to a pause in the influx of clastic sediments from the continent, there is also some evidence for a slight lowering and subsequent rise in sealevel accompanying the pause in sedimentation. Flat pebbles have always been associated with beach deposits and it is within the swash zone that the flat pebbles were probably formed. Therefore the flat pebble zone of the Tobacco Road Sand is interpreted here as being a beach deposit. The maximum width of occurrence of the flat pebble zone, across the Tobacco Road belt, is no more than 25 miles in the Twiggs County area. Because it is unrealistic to assume that there was a 25 mile wide beach during late Jacksonian time, this is taken as evidence that the beach migrated during its deposition.

The flat pebble bed may represent both a regressive, lag beach deposit and a transgressive beach deposit. During the following transgression, the earlier beach deposits, strewn across the coastal belt, would be reworked and incorporated into the transgressive beach deposit. We postulate that there was insufficient time and topographic relief in the coastal area during this sea level fluctuation, to seriously oxidize, leach, and erode the regressive beach deposit before it was incorporated into the following transgressive beach. As such, this sea level fluctuation represents a small-scale regressive-transgressive event within a large-scale, regional transgression; the transgression of the Jacksonian.

(3) Biological evidence. The common presence of bioturbation in the Tobacco Road Sand, in contrast to very little evidence of biological activity in the underlying Irwinton Sand, is taken as indirect evidence of deeper water during late Jacksonian time. In itself, evidence of biological activity is not necessarily evidence of deeper water conditions, but rather of the ability of organisms to colonize the sea bottom. In this case, the ability of organisms to colonize the sea bottom would be greatly improved by even a slight deepening of the water, especially from intertidal or beach conditions to that of open sound.

Consistent with the evidence for rising sea level after the beginning of Tobacco Road time is the Twiggs-"Cooper" stratigraphic interval in Houston and Pulaski Counties. The microfauna of the "Cooper Marl" is much richer, both in concentration of foraminifera and in species diversity than is the underlying Twiggs clay. In addition, planktonic foraminifera are common in the "Cooper" and except for one species, *Truncorotalia inconspicua* Howe, planktonic foraminifera are not normally present in the Twiggs. The only interpretation possible is that the "Cooper" represents more open, stable marine conditions than the Twiggs and therefore must be relatively transgressive over the Twiggs.

Epeirogenic change. The above discussion suggests epeirogenic uplift followed by subsidence. Regression and transgression would be manifestations of this uplift and subsidence respectively. It is our opinion that the general coarsening of the sands of the Tobacco Road, over that of the Irwinton, must be a reflection of steeper stream gradients in the Piedmont because it is unlikely that the increase in stream energy, or competence, necessary for transporting larger size particles could be attained by mere entrenchment alone.

The magnitude of such an hypothesized uplift may be examined as follows: As calculated earlier the approximate volume of the original mass of the Tobacco Road Sand is very roughly 17 cubic miles (71 km^3). Because most of the Tobacco Road consists of quartz sand, then one must derive the quartz from quartzose crystalline rocks, i.e., granites, gneisses, and vein quartz. In addition the areal extent of the source lands must be considered. This can never be known with certainty since although the coastal limit of the Upper Eocene Piedmont approximated the present Fall Line, the interior drainage divide of the late Eocene Piedmont is not known. Therefore, for the sake of discussion it is assumed to have been in the vicinity of the Brevard zone, a zone of weakness in terms of weathering and erosion in Georgia. The present drainage divide south of the Brevard is assumed to approximate the late Eocene drainage divide. This drainage divide on the average is approximately 85 miles (140 km) northwest of the Fall Line and since the present length of the Tobacco Road belt is 130 miles (210 km), the areal extent of the source province would be approximately 11,000 square miles (30,000 km^2). The following equation is used to compute the volume of Piedmont rock removed during Tobacco Road time:

Volume of quartz in Tobacco Road = Volume of quartz from Piedmont source area.

$$(\text{Area}_{TR})\ (\text{Thickness}_{TR}) = (\text{Area}_{\text{Piedmont source}})$$

$$(\text{Depth of erosion}_{\text{Piedmont source}})\ C\ ;\ C = \text{constant} = 0.2$$

This is the approximate percentage of quartz in the volume of eroded Piedmont rocks.

The depth of Piedmont erosion derived from the above equation is 35 feet (12.5 m). A regional uplift of approximately 35 feet (12.5 m) would probably cause a slight change in gradient with little more than an incision into the Piedmont saprolite. This is consistent with the absence of feldspar in the Tobacco Road Sand and is in contrast to the abundance of feldspar in some of the later Miocene deposits.

It is possible that the geologic event, or events, under examination could be solely a function of an oceanographic/climatic change. The regression and transgression may simply be a result of minor eustatic changes in sea level. The influx of more sediment, more angular sediment, and more poorly sorted sediment may have resulted from a slightly more arid coastal climate caused by cooler continental shelf waters. The enhanced aridity could have been sufficient to alter the flora and ground cover, thereby facilitating erosion and incision of the drainage system. Conceivably an increase in torrential floods could have been sufficient to move coarser clastics to the shelf. The increased aridity, however, would not have been sufficient to significantly reduce chemical weathering of the rocks since quartz is the only observed constituent of the medium to pebble-size sand fraction in the Tobacco Road Sand, feldspar being absent.

Not truth, nor certainty. These I foreswore
In my novitiate, as young men called
To holy orders must abjure the world.
"If. . ., then. . .," this only I assert;
And my successes are but pretty chains
Linking twin doubts, for it is vain to ask
If what I postulate be justified,
Or what I prove possess the stamp of fact.

Clarence R. Wylie, Jr.

PALEO-ENVIRONMENTAL SYNTHESIS

The upper Eocene, Jacksonian transgression overlapped all earlier marine, coastal plain deposits in the eastern part of Georgia.

During the late Eocene, all of the Coastal Plain of Georgia, and perhaps some of the Piedmont as well, was continental shelf and below sea level. The width of the continental shelf to the southwest and southeast varied from more than 200 to 250 miles (320 to 400 km) respectively. The peninsula of Florida at that time consisted of limestone banks that were below sealevel, similar to the Bahama Banks of today. These banks extended at least 500 miles (800 km) to the south of the Fall Line.

Judging from the distribution of, and the facies indications of the offshore Ocala Limestone, the depth of water on the shelf off of Georgia during the late Eocene was no more than one or two hundred feet (30 or 60 m). The presence of larger foraminifera, calcareous algae, and abundant, coarse rubbly bioclastic debris would suggest relatively shallow agitated water within the photic zone. If the water depth at the shelf edge was about 600 feet (180 m) based on the presently accepted world-wide average, or 180 feet (55 m) based on the depth of the present shelf edge, the average inclination of the shelf bottom from the shore to the shelf edge would have ranged from 2.5 feet to as little as 0.7 feet per mile respectively. This broad expanse of very shallow shelf waters would have been sufficient to blunt most of the energy of the oceanic waves before reaching the coast.

The coastal area at this time was probably very low lying and marshy, with broad tracts of tidal flats. There was probably only a narrow band of flat, Coastal Plain-like land of variable width between the marshy shore line and the inland Piedmont (Fig. 8). The Piedmont, at this time, was probably low, rolling country that was covered for the most part by hardwood forests. These forests consisted dominantly of hickory (*Carya*), oak (*Quercus*), chestnut (*Castanea*), holly (*Ilex*), and *Engelhardtia*, an angiosperm tree now found only along the coasts of southeastern Asia. (Darrell, 1974; Pers. Com., 1978). The rivers draining the Piedmont during the early part of late Eocene time supplied only fine sand and clay to the shelf.

Tectonism in the southeastern United States had been diminishing since Sabinian time (late Paleocene and early Eocene) when the clastic detritus of the Nanafalia, Tuscahoma, and Hatchetigbee Formations covered a large part of the Georgia continental shelf. Since that time, through Claibornian (middle Eocene) and early Jacksonian time terrigenous clastic sedimentation, sediment volume, and sediment size fraction had been diminishing. Conversely carbonate sedimentation had been increasing and the limestone shelf province had been expanding. It would not be until the subsequent early Oligocene, Vicksburgian time that terrigenous clastic sedimentation would reach a minimum.

At the end of early Jacksonian time there was a decline in terrigenous clastics being supplied from the continent that resulted in finer sediment sizes at the top of the Twiggs/Irwinton and an increase in carbonate deposition on the inner shelf. This was followed by a minor but distinct marine regression (Fig. 8A). The shoreline probably did not retreat much more than about 25 miles (40 km) seaward from the Fall Line, and farther seaward, sedimentation was probably more-or-less continuous.

An influx of coarser size clastics may have occurred at the initiation of the regression. If so, the lag beach deposits were reworked in the following transgression and were incorporated in the flat pebble beach deposit. This beach deposit, the basal flat pebble zone of the Tobacco Road Sand, formed originally at the strand line. As the marine transgression progressed, the beach migrated northward leaving the flat pebble beach deposit as a discontinuous veneer in the nearshore area (Fig. 8B).

Concurrently with the transgression, the farther offshore area was starved of terrigenous clastics due possibly to the drowning of coastal areas and the lower portions of river valleys. This resulted in a more irregular coastline indented with estuaries, and estuaries once formed would act as sediment traps for the offshore area (Fig. 8B). These offshore areas were cut off from a terrigenous clastic influx and as a result, due partly to the rising base level, deposits of biogenic origin continued to accumulate.

The new influx of clastics could have resulted in the silting up of the estuaries, followed by stabilization of the shoreline, movement of coarse terrigenous clastics onto the continental shelf, and construction of a barrier island-like system in the offshore area (Fig. 8C). Once the barrier island-like system had been constructed, a broad sound or lagoon was formed behind the barrier. The sound or lagoon water mass was characterized by low salinities, and more variable physical and chemical conditions than the shelf water mass seaward of the barrier. The sound acted as a sediment trap in which poorly sorted, perhaps muddy, medium to coarse sand and pebble-size sediment was spread out along the sound bottom. The continual ebb and flow of the tides and the probable strong coastal winds kept the clays in suspension, carrying them out into the open shelf waters through tidal channels. Seaward of the barrier, carbonates accumulated.

At the time of the beginning of the Tobacco Road transgression in the nearshore area, the current veolocities adjacent to and on the continental shelf accelerated, causing shallow upwelling of deeper waters at the shelf edge. This resulted in greater circulation of open ocean, nutrient-rich, environmentally stable water; an increase in biological production in the shelf waters; and a slightly more arid climate onshore.

There were evidently alternating periods of higher and lower energy deposition alternating with nondeposition, based on the not quite obliterated thin laminae and wisps of fine clayey sand. The sediments were commonly reexcavated by burrowing organisms, probably dominated by annelid worms, arthropods, and echinoids. Rarely, however, did the burrowing organisms completely mix the sediment into a homogeneous mass; mostly the finer and coarser sands are incompletely mixed, suggesting that originally perhaps, the sorting may have been better.

Sometime immediately before the end of the Eocene, Tobacco Road deposition ceased. The basic geographic, facies pattern that was established just after the beginning of Tobacco Road time remained fixed. At the end of Tobacco Road time, sedimentation appears simply to have ceased.

The two most likely reasons for this cessation are:

1. there was a regression and a relative lowering of sealevel, or
2. sealevel remained the same but the zone of base level oscillation on the continental shelf stabilized (subsidence ceased) and the shelf became a zone of sedimentary bypass.

The current information does not allow us to decide between these two possibilities. However, it is observed by the authors that where unweathered Oligocene deposits overlie unweathered upper Eocene deposits in Georgia, the contact between the units is not undulatory, appears conformable, and there is no indication of pre-Oligocene leaching or weathering on the top of the Eocene deposits.

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APPENDIX

LIST OF CITED LOCALITIES AND CORES

Burke County

Locality 1 On River Road, 0.75 miles southeast of the junction of Griffins Landing Road and River Road, in road cut on north side of Little Beaverdam Creek, 3.6 miles north-northwest of Girard.

Locality 2 Sand pit on north side of Thomsons Bridge Road at Hatchers Mill, 0.2 miles west of bridge over Brier Creek. Upper part of formation is exposed in a series of road cuts between 0.4 and 1.4 miles east of bridge over Brier Creek on Thomsons Mill Road.

Locality 3 South side of roadcut on county road, 300 feet east of Newberry Creek and 2.8 miles east of Shell Bluff Post Office.

Locality 4 Exposure in excavation at Georgia Power Company's Plant Vogtle.

GGS 2136 (Burke 2) on south side of Ga. 17, 4.75 miles east of Midville, and 0.4 miles east of Bark Camp Creek; 32°49′30″N, 82°09′13″W.

Houston County

Locality 5 Exposure in steephead, 100 feet south of dirt road, 0.5 miles southeast of Kathleen Observation Tower in Oakey Woods Wildlife Management Area.

Laurens County

Locality 6 Exposure in steephead, 100 feet north of US 441, 0.2 miles southeast of Wilkinson-Laurens County line.

Locality 7 Roadcut on Buckeye Road, 1.0 miles south of Laurens-Johnson County line.

Pulaski County

Locality 8 Exposures along Ga. 247 near Houston-Pulaski County line.

Richmond County

Locality 9 Babcock and Wilcox Company, Albion kaolin pit at Hephzibah, exposure in southernmost pit.

Locality 10 Type locality of Tobacco Road Sand, exposure on east side of Morgan Road, 0.35 miles north of junction of Morgan Road and Tobacco Road

Warren County

Locality 11 J. M. Huber Corporation Palmer Pit, 3.7 miles north of Jefferson-Warren County line on east side of Ga. 17.

Washington County

Locality 12 Exposure in creek bank, 1800 feet northwest of Deep Cut Bridge over Central of Georgia Railroad, in small branch of Sandy Hill Creek.

Locality 13 Type locality of Sandersville Limestone, 0.8 miles south of courthouse in Sandersville.

GGS 1168 (Washington 6 and 6A), on south side of Kaolin Road, 1.0 miles west of junction Kaolin Road and Ga. 15, located at bend in Kaolin Road; 32°57′35″N, 82°49′35″W.

GGS 1182 (Washington 10), located on north side of Oconee Road at bend in highway, about 0.6 miles west of junction Oconee Road and Ga. 15, 0.6 miles east of junction Oconee Road and Ga. 68, and 0.9 miles northwest of the center of Tennille; 32°57′00″N,82°49′20″W.

GGS 1187 (Washington 15),on west side of county dirt road, about 500 feet north of Deep Cut Bridge, about 1.3 miles south of junction Oconee Road and dirt road, and about 3.4 miles southwest of the center of Tennille;

Wilkinson County

Locality 14 Roadcut on a kaolin haul road 1.3 miles northwest of junction of Wesley Church Road and Ga. 96.

A GRAVITY SURVEY OF THE DALTON, GEORGIA AREA

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ABSTRACT

A simple Bouguer gravity anomaly map, contoured at 1.0 mGal from 694 data points between 37°37.5'N and 37°52.5'N and 84°45'W and 85°00'W, delineates the southern terminus of a sedimentary basin northeast of Dalton, Georgia. The basin is apparently outlined by sharp gradients in the Bouguer anomalies north of Dalton and 10 km west of the Great Smoky Fault. An unreversed seismic refraction line trending southeast further supports a basin edge 10 km west of the Great Smoky Fault. The depth to crystalline rocks 3 km northeast of Dalton is interpreted to be 2.1 km. Both seismic and gravity data suggest basement influence in which crystalline rocks are progressively shallower in depth toward the east.

INTRODUCTION

The Dalton, Georgia gravity survey area is located in the Valley and Ridge Province of northwest Georgia. The survey area is in the Great Valley (Fig. 1) and lies between two prominent topographic highs. To the east the massive Blue Ridge Mountains exceed 1200 meters in elevation. To the west the Armuchee Ridges rise nearly 600 meters. The gravity survey area is bounded to the east by the Great Smoky Fault and to the west by the Rome Fault. The possible involvement of basement structure in the control of sedimentation or in the origin of these faults has been a topic of significant controversy in Appalachian Tectonics (Cooper, 1968). The purpose of this report is to present an analysis of gravity and seismic data in the Dalton area, and to relate this data to the surface geology and basement structure, and perhaps, contribute to the understanding of Appalachian tectonics.

The Bouguer anomaly map of the Dalton, Georgia area is the result of a class exercise during the spring term of 1976. The unreversed seismic refraction line data were collected in an attempt to obtain seismic velocities and to elucidate the structures implied by the gravity data.

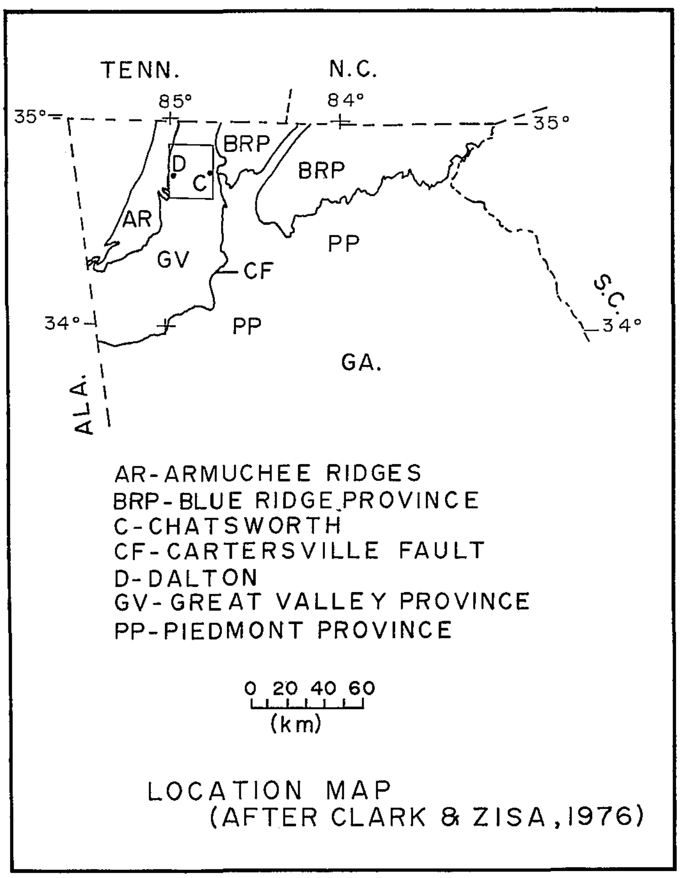

FIGURE 1. Location Map.

GEOLOGY

Butts (1946), Munyan (1951), and Cressler (1974) have investigated the geology of the Dalton, Georgia gravity survey area. Interpretation of geophysical data presented here is based mainly on the more recent work of Cressler (1974) (Fig. 2).

Just east of the Dalton gravity survey area lie the predominantly east to southeast dipping crystalline rocks of the Blue Ridge and Piedmont Provinces. The gravity survey area is assumed to be underlain by Cambrian or Precambrian crystalline basement rocks. The principle rock types are metamorphosed sedimentary rocks and granitic rocks.

The basement is overlain by the folded Paleozoic sedimentary rocks ranging in age from Cambrian through Mississippian. The sedimentary rocks were originally horizontal and were subsequently compressed, faulted, uplifted, and folded into their present form.

The Paleozoic rocks shown in Figure 2 as described by Cressler (1974) are:

MISSISSIPPIAN DEVONIAN -
- MDc - includes Fort Payne Chert at top, Armuchee Chert at bottom and Chattanooga Shale between

ORDOVICIAN -
- Ob - Bays Formation
- Oa - Athens Shale
- On - Newala Limestone
- Oєk - Knox Group

CAMBRIAN -

Ꞓcs - Conasauga Shale
Ꞓcl - Conasauga scattered limestone outcrops
Ꞓcls - Conasauga limestone including some shale
Ꞓcsl - Conasauga Shale with some limestone
Ꞓch - Chilhowee Group - massively bedded quartz-pebble conglomerate

STRUCTURE

The Dalton gravity survey area is characterized by broad open folds and major and minor faults with the western margin more steeply folded and faulted. The Rome Fault is one of the major thrust faults in the southern Appalachians, extending hundreds of kilometers from Alabama across Georgia and Tennessee. According to Cressler (1974) the Rome Fault in the Dalton gravity survey area brought the middle Cambrian Conasauga Formation into contact with the Mississippian Floyd Shale with an accumulated stratigraphic throw of about 2100 meters. Cressler believed the fault to be a flat-lying bedding-plane thrust that originated in the Conasauga or Rome Formation. Remnants of the thrust sheet indicate rock was displaced westward 8 to 16 km during the time in which the Conasauga was uplifted. The area just north of Chatsworth is cut by several smaller high-angle faults in which displacements are thought to have been small. Unlike the major thrust faults which are mainly in shale these high angle faults cut through carbonate rocks.

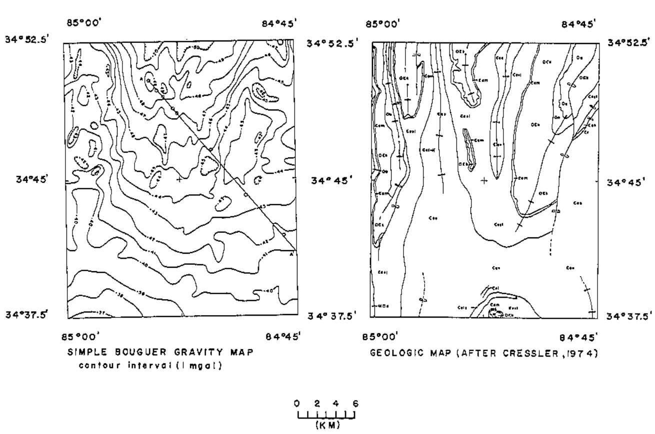

FIGURE 2. Simple Bouger gravity map and geologic map.

REGIONAL GRAVITY

Figure 3 shows the Dalton gravity survey area outlined on the Bouguer anomaly map of Georgia (Long and others, (1972). The regional trend of the Bouguer anomalies is evident. The regional Bouguer anomaly map, which was compiled from a station spacing of 7 to 10 km, indicates the influence of deeper or regional features. The outlined area shows from south to north a progressively more negative trend from -40 to -50 mGal. This trend can be attributed to the increasing thickness of unmetamorphosed sedimentary rocks west of the Great Smoky Fault.

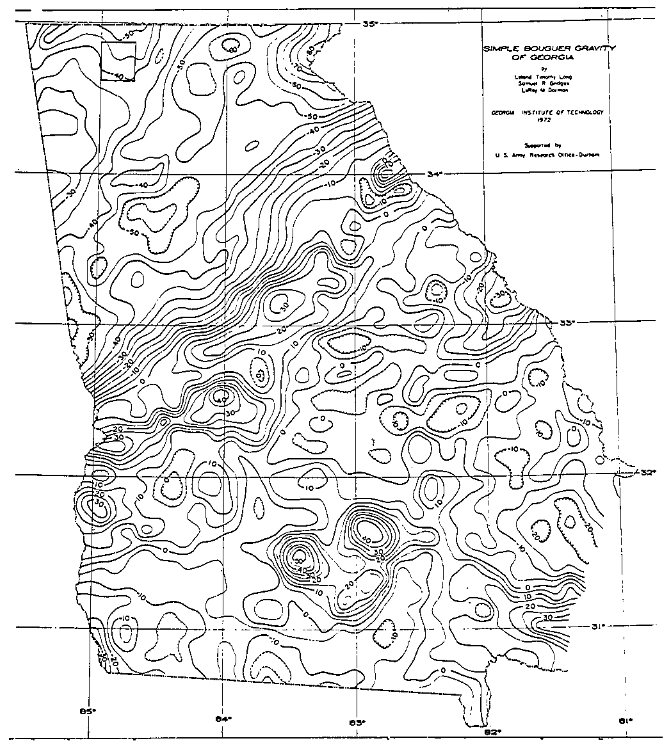

FIGURE 3. Simple Bouger gravity map of Georgia.

DALTON AREA SIMPLE BOUGUER ANOMALY MAP

The simple Bouguer anomaly map shown with the corresponding geology (Fig. 2) was drawn from 694 data points with a spacing of 0.5 km east-west and 1.0 to 1.5 km north-south. The raw data were obtained with Worden and Lacoste-Romberg gravity meters with instrumental precisions (one standard deviation) of ± 0.2 mGal and ± 0.05 mGal respectively. The map shown in Figure 2 was contoured at 1 mGal with an estimated precision of ± 0.4 mGal for interpolated values between data points.

The regional trend shown on the State map (Fig. 3) is evident in the detailed map. However, the closer spacing of data points allows a more detailed observation of the local effects of surface geology. In the upper half of the Bouguer anomaly map of the Dalton area the contours form a U-shaped configuration open to the north with Bouguer gravity anomalies becoming progressively more negative in the same direction. The Bouguer anomalies become more positive to the east, west and south. Within this U-shaped configuration are two local anomalies which conform well to outcrops of the Knox (OꞒk). In general the Knox Group would be expected to be more dense than the surrounding shales but there are numerous exceptions. For the most part where the Knox crops out forming a slight topographic high, the Bouguer anomalies indicate either negative contour closure or local negative anomalies. However, where the Knox is below the surface, generally more positive anomalies are observed. This relation is best illustrated near the northern edge of the Dalton map area (34°53′) where the two -50 mGal closures denote the edge of the Knox in a synclinal structure. The northeastern portion of the map also indicates local negative Bouguer anomalies since the contour lines are deflected south. The Knox is present in this vicinity also. The northwestern portion is more complex and difficult to interpret due to the tight folding, more numerous faults, and the resulting closely spaced lithologic units. Just east of the center part of the Dalton Bouguer anomaly map is a large exposure of the Knox. The gravity data near this exposure of the Knox indicate that the Bouguer anomalies become more positive, forming a Bouguer anomaly plateau around -42 mGal. The negative Bouguer anomaly previously associated with the Knox Group is reflected in this area by a steepening of the gradient. The negative influence of the Knox Group is possibly being overridden by a prominent structure or dense rocks in the basement. The Bouguer anomalies and geology in the southern half of the map area are uniform and monotonous for the most part.

SEISMIC REFRACTION DATA

An unreversed seismic refraction line was obtained along the profile designated A-A′ in an attempt to further define the sub-surface structure and determine the relation between the surface geology and gravity data. The circles plotted on line A-A′ represent seismic recording sites.

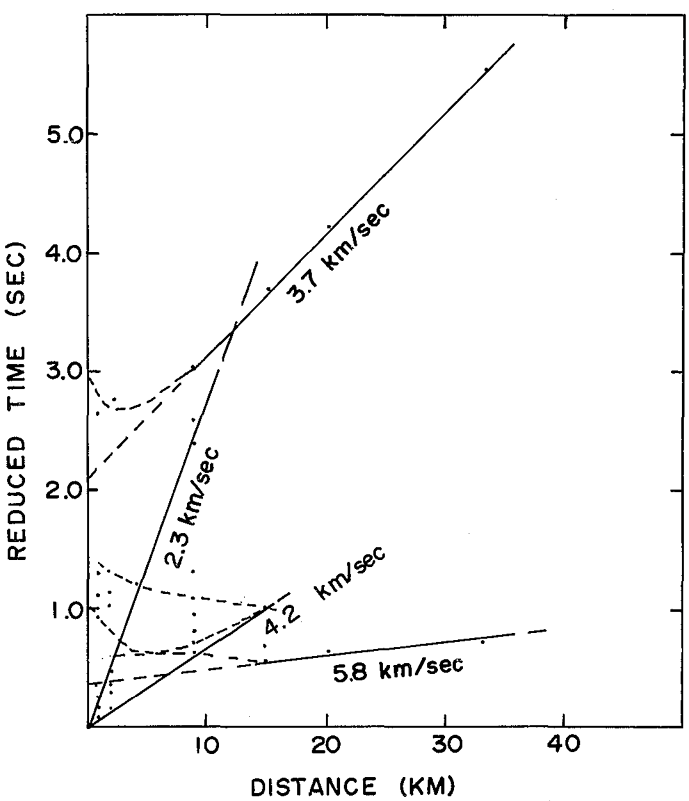

FIGURE 4

The Dalton, Georgia, Rock Products quarry provided blast data for the curves shown in Figure 4. Origin times were obtained within ± 0.01 seconds by recording the blast a few feet from the detonation site. The curves shown in Figure 4 contain traces of the blast data recorded on magnetic tape. The curves (Fig. 4) look similar to the two-layer horizontal-interface model but upon closer examination of the record at 9 km a significant time delay is observed. The P and S direct wave velocities in the unmetamorphosed sedimentary rocks (Fig. 4 or 5) are 4.2 km/sec and 2.3 km/sec respectively. The secondary arrivals or refracted phases travel with velocities characteristic of the lower media, in this case the proposed basement. At distances of 20 to 35 km the P and S waves travel at 5.8 km/sec and 3.7 km/sec respectively. No depth computations were made based on the S wave since there was some uncertainty as to the actual first arrival times. The computation of depths are based largely on the P phases. The arrival interpreted as the reflected P-wave (Fig. 4 or 5) implies a two-way travel time of about 1.0 sec or a depth of 2.1 km. Of particular interest is the implied time delay in the vicinity of 0.0 to 10.0 km (Fig. 4) for the refracted P-phase. Since the refracted P-phase travels with the velocity of the lower media this time delay is apparently caused at least in part by the basement surface configuration. The basement is interpreted as being deeper near point A and shallower toward A′. By extrapolating the delayed refracted P-phase to the ordinate axis and computing the depth below shotbreak one obtains a depth of 2.1 km which is in agreement with the two-way time for the reflected phase. The calculated depth at a distance of 9 km is approximately 1.8 km and near 25 km it is about 1.2 km. The certainty of the depth estimate near the origin is dependent on the interpretation of the delay time and the reflection and hence should be within 10 percent. Because of the sparse spacing of seismic recording sites in the 5 to 15 km range the depth estimates are less certain at greater distances. The important result is that a significant change in the depth to basement does occur and the basement rocks are interpreted as being progressively closer to the surface to the southeast.

GRAVITY PROFILE AND MODEL

In order to integrate the refraction line interpretation with the gravity data, a profile was interpolated along the same line A-A′ (Fig. 6). The solid line represents the observed relative Bouguer anomaly along the profile. The dashed line indicates the computed gravity produced by the model, shown in the lower portion of the figure. The reasonable fit of the theoretical curves implies that the discontinuity in the basement as determined from the seismic data can explain the gravity anomalies. The model was constrained to fit the calculated depths to basement obtained from the travel time curves. A density contrast of -0.18 gm/cm^3 was used for modeling.

CONCLUSION

The strong regional trend shown in the regional Bouguer anomaly map is observed also in the Dalton area Bouguer

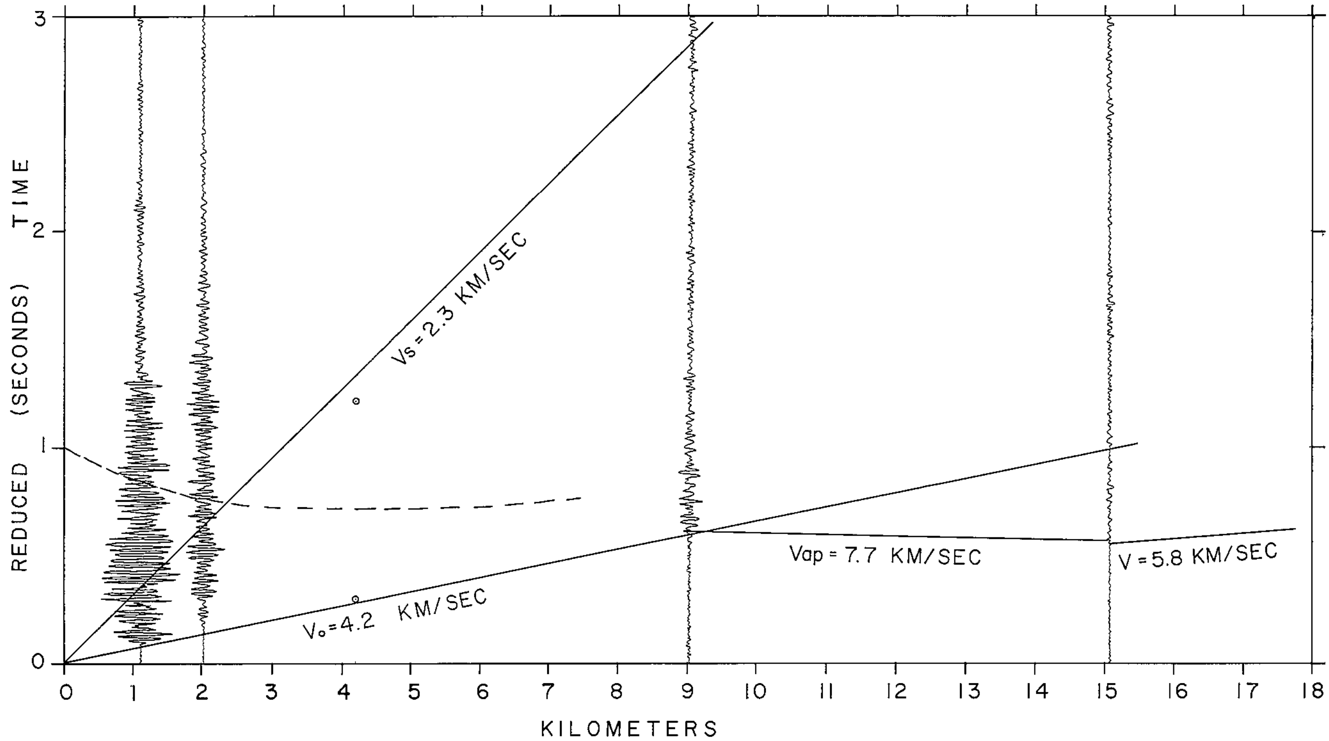

FIGURE 5

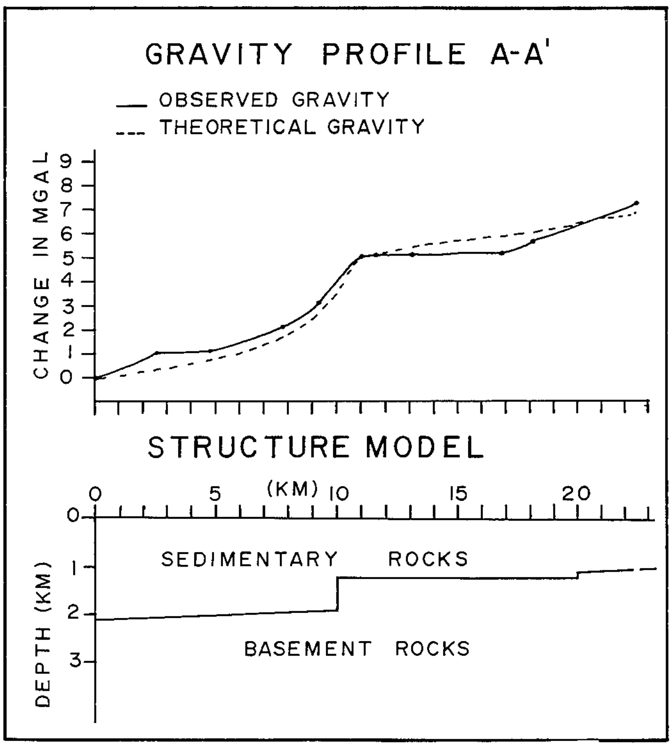

FIGURE 6. Gravity profile A-A'.

anomaly map. The U-shaped configuration of the contours may indicate a basin like structure which shallows to the south and deepens northward. Both seismic and gravity data suggest the existence of a sedimentary basin in which basement rocks are progressively shallower in depth toward the east. The eastern edge of the basin which is located 10 km west of the Great Smoky Fault, may be fault controlled. The depth to crystalline rocks 3 km northeast of Dalton is interpreted to be 2.1 km. The Knox Group produces a local negative anomaly where it crops out.

ACKNOWLEDGMENTS

The authors wish to thank Mr. C. L. Brown, Mr. David Dunbar, and Mr. Jim Fulford who also participated in gravity data acquisition and reduction. The research was supported in part by the National Science Foundation (Grant DES-75-15765).

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THE ALTO ALLOCHTHON: A MAJOR TECTONIC UNIT OF THE NORTHEAST GEORGIA PIEDMONT

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ABSTRACT

A major allochthonous sheet, named the Alto Allochthon, has been recognized southeast of the Brevard Zone in northeast Georgia and northwestern South Carolina. It was recognized by occurrence of inverted metamorphic zones and truncation of structures and geologic contacts.

The allochthon is post-metamorphic and either coeval with or later than F_2 folding. A Brevard Zone fault cuts a klippe of the allochthon in South Carolina, thereby providing a minimum age. Similarity of timing of emplacement of this structure to others in this region suggests a possible Devonian age of emplacement.

The root zone for the allochthon is unknown. It may be rooted in the Inner Piedmont, but lithologic similarities to Blue Ridge units raise the possibility that it is rooted northwest of the Brevard Zone.

INTRODUCTION

The purpose of this paper is to describe a structural feature located southeast of the Brevard Zone in part of the Chauga Belt of northeast Georgia and northwestern South Carolina (Fig. 1). This feature was recognized during the course of detailed geologic mapping in the Chauga Belt and Inner Piedmont. It was originally recognized as a series of klippen resting upon some of the higher hills in the Chauga Belt of South Carolina. Later is was discovered that these klippen are remnants of a more continuous sheet of allochthonous rocks farther southwest in Georgia. I propose to call this feature the Alto Allochthon after the small town of Alto, Georgia, which is centrally located in the structure.

The extent of the Alto Allochthon is now known to be from the edge of northwestern South Carolina to just east of Gainesville, Georgia. Its internal structure is not fully known at the present time, but its extent has been well delineated by detailed geologic mapping of the northeastern end and reconnaissance of the remainder (Fig. 1).

Interpretation of aeroradioactivity and aeromagnetic maps of a major part of the northeast Georgia Blue Ridge and part of the Piedmont by Higgins and Zietz (1975) and resulted in a general delineation of the Alto Allochthon as a separate feature. Moreover, the contacts were interpreted as faults.

EVIDENCE SUPPORTING THE ALLOCHTHON

The rocks within the Alto Allochthon consist of sillimanite grade mica and granitic gneisses, muscovite-biotite schist, aluminous schist, amphibolite, and quartzite.

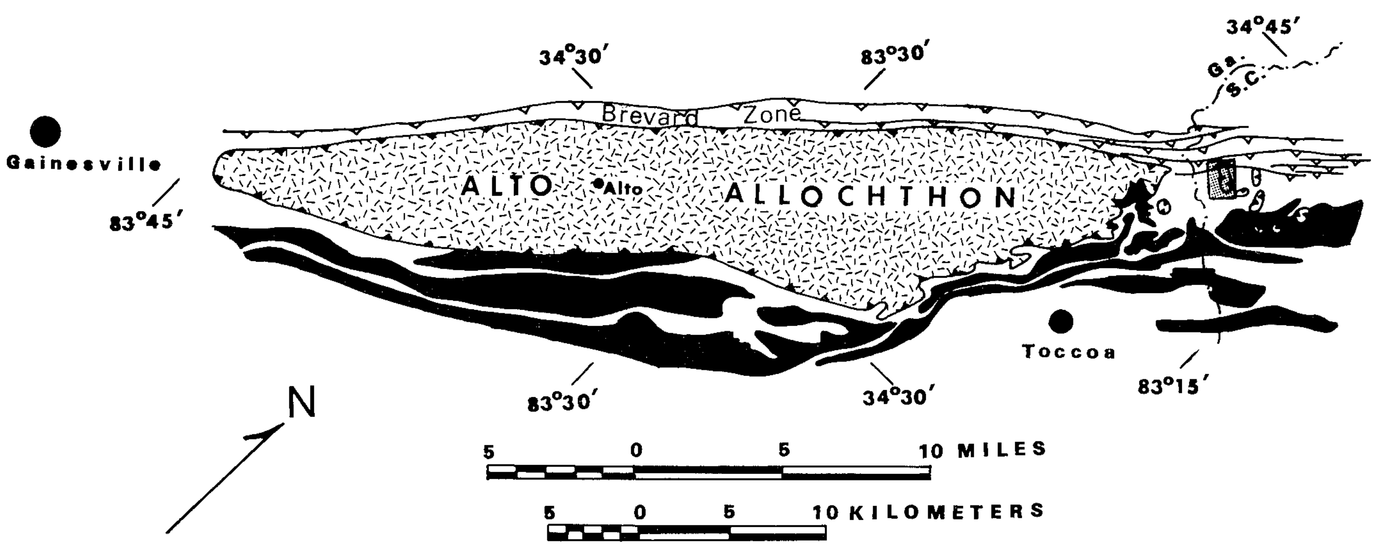

FIGURE 1. Generalized geologic map showing the location of the Alto Allochthon in relation to the Brevard Zone in some of the nearby geology. Chauga Belt and Inner Piedmont granitic rocks are shown in black. Other rocks (Poor Mountain Formation, Chauga River Formation and Inner Piedmont amphibolite) are unpatterned. Small shaded area at northeast end of map is the area of Figure 2.

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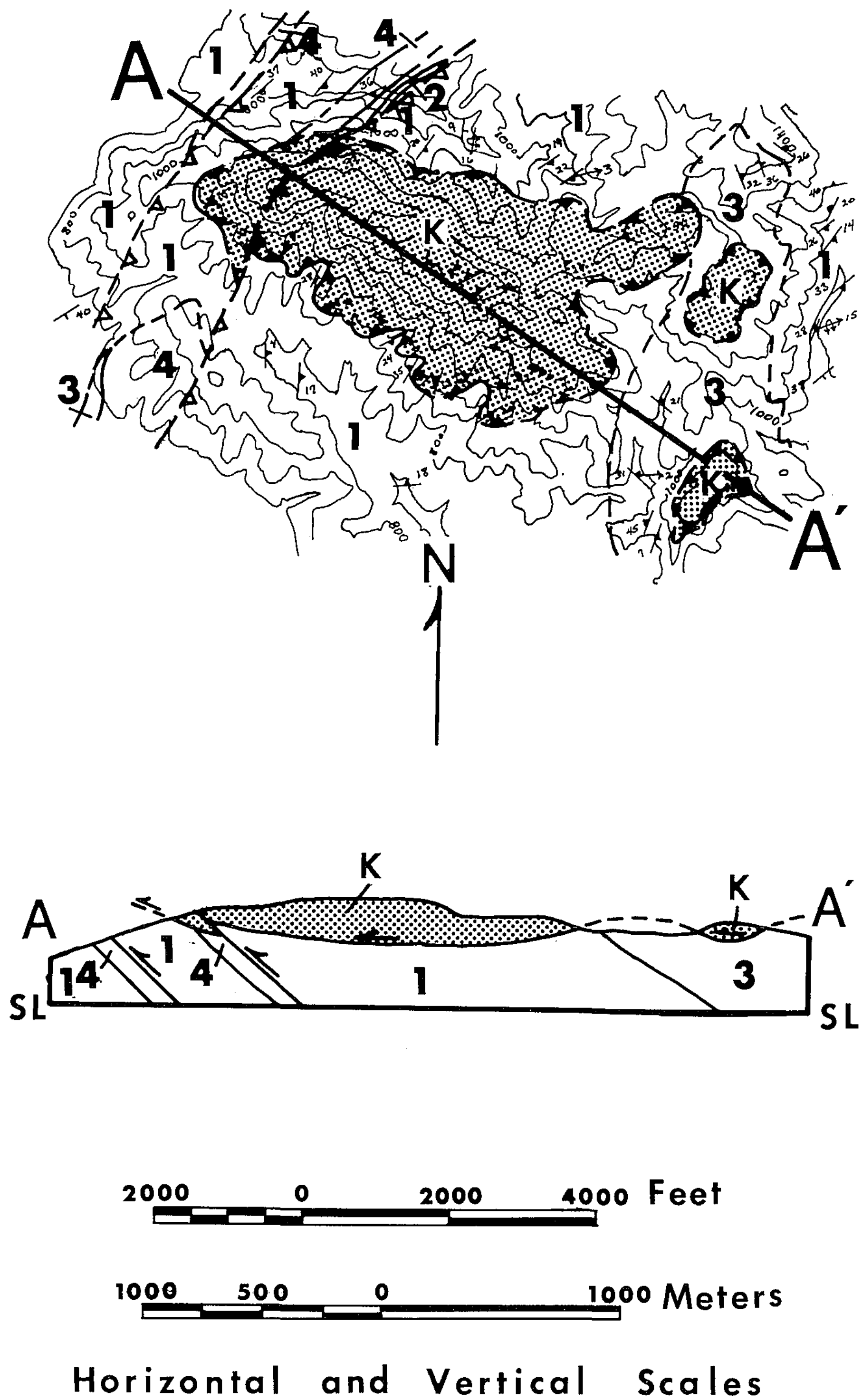

FIGURE 2. Detailed geologic map of a portion of Tugaloo Lake Quadrangle, South Carolina-Georgia and cross-section showing relationships between several klippen (K) of the Alto Allochthon and underlying rocks and with Brevard Zone faults which cut them. One of these faults was reactivated to cut the large klippe after the Allochthon was emplaced. K-muscovite-biotite schist, biotite gneiss, amphibolite, quartzite. 1-Brevard phyllite and Brevard-Poor Mountain Transitional Member of Poor Mountain Formation. 2-Chauga River carbonate. 3-Poor Mountain Amphibolite. 4-Henderson Gneiss. Contacts-solid, exact; dashed, approximate. Contours in feet.

The rocks are coarser grained and of higher metamorphic grade than the surrounding Chauga Belt rocks, which may be as high as staurolite grade southeast of Gainesville, Georgia (W. H. Grant, oral commun., 1973), but as low grade as upper greenschist facies farther northeast (Hatcher, 1969).

Chauga Belt folds and contacts are truncated by the allochthon. Post-metamorphic F_2 folds[1] are truncated by this structure, and several contacts between rock units of contrasting lithology have also been overridden. However, a subparallel relationship of S-surfaces (S_1) exists along the boundary of the allochthon and the rocks beneath it. Rocks on the southeast side of the allochthon and in the Chauga Belt dip northwest; rocks along the northwest boundary dip southeast.

TIME OF MOVEMENT ON THE ALLOCHTHON

Relative time of emplacement of the Alto Allochthon is estimable from the features present and structures cutting it. This is particularly true in the area of detailed investigation along its northeastern end.

Post-metamorphic F_2 folds truncated by the allochthon provide a maximum age for its emplacement. The fact that it truncates these folds shows that the allochthon post-dates them. However, rocks in the vicinity of the contact do not appear very cataclastic, except along its northwestern edge, but this cataclasis is probably related to the Brevard Zone and not to the Alto Allochthon.

A Brevard Zone related fault cuts one of the klippen of the allochthon in South Carolina (Fig. 2), thereby establishing a minimum age for the emplacement of the allochthon. The fault which cuts the klippe was probably formed during a late brittle phase of movement on the Brevard Zone. This particular fault exhibits little offset but definitely displaces the contact of the klippe with the underlying rocks. However, mylonitic Henderson Gneiss has overridden the klippe, thereby dating its emplacement as later than this event.

Truncation or inversion of metamorphic zones, like the F_2 folds, places a maximum age on emplacement of the allochthon. Apparently, the mass had sufficiently cooled when it was emplaced that it did not reequilibrate the rocks beneath, since there appears to be no interruption of the regional southeastward increase of metamorphic grade from the Chauga Belt into the Inner Piedmont. If the regional thermal peak occurred about 450-480 m.y. ago, as Butler (1972) suggested, the allochthon must have been emplaced later.

Russell (1976) has determined that mylonites along several faults experienced equilibration of rubidium and strontium isotopes during the Devonian (about 350-375 m.y.). Odom and Fullagar (1970, 1973) found the same to be true of mylonites in the Brevard Zone. If this relationship is generally true for most of the mylonites in the Blue Ridge and Piedmont, and if this equilibration of isotopes represents the principal episode of movement on these faults, then the Alto Allochthon may also have been emplaced at this time, or possibly before, since cataclastic rocks are not present on the contact. A Devonian age of movement would fit with other data placing the time of movement after F_2 folding and before brittle deformation in the Brevard Zone.

ROOTS OF THE ALLOCHTHON

Perhaps the most important questions related to the Alto Allochthon are those of its souce and amount of transport, as details of the internal structural geometry of this feature remain unknown. The simplest solution to this problem is to assume that the Alto Allochthon is rooted in the Inner Piedmont to the southeast. It would therefore have experienced northwestward tectonic transport. The occurrence of sillimanite grade nappes in the Inner Piedmont (Griffin, 1971, 1974) lends support to this line of reasoning.

Rocks within the Alto Allochthon are lithologically similar to rocks of the Tallulah Falls Formation (Hatcher, 1971) west of the Brevard Zone in the Blue Ridge. All marker units used in mapping the Tallulah Falls Formation rocks in the Blue Ridge have been recognized within the allochthon. Moreover, the most common rock types in the allochthon (biotite gneiss, muscovite-biotite schist and amphibolite) are likewise the most abundant rock types in the Tallulah Falls Formation. Hence the possibility exists that the Alto Allochthon was derived from the Blue Ridge and crossed what is now the Brevard Zone through some means of southeastward tectonic transport.

CONCLUSIONS

1. An allochthonous sheet of high grade rocks rests upon lower grade Chauga Belt rocks southeast of the Brevard Zone in northeast Georgia and northwestern South Carolina. This sheet is named the Alto Allochthon.
2. The time of emplacement of the allochthon is bracketed by the regional metamorphic-thermal peak followed by post-metamorphic F_2 folds and later brittle deformation along the Brevard Zone. Within these limits, it is concluded that the allochthon may have been emplaced during the Devonian, a time of equilibration of isotopes and possible movement on several other faults in the crystalline southern Appalachians.
3. The roots of the allochthon are unknown. It could be rooted in the Inner Piedmont, but similarity of rocks within the allochthon to rocks in the Blue Ridge raises the possibility that it may have been derived from the northwest.

ACKNOWLEDGMENTS

Support for the field work involved with this study was provided by the National Science Foundation Grant GA-1409, the South Carolina Division of Geology, and Clemson University. Several student assistants contributed significantly to the detailed geologic mapping. Most notable is Louis L. Acker, who first brought in some high grade amphibolite and quartzite from a high hill deep within the Chauga Belt. Stephen L. Wood assisted in walking the

[1] A discussion of the structural sequence in the region may be found in Hatcher (1974, 1976, in preparation).

contact on the faulted klippe (Fig. 2). William M. Rivers assisted on several traverses in northwestern South Carolina.

Critical review by M. W. Higgins and K. I. McConnell is very much appreciated. However, the author remains responsible for all errors and misinterpretations.

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STRUCTURAL AND LITHOLOGIC CONTROL OF SWEETWATER CREEK IN WESTERN GEORGIA

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ABSTRACT

Sweetwater Creek, draining portions of Cobb, Douglas and Paulding Counties, Georgia, shows a complex interrelation of structural and lithologic control on streamflow direction. Differences in the degree of control allow the course of Sweetwater Creek to be divided into two distinct parts. North and west of Austell, Sweetwater Creek flows within northeast trending hornblende gneisses, felsic gneisses and mica schists. In this section, Sweetwater Creek flows in a northeastward direction for approximately 9 miles (14 km) following the axial trend of the Austell-Frolona antiform. Sweetwater Creek is the only stream in this part of the Piedmont to flow in this direction for a significant distance. Large scale structures control stream flow direction along this part of the creek while small scale structures exhibit little control. Jointing, although moderately well developed, has little or no affect on the direction of flow. Control of the creek seems to result from the trend and plunge of the Austell-Frolona antiform and the presence of the massive Austell Gneiss in the nose of the fold. The southeastern part of the creek, extending from just south of Austell to the Chattahoochee River, crosses a series of northeast trending quartzites, aluminous schists, felsic gneisses, and amphibolites of the Sandy Springs Group. Here stream direction is primarily influenced by a well developed set of steeply dipping joints striking approximately N 50°W. Resistant and poorly jointed lithologies provide a second, but less important, controlling factor.

INTRODUCTION

The drainage basin of Sweetwater Creek encompasses an area of approximately 245 square miles (635 sq. km.) including within it the cities of Austell, Villa Rica, and part of Marietta (Fig. 1). Sweetwater Creek's drainage basin can be divided into two distinct sections, a northwestern section and a southeastern section with Austell being the point at which the sections are separated. This division is based on differences in physical characteristics and geologic controls exhibited by the two sections. The purpose of this paper is to describe and interpret the relationship between geologic controls and the physical characteristics of Sweetwater Creek.

LOCATION AND PHYSIOGRAPHY

Sweetwater Creek flows through portions of Cobb, Douglas and Paulding Counties, Georgia. Its headwaters are in southern Paulding County near New, Georgia and it

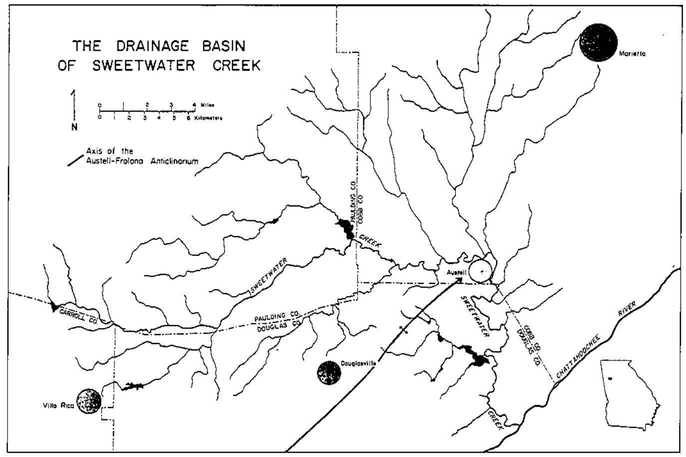

FIGURE 1. Drainage basin of Sweetwater Creek.

terminates in southeastern Douglas County where it flows into the Chattahoochee River. The creek is a major supplier of water for the city of East Point and until 1971 was also a source of water for Austell.

Sweetwater Creek flows across portions of the Central Uplands and Gainesville Ridge Districts of the Piedmont Physiographic Province (Clark and Zisa, 1976). The boundary between these two districts also serves as the approximate boundary between the distinctly different northwestern and southeastern sections of the creek. Streams in the Central Uplands District flow in broad, open valleys which is characteristic of the northwest section of Sweetwater Creek. To the south, in the Gainesville Ridges District, streams occupy narrow, v-shaped valleys and flow is, for the most part, perpendicular to the regional geologic structures. While drainage in both sections of Sweetwater Creek and in both districts is rectangular in form, the northwestern section exhibits a more modified rectangular drainage pattern than the southeastern section.

GENERAL GEOLOGIC SETTING

The entire course of Sweetwater Creek lies within the metamorphic and igneous rocks of the geologically complex northern Piedmont of Georgia. The northwestern section of Sweetwater Creek is underlain by amphibolite grade metasedimentary and metaigneous rocks (Fig. 2). North and west of the Austell-Frolona antiform rock units are composed primarily of amphibolite and felsic gneisses with lesser amounts of garnet-mica schist and metagabbro. Rock types present in the Austell-Frolona antiform are: granite gneiss containing relict feldspar phenocrysts, muscovite schist, garnitiferous biotite-muscovite schist, and amphibolite (Fig. 2).

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As Sweetwater Creek begins to flow to the south near Austell (Fig. 2) it crosses the Chattahoochee Fault Zone and flows into the Sandy Springs Group (Higgins and McConnell, in press). The Sandy Springs Group is composed of northeast trending quartzites, aluminous schists, felsic gneisses and amphibolites.

Continuing southeastward, Sweetwater Creek enters the lithologies characteristic of the Brevard Fault Zone. Rocks present here are primarily mylonites, blastomylonites, phyllonites, and "button" schists.

FOLDING AND METAMORPHISM

The rocks in the Sweetwater Creek area have undergone at least three generations of folding and one or more episodes of progressive regional metamorphism. Folding (F_1) during amphibolite grade metamorphism has produced an axial-plane foliation now expressed by the regional foliation (Fig. 3). Second generation (F_2) folds trend approximately N40° E. This generation of folding, of which the Austell-Frolona antiform is typical, dominates the outcrop pattern in the northwestern section of the drainage basin. Third generation (F_3) folds are open, gentle crossfolds which trend approximately N55° W. Localized retrograde metamorphism has occurred in the rocks of the Brevard Fault Zone in the southeastern section of the drainage basin.

FAULTS

The southeastern section of Sweetwater Creek's drainage basin crosses three major fault zones. Rocks present in and northwest of the Austell-Frolona antiform have been thrust over by the Sandy Springs Group along the Chattahoochee Fault (Hurst, 1973). Recent reconnaissance mapping indicates that the rocks of the Chattahoochee Fault block have themselves been overthrust along another fault zone, here informally named the Blair Bridge fault. This second fault is believed to have formed as a result of the presence of the massive Austell Gneiss. Where it is present, the Austell Gneiss impeded the northwestward movement along the Chattahoochee Fault and precipitated movement along a second fault, the Blair Bridge fault, which overthrust the rocks of the Chattahoochee Fault block. Northeast of where the Austell-Frolona antiform plunges beneath the surface, the Blair Bridge fault dies out due to the absence of the Austell Gneiss. Rocks present near the Blair Bridge fault zone show flattened and in some cases rotated garnets; however, the main evidence for the presence of a fault in this area is the termination of units trending down from the northeast (see Fig. 2). Further to the southeast, as it nears the Chattahoochee River, Sweetwater Creek crosses the Long Island Fault (Higgins, 1966) and then flows across the cataclastic rocks of the Brevard Fault Zone.

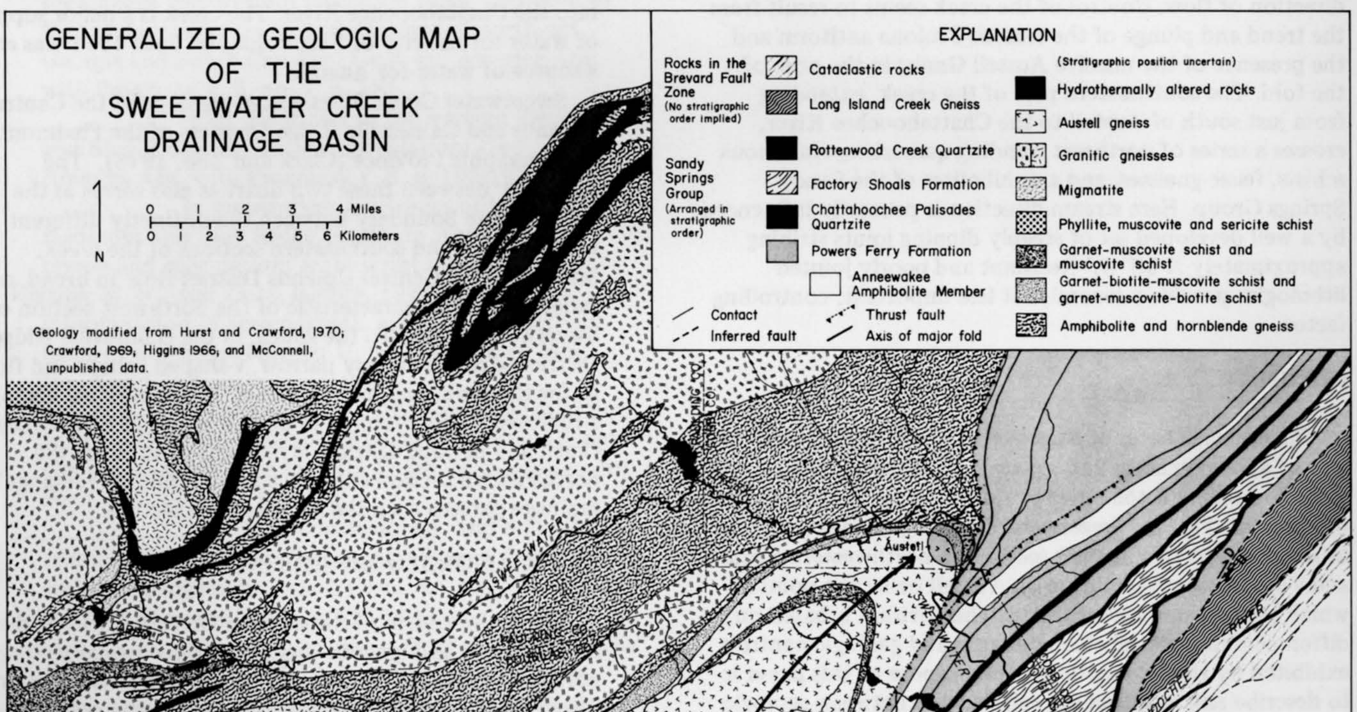

FIGURE 2. Geologic map of the Sweetwater Creek drainage basin.

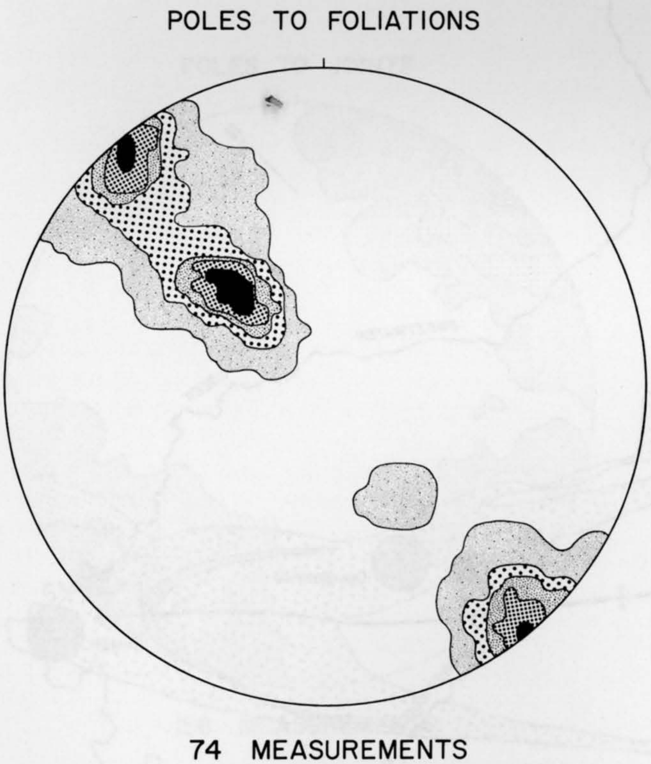

FIGURE 3. Equal-area projection of poles to foliation.

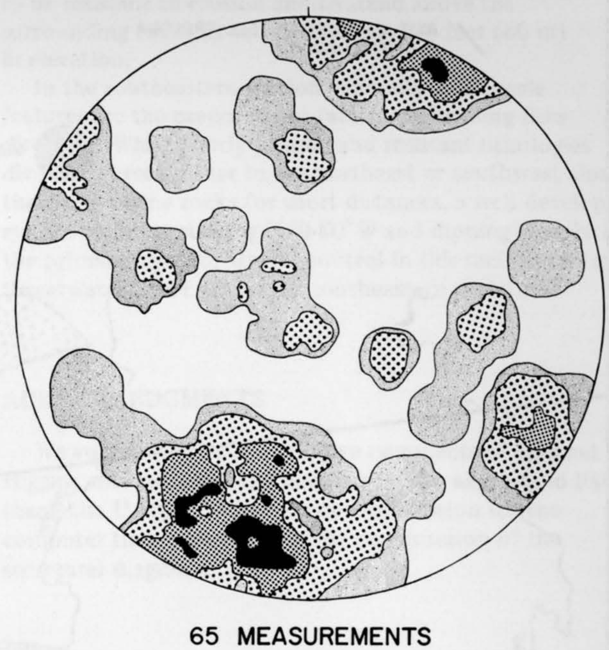

FIGURE 4. Equal-area projection of poles to jointing in the Austell gneiss.

RELATION OF STREAM FLOW TO STRUCTURE AND LITHOLOGY

Northwestern Section of Sweetwater Creek

Drainage in the northwestern section of Sweetwater Creek's drainage basin has a modified rectangular pattern with swampy areas and a wide stream valley. Outcrops along this section of the creek are relatively sparse, but where exposed, the rocks are usually moderately well jointed. Small scale structures, such as joints, appear to have little or no affect on stream flow in this section. However, the Austell-Frolona antiform plays a significant role in controlling stream flow direction. Sweetwater Creek and its tributaries have been unable to breach the ridge formed by the Austell-Frolona antiform near Austell. This ridge is approximately 100 feet (30 m) higher in elevation than the surrounding terrain including the quartzites of the Sandy Springs Group to the southeast. The apparent control of Sweetwater Creek by the Austell-Frolona antiform can be explained by the presence of the resistant Austell Gneiss in the nose of the antiform. The main factor influencing the resistance of the Austell Gneiss is its presence in the crest of an antiform. The fact that the gneiss' foliation is subhorizontal causes it to form pavement-type outcrops and weather less rapidly than the surrounding rocks.

A second factor relating to the Austell Gneiss' resistance to erosion is that jointing in the Austell Gneiss is irregular in orientation (Fig. 4) and discontinuous. Unlike the quartzites to the southeast, the Austell Gneiss does not have the consistently oriented, continuous joints to aid the erosion process.

The Austell-Frolona antiform and the presence of the massive Austell Gneiss in the nose of the fold combine to keep Sweetwater Creek flowing northeastward along its northwestern flank for approximately 9 miles (14 km). Sweetwater Creek is the only stream in this part of the Piedmont to flow northeastward for this distance. In contrast, to the southwest where the Austell Gneiss is absent, both the Dog River and Snake Creek are able to cross the axis of the fold and flow in a general southeastward direction (Fig. 5).

Southeastern Section of Sweetwater Creek

Sweetwater Creek follows the outline of the Austell-Frolona Antiform as it plunges beneath the surface at Austell. Bending around the nose of the antiform, Sweetwater Creek crosses the Chattahoochee Fault and the Blair Bridge fault and flows across the rocks of the Sandy Springs Group. In this section of the creek, the stream valley is deeper and more v-shaped and drainage becomes more rectangular in form.

Stream flow in the southeastern section of Sweetwater Creek is strongly controlled by well developed joints and by erosion resistant and (or) poorly to non-jointed lithologies. Lithologic control is due primarily to the presence of resistant quartzites and poorly jointed concordant pegmatites

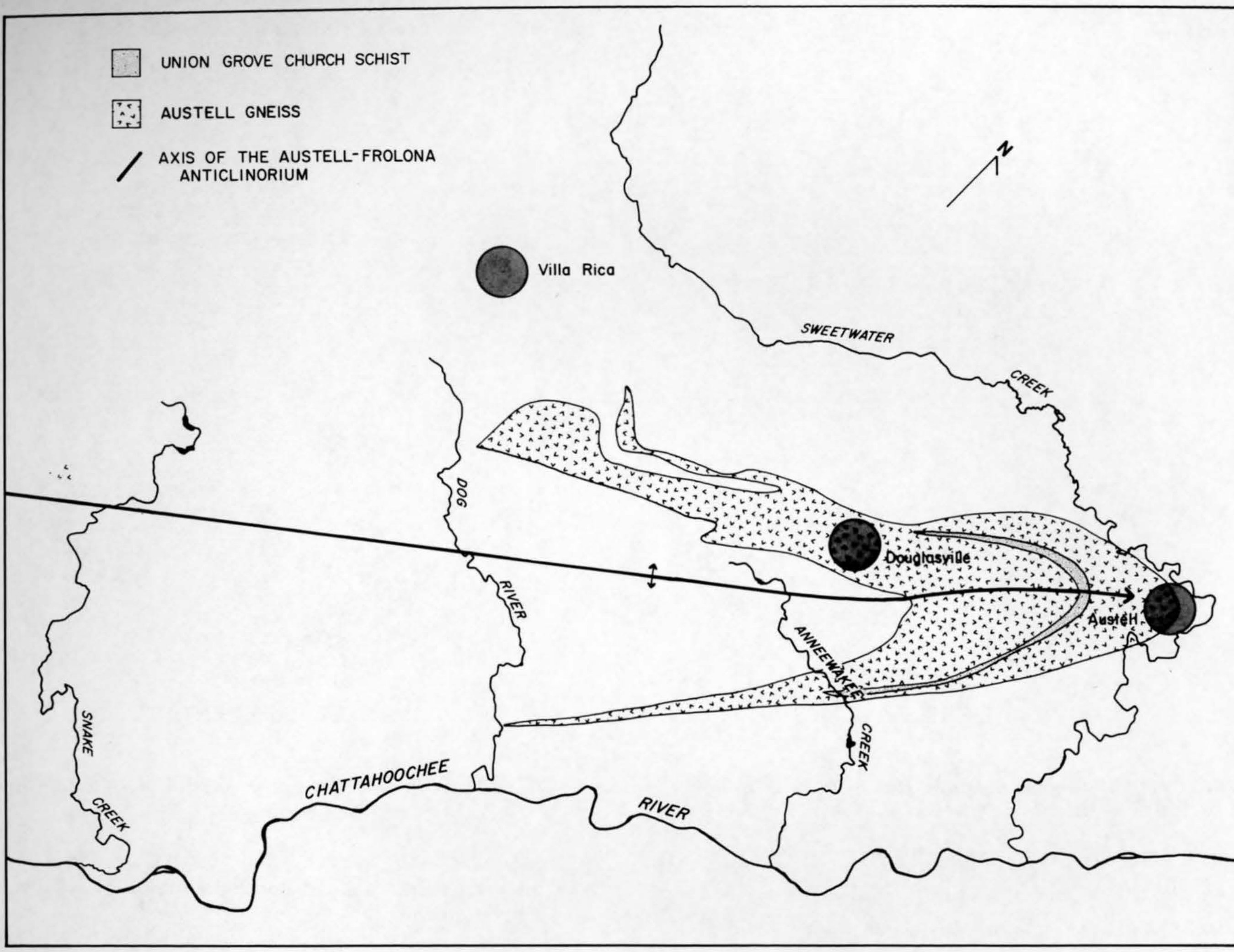

FIGURE 5. Relationship of Snake Creek and the Dog River to the Austell-Frolona antiform.

which divert stream flow to a direction parallel to the strike of the rocks. In this area lithologic banding and (or) bedding is parallel to the regional foliation which strikes approximately N50°E and dips moderately to steeply (Fig. 3). This type of stream control is first seen where the creek crosses a thick quartzite just southeast of Austell (Fig. 2). The quartzite diverts the creek along strike to the northeast until the creek is able to breach the resistant layer through well developed joints. Further downstream two more quartzites and several thick (approximately 5 feet or 1.5 m) pegmatites divert the creek to flow in a southwest direction.

The course of Sweetwater Creek shows excellent joint control between the areas where stream flow is diverted by resistant lithologies and (or) poorly jointed pegmatites. Flow direction is controlled by one well developed system of joints (Fig. 6) that strikes N30-60°W and dip steeply. A second poorly developed system of joints, striking approximately N70°E and dipping steeply to the northwest, does not appear to exert much control on stream flow.

In order to compare the control of stream flow by different structural elements in the southeastern section of Sweetwater Creek, rose diagrams of stream direction, jointing and foliation were constructed (Fig. 7). The predominant stream directions are from N10-50°W and N20-50°E (Fig. 7A). Jointing (Fig. 7B) has its direction maximum at N30-60°W and corresponds to the N10-50°W stream direction maxima. Foliation (Fig. 6B), which in this area is generally parallel to lithologic layering, has a maximum concentration at N40-60°E (Fig. 7C) which corresponds well with the N20-50°E maxima of stream direction.

These diagrams (Fig. 7) show that both jointing and erosion resistant lithologies play significant roles in controlling stream flow. However, it is evident from the dominant southeast trend of the creek in this section that jointing exerts the primary control on stream direction whereas resistant lithologies play a significant, but secondary role in control of stream direction.

SUMMARY

Sweetwater Creek provides an interesting contrast in tectonic control of streamflow direction. Both a major

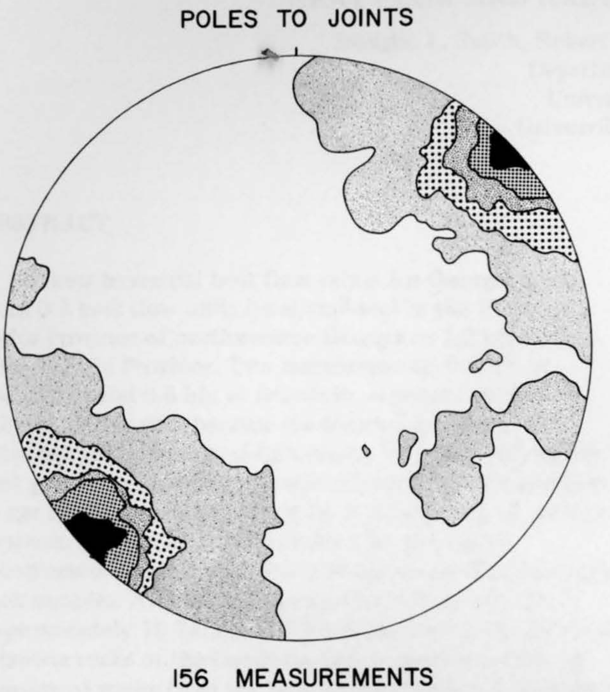

FIGURE 6. Equal-area projection of poles to jointing in the southeastern section of Sweetwater Creek.

structure, the Austell-Frolona antiform, and smaller scale features, jointing and lithology, control stream direction and water flow. In the northwestern section of the creek where small scale structures such as jointing do not appear to play a significant role in stream control, the Austell-Frolona antiform influences the direction of flow of Sweetwater Creek. Here the creek flows along the northwestern limb of the antiform in a northeastern direction for approximately 9 miles (14 km). Sweetwater Creek is the only creek in this part of the Piedmont to flow in this direction for a significant distance. This control of stream flow by the antiform results from the presence of the massive Austell Gneiss in the nose of the fold. The combination of its position in the nose of a plunging fold and its massive character cause the gneiss to be resistant to erosion and to stand above the surrounding rocks by approximately 100 feet (30 m) in elevation.

In the southeastern section relatively small scale features are the predominant factors controlling flow direction. While poorly jointed and resistant lithologies divert the creek either to the northeast or southwest along the strike of the rocks for short distances, a well developed system of joints striking N30-60° W and dipping steeply is the primary agent of stream control in this section and gives Sweetwater Creek its overall southeastern direction.

ACKNOWLEDGMENTS

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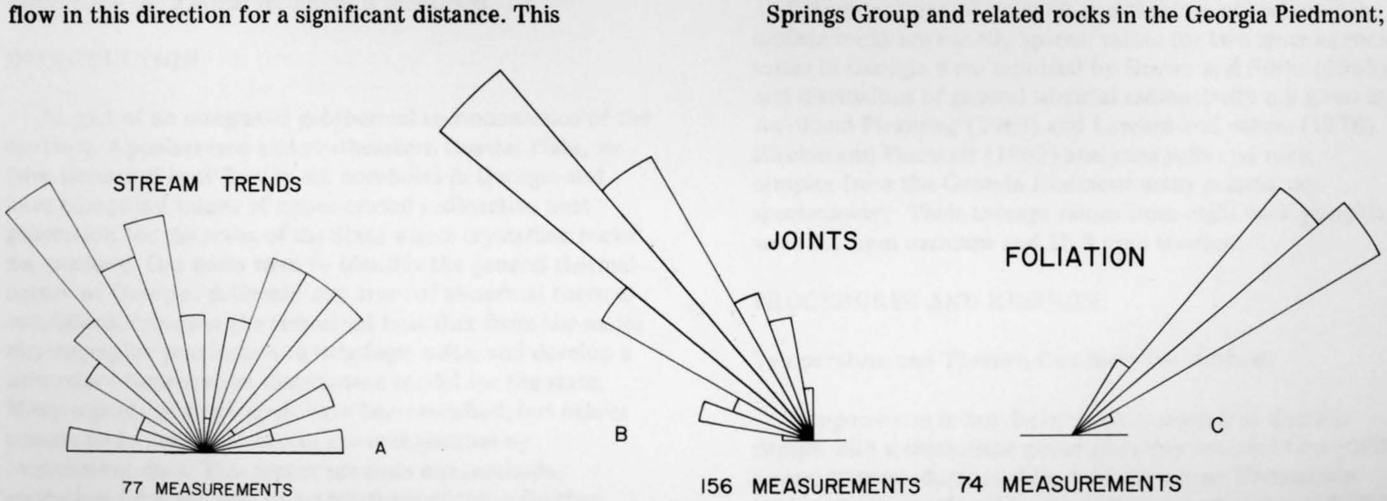

FIGURE 7. Rose diagrams of foliation, jointing and stream direction

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A THERMAL RECONNAISSANCE OF GEORGIA: HEAT FLOW AND RADIOACTIVE HEAT GENERATION

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ABSTRACT

Six new terrestrial heat flow values for Georgia range from 0.3 heat flow units (μcal/cm^2 sec) in the Valley and Ridge Province of northwestern Georgia to 1.2 hfu in the Coastal Plain Province. Two measurements, 0.5 hfu at Brunswick and 0.6 hfu at Braselton, represent minimum values at those sites because the thermal gradients demonstrate hydrological influences. Values of radioactive heat generation for various zones of the Piedmont and Blue Ridge Provinces are calculated from abundances of uranium, thorium, and potassium determined by gamma ray spectrometric analyses of over 130 igneous and metamorphic rock samples. Average heat generation values range from approximately 10-14 hgu (10^{-13}cal/cm^{3}sec) in the Paleozoic intrusive rocks of the Charlotte Belt to less than 4 hgu in samples of gneiss from the Blue Ridge Province. Comparisons of the heat flow and heat generation values yield a general linear relationship between the two variables, but fail to verify the reduced, or mantle heat-flow value of 0.8 hfu established for the central and northeastern United States. Instead, the values suggest that a portion of Georgia is thermally anomalous, and a separate thermal subprovince with a mantle heat flow of 0.3 hfu is identified. The low mantle heat flow is attributable to an upper mantle zone which is cooler, more mafic, and more radioactively impoverished than normal, and which may underlie much of the southern Appalachians. Simple temperature calculations for the crust indicate temperature variations at 30 km depth of at least 80°C and perhaps over 200°C among various portions of Georgia. No obvious evidence for exploitable geothermal resources is identified.

INTRODUCTION

As part of an integrated geothermal reconnaissance of the southern Appalachians and southeastern Coastal Plain, we have measured heat flow in six boreholes in Georgia and have computed values of upper-crustal radioactive heat generation for the areas of the State where crystalline rocks are exposed. Our goals were to identify the general thermal nature of Georgia, delineate any areas of abnormal thermal conditions, compare the terrestrial heat flux from the major physiographic provinces and lithologic belts, and develop a subsurface temperature distribution model for the state. Many aspects of these goals have been satisfied, but others remain to be accomplished or are complicated by inconsistent data. This report presents our methods, numerical findings, and interpretations of those findings.

Previous Work

Existing descriptions of geothermal conditions in Georgia are limited to an inclusion of the State in broad areas discussed on a regional basis. Roy and others (1968a, 1972) described the central and eastern United States as an area of "normal" heat flow with local anomalies attributable to excessive radiogenic heat production within the upper crust. Diment and others (1972) also discussed heat flow trends in the eastern United States and, although the preponderance of the data was from the northeastern United States, they presented similar values of approximately 1.2 heat flow units (1 hfu = 10^{-6} cal/cm^{2}sec). Conjectured heat flow contour patterns presented by Sass and others (1976) include western Georgia in an anomalous zone of less than 1.0 hfu, while eastern Georgia is portrayed as an area with a heat flow of 1.0 - 1.5 hfu.

Actual measurements in support of these suggestions, however, are limited. Determinations of heat flow at LaGrange and Griffin (Fig. 1) based on data from Birch and Spicer were reported by Diment and Robertson (1963). Values at both sites are 1.0 hfu, but should be considered as estimates because thermal conductivity values were estimated rather than measured. Very little additional information concerning subsurface temperatures in Georgia exists. Descriptions of thermal springs in Georgia generally place water temperatures at 23° to 30°C (McCallie, 1913; Hewett and Crickmay, 1937; Waring, 1965). Limited temperature measurements in boreholes in Glynn County (Wait and Gregg, 1973) were interpreted as indicative of temperature gradients of approximately 28°C/km.

Determinations of uranium and thorium abundances in surface rocks are equally sparse; values for two igneous rock suites in Georgia were reported by Brown and Silver (1955), and discussions of general surficial radioactivity are given by Auvil and Pickering (1969) and Lawton and others (1976). Ritchie and Plummer (1969) analyzed soils and rock samples from the Georgia Piedmont using gamma ray spectrometry. Their average values from eight rock samples were 3.7 ppm uranium and 15.8 ppm thorium.

PROCEDURES AND RESULTS

Temperature and Thermal Conductivity Methods

Temperatures in boreholes were measured at discrete depths with a thermistor probe assembly coupled to a 1000 meter, four-conductor cable. A Mueller-type Wheatstone bridge, similar to that described by Roy and others (1968b),

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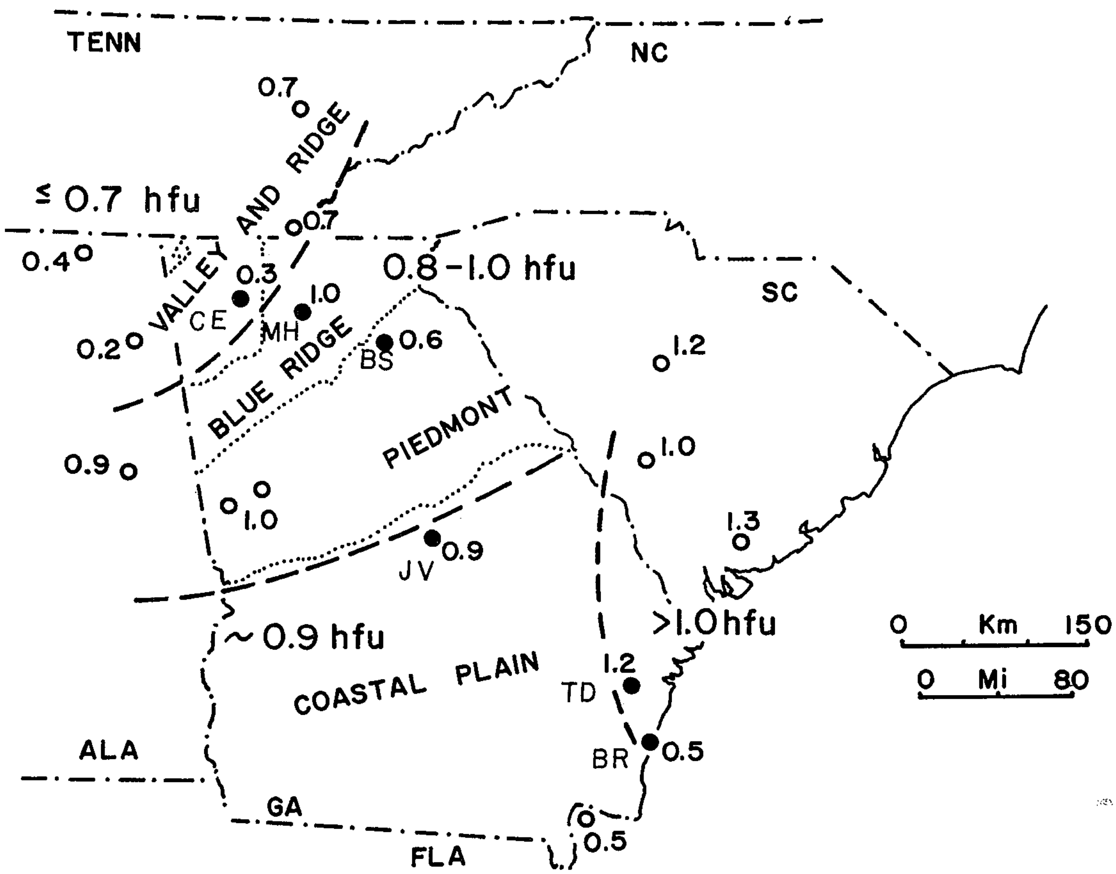

FIGURE 1. Distribution of heat flow values in and adjacent to Georgia. Closed circles represent new values reported herein. Key: CE, Crane Eater; MH, Marblehill; BS, Braselton; JV, Jeffersonville; TD, Townsend; BR, Brunswick. Open circles designate data from Diment and Robertson (1963), Roy and others (1968b), Diment and others (1965), Smith (1976), and Smith and others (1977). Values are given in heat flow units and generalized heat flow contours are shown as dashed lines. Boundaries between physiographic provinces are from Hatcher (1972).

with an Electroscientific Industries six-decade variable resistor and a Leeds and Northrup 2437 null detector was used to match thermistor resistances. The system was calibrated at 10° intervals from 10° to 40°C using a Hewlett-Packard quartz thermometer. The estimated accuracy of individual temperature measurements is better than ± 0.1°C. The accuracy of successive temperature differentials used for gradient values is considered to be much better, perhaps 0.01°C/km.

Temperature gradients were corrected for local topographic variations where necessary by using a modification of the method developed by Birch (1950). The differences between average elevations in geographic rings and the surface elevation of a drill hole were determined from topographic maps. Figure 1 shows the location of gradient measurements in Georgia and parts of adjacent states.

Thermal conductivity analyses of rock samples from boreholes were made with a divided-bar apparatus similar to that described by Sass and others (1971a). The instrument consists of 3.81 cm diameter upper and lower assemblages of 0.317 cm thick copper disks enclosing 0.635 cm thick lexan disks. Circulating baths were used to establish a temperature differential of approximately 10°C across the specimen and maintain its temperature to within 10°C of its *in situ* value. Copper-constantan thermocouples were inserted in the copper disks to determine temperature values. All core samples were machined to 2.54 cm by 3.81 cm diameter cylinders with a tolerance of 0.0005 cm, vacuum saturated with water, and coated with silicone grease to facilitate thermal contact. Measurements were made with the samples under an axial pressure of 50 to 100 bars.

Thermal conductivity values were determined on drill-hole cuttings from five sites for which core samples were not available. The method employed was similar to that described by Sass and others (1971b). Cuttings were packed into plastic-walled cells (1.59 cm by 3.81 cm diameter) having machined copper bases. A known volume fraction of water was added to saturate the cell, and a conductivity value for the cell as a whole was determined using the divided-bar apparatus. The conductivity of the rock fragments was then isolated by calculating the influences of the plastic walls, the volume fraction of the water, and the natural porosity of the rocks according to Sass and others (1971b). Mean harmonic conductivity values are listed in Table 1.

Heat Flow Sites

Terrestrial heat flow values were calculated for each measurement site as the product of the least-squares temperature gradient and the mean harmonic conductivity. Figure 2 displays the temperature depth relationships for each site. The new values (Table 1) are for three sites in the Coastal Plain Province, two in the crystalline rock areas north of the Fall Line, and one site in folded rocks of Paleozoic Age in the northwestern corner of Georgia.

Holes were logged in the Coastal Plain at Brunswick (Glynn County), Jeffersonville (Wilkinson County), and near Townsend (Long County). In Brunswick, the 350 m deep U.S.G.S. water-monitoring well No. 19 displayed a very erratic temperature distribution with depth (Fig. 2) indicative of subsurface water flow. The well is near the coastline and three separate levels of water flow with distinctive gradients can be surmised. An overall average gradient value was used for computations. Contrasting gradient values were also observed in the 350 m deep oil test well near Jeffersonville. The temperature gradient is 8.16°C/km in the interval from 80 to 140 m, but jumps to 15.53°C/km between 140 and 300 m. The deeper gradient value was used to determine the heat flow in order to avoid possible hydrologic effects at shallower depths. The observed gradient in a 260 m deep U.S.G.S. water-monitoring well at a cemetery near Townsend is linear, except over the depth interval 160-200 m.

Neither core nor cutting samples were available for any of the Coastal Plain holes; consequently, cuttings from nearby holes with representative lithologies were utilized for conductivity determinations. Measurements were made of cutting from GGS 363 in Liberty County and GGS 3145 in Wayne County for the Townsend site, from GGS 3147 in Twiggs County for Jeffersonville, and from GGS 362, five miles SSW of Brunswick, for the Brunswick site.

A 370 m deep exploration hole (Georgia Marble Company No. 66-120) at the marble quarry near Marblehill in Pickens County displayed a linear temperature profile with a gradient of 15.94°C/km. Core samples from the same hole were used for conductivity determinations. This hole is in the Murphy Marble belt of the Blue Ridge Province.

In the Piedmont Province, a relatively shallow (146 m) unused water well was logged just west of Braselton in Jackson County. The temperature profile shows a relatively low gradient (8.46°C/km) over the upper 130 m, but three data values in the lower portion (130-146 m) of the hole show a gradient of over 36°C/km. It is very probable that flowing water has disturbed the true gradient at this site, and neither value is valid. Although cutting chips collected at the site were probably from the lower segment (first hard rock encountered), a weighted average gradient (10.68°C/km) was estimated as representative. This has resulted in a

TABLE 1. Temperature gradients (Γ), mean harmonic thermal conductivity values (K) from N samples, and surface heat flow (Q) values.

Location	N. Lat.	W. Long.	depth (m)	Γ(C /km)	N	K (mcal/cm sec°C)	Q(hfu)
Crane Eater (near Calhoun, Gordon Co.)	34°32′	84°52′	200 m	10.05*	14	3.425 ± .264	0.34 ± .03
Braselton, Jackson Co.	34°5′	83°46′	146 m	10.68	14	6.015 ± .497	0.64 ± .04
Brunswick, (USGS Test Well 19) Glynn Co.	31°8′	81°30′	350 m	12.24	9	4.163 ± .888	0.51 ± .16
Jeffersonville, Wilkinson Co.	32°43′	83°15′	300 m	15.53	8	5.899 ± .947	0.92 ± .05
Marblehill, (Georgia Marble Co. No. 66-120) Pickens Co.	34°25′	84°21′	370 m	15.94*	12	6.291 ± .936	1.00 ± .04
Townsend, Long Co.	31°36′	81°36′	260 m	33.41	13	2.710 ± .536	1.24 ± .06

*Corrected for local topography

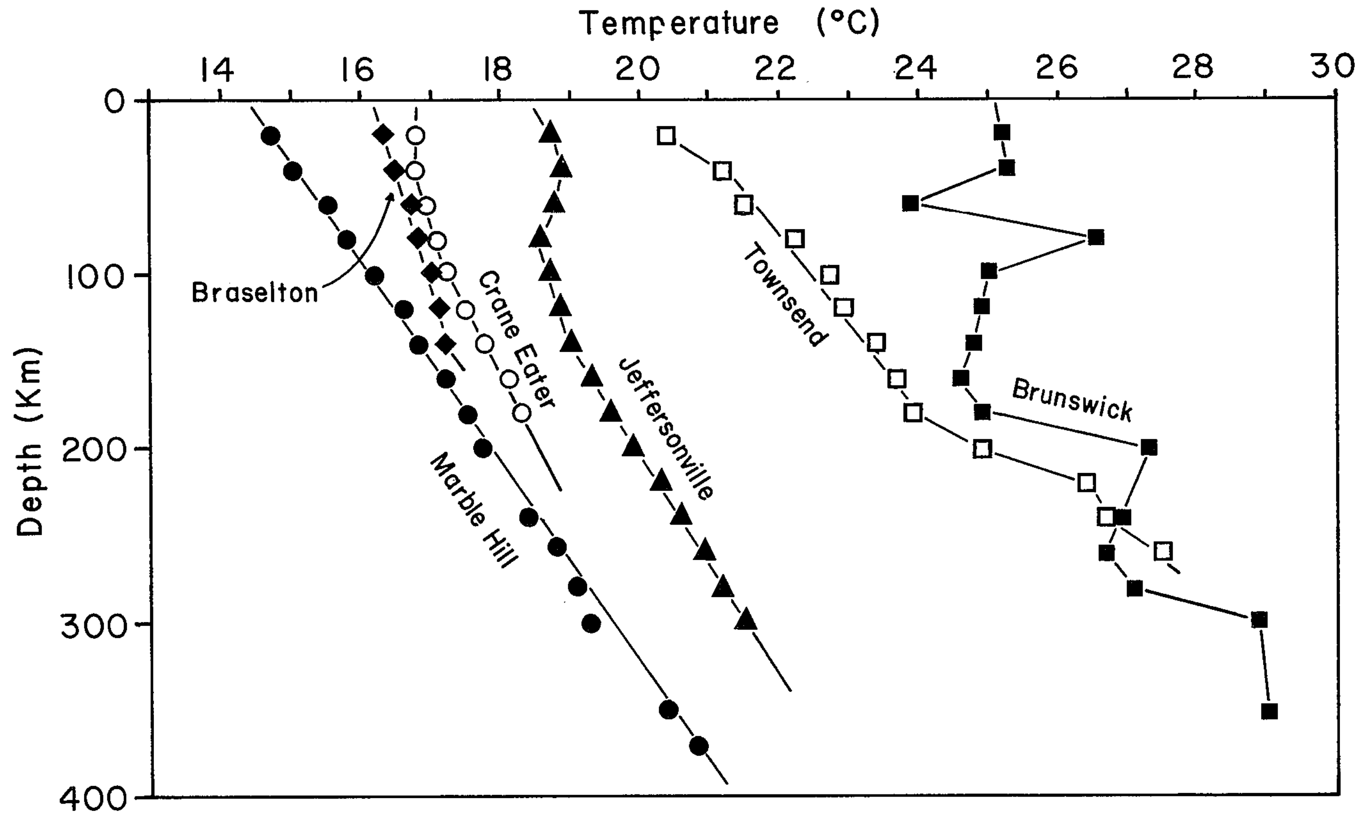

FIGURE 2. **Temperature profiles recorded in Georgia boreholes. Gradient values are reported in Table 1.**

considerably lower heat flow value than might be acquired by using the chip conductivity and the high gradient at the bottom of the hole.

One hole in the Valley and Ridge Province, an oil test well at Crane Eater in Gordon County, was logged. The gradient changed abruptly from 7.58°C/km over 60 to 120 m to 13.95°C/km over 120-200 m, and an overall average value was assigned as representative. Cutting chips from unknown depths and orientations in the hole were collected at the site for conductivity measurements. No other samples were available.

The most reliable of the new heat flow values listed in Table 1 are, we believe, those for the Marblehill, Townsend, and Jeffersonville sites. The obvious distortion of the thermal gradient at Brunswick by groundwater movement relegates the 0.5 hfu heat flow value for that site to a minimum value. Distortion of the heat flow by groundwater was also observed 50 miles to the southwest at Boulougne, Florida, where a value of 0.5 hfu was measured (Smith and Fuller, in press).

Marginal conditions (no casing in the hole, core cuttings for conductivity measurements, probable groundwater disturbance) at the Braselton site (Fig. 1) lead us to assess the calculated heat flow value for that area as a minimum value. The higher temperature gradient at the base of the drill hole where the core cuttings used for conductivity measurements originated cannot be quantitatively evaluated, but probably should be considered as indicative of a slightly higher flux than that computed. In the same sense, the 0.3 hfu value for Crane Eater in the Valley and Ridge Province is considered an estimate because shale chips from the drill site were used for conductivity measurements, but no information concerning the depth, extent, or orientation of the shale is available for the measurement site. Anomalously low heat flow values are also observed in the Valley and Ridge and Appalachian Plateau Provinces of northern Alabama (Smith and others, 1977); thus the Crane Eater value, although an estimate, is probably representative of the area.

Radioactive Heat Generation

A total of 112 composite one-kg samples of igneous and metamorphic rocks representative of local exposures were collected throughout the crystalline areas of northern Georgia. Collection and preparation procedures were outlined by Garvey (1975). The samples were analyzed by gamma ray spectrometry for abundances of the heat producing elements uranium, thorium, and potassium (Table 2). The spectrometer system consisted of a Harshaw 6-inch diameter NaI(Tl) detector crystal optically coupled to a photomultiplier tube and housed in a 4-inch thick lead-brick well with a rail-mounted, rolling lead-brick top. Scintillation signals were recorded by a Nuclear Data 100 multi-channel analyzer and later transferred to magnetic tape.

Calibration was achieved with small amounts of cobalt-60 and cesium-137, and standard samples with certified abundances of uranium and thorium were acquired from the New Brunswick A.E.C. Laboratory for comparison against

TABLE 2. Average abundances of radioactive elements and heat generation values for 1 kilogram composite samples of surficial rocks in Georgia.

Area	Rock Types	No. Samples	U ppm	Th ppm	K%	A (hgu)
Paleozoic Plutons						
(Elberton)	granite	5	6.8	51.6	4.7	14.1
(Danburg)	granite	4	4.6	34.6	4.5	9.7
(Sparta)	granite	3	4.0	18.6	4.0	6.7
(Siloam)	granite	2	6.8	38.5	4.6	11.8
(Stone Mountain)	quartz monzonite	4	4.5	6.4	4.6	4.9
(West Central Georgia)	granite	6	4.3	26.4	3.4	12.9
Inner Piedmont						
(eastern)	granite gneiss	23	4.2	18.5	3.9	6.4
(western)	granite gneiss	15	2.7	12.3	2.9	4.4
Charlotte Belt						
(eastern)	granite gneiss	15	3.3	18.5	3.8	5.8
(western)	hornblende gneiss	6	1.5	15.8	4.6	4.7
Blue Ridge						
(Near Brevard Fault Zone)	biotite gneiss & granite gneiss	6	4.5	15.6	3.4	6.2
(north eastern Counties)	mica gneiss	24	1.9	9.7	2.0	3.6

unknowns. Calculation of radio-isotope abundances was accomplished with a 256-channel, least-squares computer program for three unknowns modified from Salmon (1961). Heat generation values were calculated using Birch's (1954) heat production coefficients for uranium, thorium, and potassium. Figure 3 shows the distribution of sample collection sites and Table 2 summarizes the values in heat generation units (10^{-13}cal/cm^{3}sec). A complete listing of individual sample values is given in Garvey (1975) and Gregory (unpub. data).

DISCUSSION

Heat Flow Values

It is difficult to identify specific heat flow trends from the reported values because of the uncertainty associated with some of the values and the general paucity of data. Upon consideration of previously determined heat flow values for the southeastern United States (Diment and others, 1972; Smith, 1976), however, we can fix rudimentary heat flow divisions within Georgia. Figure 1 depicts four surficial heat flow regions for Georgia, each based on a limited number of values within the state and extending to areas of adjacent states with congruent data.

The area of relatively low (<0.7 hfu) heat flow in northwestern Georgia includes the Valley and Ridge sequence and that part of the Blue Ridge Province northwest of the Murphy Belt. Low heat flow values at Ducktown, Tennessee (0.7 hfu), and Calhoun County, Alabama (0.2 hfu), substantiate the presence of this thermal region. The Piedmont Province and most of the Blue Ridge Province are grouped together in a thermal region considered to be typical of the southern Appalachians with heat flow values of 0.8 - 1.0 hfu. Two anomalously low values of heat flow from Charlotte and Winston-Salem, North Carolina (Emhof, 1977), suggest separate conditions for the Charlotte Belt; however, no data are available from the Charlotte Belt in Georgia and it is tentatively considered thermally similar to the Inner Piedmont.

A gradational boundary for heat flow is proposed (Fig. 1) at the Fall Line between the Piedmont and western Coastal Plain. Rock types dissimilar to those of the Piedmont may form the buried basement for much of the Coastal Plain (Milton and Hurst, 1965); consequently thermal conditions might be expected to vary from the Piedmont to the Coastal Plain. The heat flow region with $\sim$0.9 hfu extends to the southwest to include similar values in southern Alabama and north Florida (Smith and Fuller, 1976; Gregory, unpub. data).

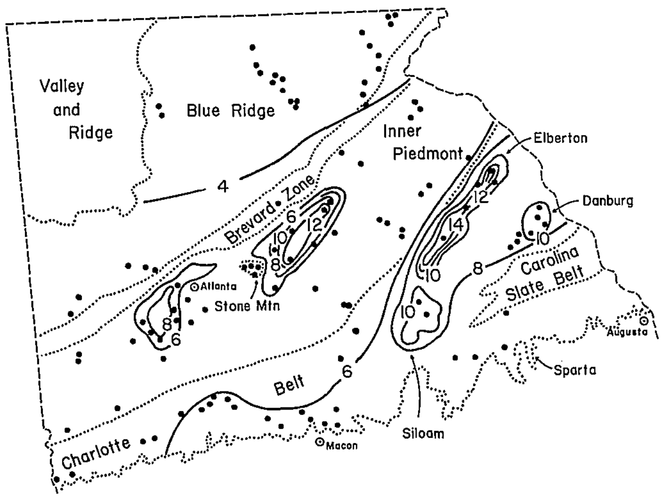

FIGURE 3. Distribution of sample locations for radioactive heat generation measurements and generalized contour lines of average values. Units are heat generation units.

The 1.2 hfu value at Townsend, however, justifies an extension into eastern Georgia of a relatively high heat flow region identified in the North Carolina and South Carolina Coastal Plain by Smith and others (1977). This disparity may be a result of contrasting basement rock types or crustal thickness variations.

Distributions of Radioactive Elements and Heat Production

The numerous surface rocks sampled for analyses of radioactive heat-producing elements represent the diverse array of crystalline rocks exposed in the Piedmont and Blue Ridge areas of Georgia. A wide range of results was obtained among individual samples, and average values for groupings of major rock units or areas (Table 2) demonstrate regional variations of radioelement abundances and heat generation.

Higher abundances of radioelements appear in the granitic intrusives of the Charlotte Belt than in the more abundant metamorphic rocks. Five composite samples from the Elberton Batholith average 6.8 ppm uranium, 51.6 ppm thorium, and 4.7 percent potassium. Lower average abundances are observed in the other plutons (Table 2) in the eastern portion of the Charlotte Belt, but six samples from various igneous intrusives of the west-central portion of the Piedmont average 4.3 ppm uranium, 26.4 ppm thorium, and 3.4 percent potassium. These abundances are roughly consistent with values from the Sparta and Danburg Plutons and suggest that no regular variation in radioelement abundance exists among igneous intrusives of different belts in the Piedmont.

Analyses of samples from the Stone Mountain Pluton near Atlanta in the Piedmont Province, however, suggest that it is chemically distinct from the other Georgia plutons. Ratios of thorium to uranium for the Charlotte Belt Plutons and those from the west-central Piedmont range from 7.6 to 4.6 (somewhat higher than the world average value of 3.7), but four Stone Mountain rocks show an average Th/U value of 1.4. A fifth sample (19.2 ppm uranium, 6.7 ppm thorium, and 4.6 percent potassium) has a Th/U value of 0.3, similar to the 0.5 value derived from data reported by Ritchie and Plummer (1969). If Stone Mountain is the result of anatexis (Whitney and others, 1974), the uranium and thorium abundances are related to abundances in the country source rock, and low Th/U values can be attributed to preferential mobilization of uranium during the melting process. Heat generation from the Stone Mountain granite is lower (4.9 hgu) than that observed in the other plutons samples (6.7 to 14.1 hgu).

Average radioelement abundances for the metamorphic rocks analyzed are, in general, less than those for the igneous samples. Rock types sampled in the Inner Piedmont were primarily biotite gneiss and granite gneiss. Heat generation values are slightly higher in the eastern Inner Piedmont than in the west (6.4 hgu vs. 4.4 hgu), but still somewhat lower than the igneous rock values. Six samples of biotite gneiss and granite gneiss from the controversial area north of the Brevard Zone (described as Blue Ridge by Hatcher (1972); described as Piedmont by Crawford and Medlin (1973)) averaged 6.2 hgu and are thermally similar to those of the Piedmont. Fifteen metamorphic rock samples, primarily granite gneiss, from the Charlotte Belt averaged 5.8 hgu and six samples of hornblende gneiss from the western end of the Charlotte Belt averaged 4.7 hgu. Heat generation in the Charlotte Belt appears to be similar to that in the Inner Piedmont with each belt demonstrating a westerly decrease. Eighteen composite samples of mica schist and various gneisses from the Blue Ridge of northeastern Georgia (primarily Murray, Fannin, Union, and Lumpkin counties) average significantly lower radioactive heat production (3.6 hgu) than the crystalline rocks of the Piedmont.

Figure 3 depicts a generalized contour pattern for radioactive heat generation in the Georgia Piedmont and Blue Ridge. Higher values in the Piedmont can be correlated with larger granitic plutons while the lowest values (2 to 4 hgu) are more representative of the Blue Ridge rocks. Although the distribution of values in Figure 3 does not preclude an assemblage of continental fragments about the Brevard Zone (Fullagar and Butler, 1977), we prefer to attribute the major heat production variations (exclusive of those resulting from Paleozoic intrusives) to compositional changes and erosive levels in a crust marked by a decreasing radioelement abundance with depth (Lachenbruch, 1968). Thus, the exposed rocks in the Blue Ridge represent an older and deeper assemblage than those of the Piedmont.

Mantle Heat Flow

Determinations of radioactive heat generation in the surficial crystalline rocks of northern Georgia permit calculations of reduced heat flow - that part of the surficial heat flow not generated in the upper crust, but considered to be indicative of flux from the mantle. Birch and others (1968), Roy and others (1968a, 1972), and Lachenbruch (1968) found a linear relationship between

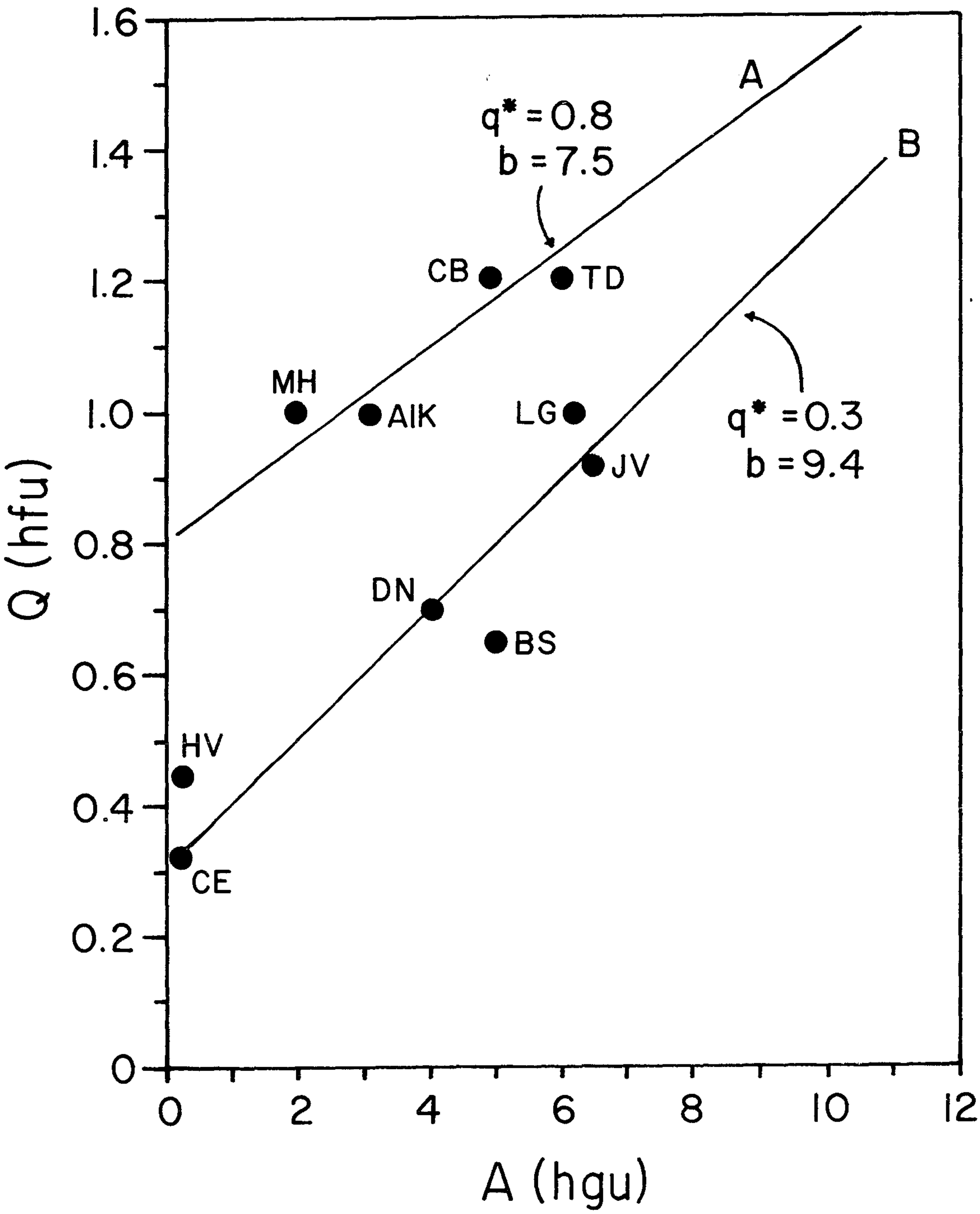

FIGURE 4. Plot of surficial heat flow and radioactive heat generation for Georgia and surrounding areas. Line A is from Roy and others (1968a), and represents the "normal" eastern United States thermal province ($q^* = 0.8$ hfu; $b = 7.5$ km). Line B is constructed to fit the surrounding data points and indicates a separate thermal subprovince with $q^* = 0.3$ hfu and $b = 9.4$ km. Key: MH, Marblehill; AIK, Aikin; CB, Columbia; TD, Townsend; LG, LaGrange; JV, Jeffersonville; BS, Braselton; DN, Ducktown; HV, Huntsville; CE, Crane Eater; CG, Colvin Gap (Ala.).

surficial heat flow and radioactive heat generation, and demonstrated that reduced heat flow values can be defining parameters for thermal provinces. As expressed by Roy and others (1972), the reduced heat flow is

$$q^* = Q - bA \qquad (1)$$

where Q is the surface heat flow and A is the heat generation of surface rocks as determined from radioelement abundances. The parameter b has units of distance and is associated with the possible thickness of a crustal layer with characteristic radioactivity, A. The definitive parameters for the Central Plains - Eastern United States thermal province are $q^* = 0.8$ hfu and $b = 7.5$ km (Roy and others, 1972). These data, however, are derived primarily from measurements in the northern Appalachians.

As a test of the linear relationship between heat flow and surface rock heat generation and its applicability to the southern Appalachians, values from Georgia and adjacent areas are plotted in Figure 4. The heat generation values, A, used for the Georgia sites are averages of U, Th and K analyses from rocks in the immediate area of each of the heat flow sites. Because the Crane Eater site was in the Paleozoic sediments of the Valley and Ridge Province, a negligible value is assigned that site. Heat generation values associated with the Coastal Plain measurements at Townsend and Jeffersonville are estimated from the trends of Figure 3.

Several of the values in Figure 4, including those for Townsend and Marblehill in Georgia, fall near the line established for the eastern United States by Roy and others (1972), but the remainder appear to describe a separate linear relationship with a reduced, or mantle, heat flow of 0.3 hfu and a "b" value of 9.4 km. These data provide evidence for a thermal subprovince of relatively low heat flow that is distinct from the "eastern United States" province.

The q^* and b parameters are similar to those for the Sierra Nevada province ($q^* = 0.4$ hfu and $b = 10$ km) as proposed by Roy and others (1972), but must be regarded as only approximate because of the limited number of values. Indeed, if the intercept (or mantle heat flow) of line B in Figure 6 is approximately 0.3 hfu and the slope (or b value) is approximately 7.5 km or parallel to line A, then an argument for some subsurface thermal loss equivalent to 0.5 hfu might be advanced. This could be based on a uniform 0.8 hfu flowing from the mantle into the southern Appalachian crust, but a portion of that amount would be deviated elsewhere. Reiter and Costain (1973) discussed the possibility of thermal refraction at depth to explain contrasts in heat flow values in Virginia. Reliable models of the lower crust and upper mantle for the southern Appalachians and Coastal Plain are elusive, precluding exact calculations. The higher heat flow values reported for the North and South Carolina Coastal Plain Province (Smith and others, 1977; Ziagos and others, 1976), however, suggest that this area is (1) not a portion of the subprovince defined by Figure 4 and (2) perhaps a recipient of heat refracted from the immediate interior of the continent.

An alternative interpretation for the implied subprovince may evolve from the simplistic approach that lower crust and upper mantle thermal conditions are not uniform on a regional basis (eastern United States), but that areas of local extent (e.g. southern Appalachians) may display characteristic properties. Combs and Simmons (1973) have outlined evidence for thermal subprovinces in the central United States, and propose subtle, but basic, differences in the mantle heat flow. The abundant occurrence of anorthosite bodies of impoverished radioactivity in the northern Appalachian crust has been cited (Diment and others, 1972) as a possible reason for locally lower heat generation and surface heat flow in the northern Appalachians. No evidence of such plutons exists in the southern Appalachians.

Temperature Distributions

Calculations of subsurface temperatures are based on the linear relationship between surface heat flow and heat generation providing some reasonable estimates of thermal conductivity and heat generation exist for the subsurface. Lachenbruch (1968; 1970) has mathematically demonstrated an exponential decrease of radioactive heat generation within the crust with b (slope in Fig. 4) as an exponential decrement. If we assume a constant thermal conductivity, K, with depth, then the relationship between temperature, T, and depth, z, becomes

$$T(Z) = To + \frac{q^*Z}{K} + \frac{Aob^2}{K}\left[1 - e^{-z/b}\right] \qquad (2)$$

where T_o is the mean annual surface temperature and A_o is the surface rock radioactivity. Using equation (2) and setting $T_o = 10^\circ C$ and $K = 6$ mcal/cm$^\circ$C sec, we have calculated separate temperature distributions for typical surface radioactivity values of Georgia and the q^* (0.3 hfu) and b (9.4 km) values from Figure 4. Three of the four distributions (Table 3; Figure 5) correspond to surface radioactivity values of 4 hgu (Blue Ridge), 6 hgu (Piedmont metamorphic rocks), and 10 hgu (Paleozoic intrusive rocks).

Curves A, B, and C of Figure 5 represent temperature profiles for the various configurations of upper-crustal heat generation in Georgia that are within the thermal subprovince of anomalously lower mantle heat flow. We estimate this area to include the Valley and Ridge, the southwestern Blue Ridge, the Inner Piedmont, and perhaps the western Coastal Plain of Georgia. Curve D (Fig. 5) represents the standard temperature distribution of Roy and others (1972) for the eastern United States. Based on the distribution of data in Figure 4, this curve would be applicable to the northeastern Blue Ridge area, the northeastern Coastal Plain area (adjacent to South Carolina), and perhaps the younger intrusives of the Charlotte and Carolina Slate Belts. The curves suggest lower crustal temperatures (30 km depth) ranging from approximately 200°C in extreme northwestern Georgia to over 400°C for the Coastal Plain Province near Savannah.

Although the complex composition of the southern Appalachians seemingly resists a simple pattern of thermal

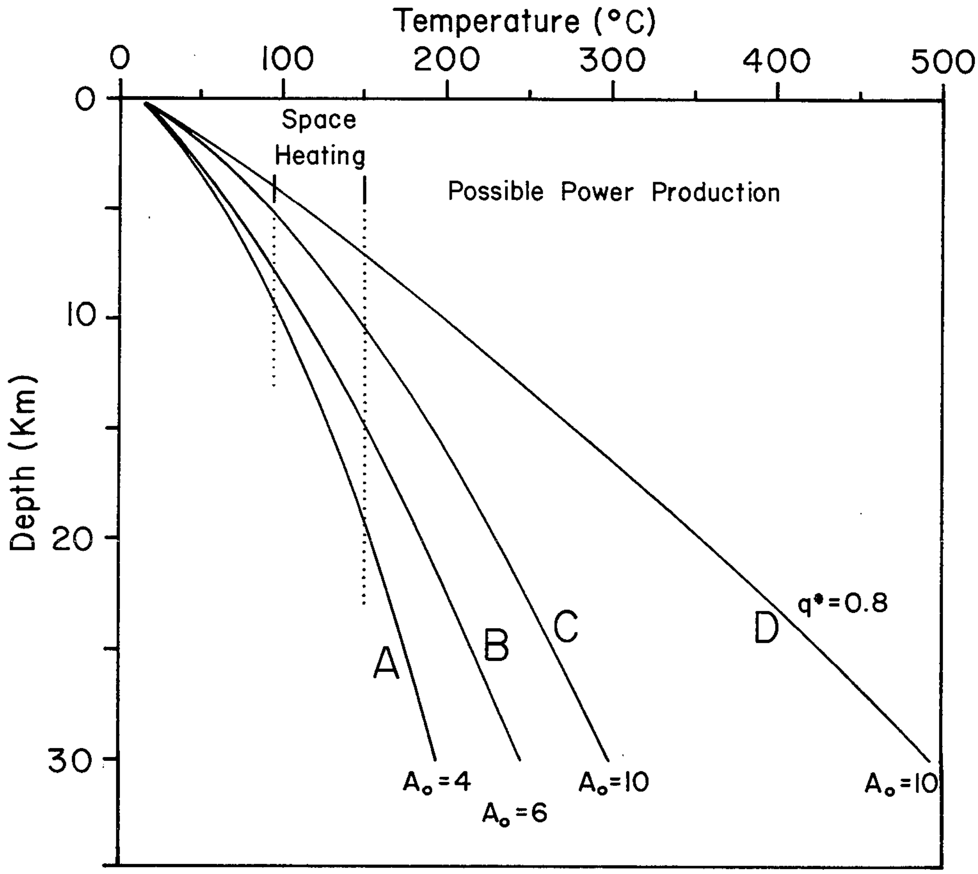

FIGURE 5. Plot of temperature calculations for the Georgia subsurface using a constant thermal conductivity of 6 mcal/cm sec°C and a surface temperature of 10°C. Curves A, B, and C are based on a reduced heat flow of 0.3 hfu and correspond, respectively, to typical surface rock heat generation values of 4 hgu (Blue Ridge), 6 hgu (Piedmont metamorphic rocks), and 10 hgu (Paleozoic intrusive rocks). Curve D (from Diment et al., 1975) represents the temperature distribution for the normal eastern United States thermal province (Roy and others, 1968a) with $q^* = 0.8$ hfu and $A_0 = 10$ hgu. Economically useful temperature ranges for space heating and possible power production (Diment and others, 1975) are superimposed to illustrate probable depths for geothermal resources.

models and some compromises among the curves of Figure 5 may be in order, the lower subsurface temperatures associated with the anomalous mantle heat flow of 0.3 hfu imply a more mafic, radioactivity-impoverished, and cooler upper mantle under much of northern Georgia. The extreme subsurface temperature contrast between areas of low surface heat generation in the Blue Ridge Province and the areas of higher, more normal heat flow on the Coastal Plain Province can be emphasized by noting (Table 3) the 20 km depression of isotherms necessary under the Blue Ridge and Valley and Ridge Provinces.

Geothermal Energy Potential

No obvious sources of geothermal energy in Georgia have been encountered in this study. Temperature ranges (from Diment and others, 1975) to accommodate limited space heating and power production are marked on Figure 5. Useful temperatures in Georgia, according to our theoretical calculations, lie several (4 to 8) kilometers below the surface and would seem to economically preclude drilling for geothermal purposes. The presence of thermal springs in western Georgia (Waring, 1965) is probably related to the Towaliga Fault Zone, but typical temperatures are not excessive (30°C) and are probably a result of water circulation through only 1 to 2 km.

CONCLUSIONS

Six new heat flow values for Georgia augment two previously determined values. The new values range from 0.3 hfu in the Valley and Ridge Province of northwestern Georgia to 1.0 hfu and 0.6 hfu values in the Blue Ridge and Piedmont Provinces, respectively, and to a high of 1.2 hfu in the Coastal Plain Province. Several of the measurements show influences of circulating groundwater and are, therefore, only estimates.

The data do not show a sharp contrast of heat flow values along boundaries between physiographic provinces, but do substantiate the concept of a region in the southern Appalachians of relatively low heat flow. A paucity of values in the Piedmont prevent our interpretation of

TABLE 3. Temperature calculations for the Georgia subsurface using a constant thermal conductivity of 6 mcal/cm sec°C and a reduced heat flow (q^*) of 0.3 hfu. Surficial radioactive heat generation (A_0) is in hgu and is considered to decrease exponentially with depth. The exponential decrement is b = 9.4 km. A surface temperature of 10°C is assumed throughout.

Depth (km)	Blue Ridge ($A_0 = 4$)	Piedmont ($A_0 = 6$)	Paleozoic Intrusives ($A_0 = 10$)
1	21	24	30
2	31	37	48
4	50	60	80
6	68	82	109
8	84	100	134
10	98	117	156
20	162	188	240
30	217	245	301

possible thermal differences among the various Piedmont Belts.

Measurements of uranium, thorium, and potassium abundances in over 100 surficial igneous and metamorphic rocks by gamma ray spectrometry have resulted in calculations of the radioactive heat generation in the upper crust. The greatest radioactive heat generation, averaging over 10 heat generation units, is found in the Paleozoic igneous intrusives of the Charlotte Belt and Inner Piedmont. The metamorphic rocks of the Piedmont range typically from 4 to 8 hgu, while the lowest heat production (3.6 hgu) is found in the Paleozoic igneous intrusives of the Charlotte Belt and Inner Piedmont. The metamorphic rocks of the Piedmont range typically from 4 to 8 hgu, while the lowest heat production (3.6 hgu) is found in the Blue Ridge Province. Thorium to uranium ratios in the igneous rocks are generally 4 to 8, somewhat higher than the world-wide average. Stone Mountain samples, however, have an anomalous Th/U value of 1.4, supporting the theories of an anatectic origin for that pluton.

Comparison of the heat flow and heat generation data tends to verify a rough linear relationship between the two variables, but fails to confirm a definite continuity of the eastern United States thermal province throughout the southern Appalachians. Instead, a subprovince with a mantle heat flow of 0.3 hfu (vice 0.8 hfu) and correspondingly lower than average surface heat flow is suggested. The anomalous thermal conditions which constitute the thermal subprovince are enigmatic, but may be a result of thermal refraction across layers of contrasting conductivity at depth, variable crustal thicknesses, or discontinuous mantle compositions. No superior model can be developed without additional thermal evidence and better concepts of the structural nature of the lithosphere of the southern Appalachians.

Temperature calculations for the Georgia subsurface indicate deep crustal values (at 30 km) ranging from 217°C in the Valley and Ridge Province to 300°C beneath the Paleozoic intrusives. Temperatures of over 400°C may be expected at 30 km depth in certain areas of the Coastal Plain Province. No conditions favorable to the exploitation of geothermal energy under present technologic and economic restraints exist.

ACKNOWLEDGEMENTS

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MEMORANDUM

TO: Readers of Short Contributions
Bulletin 93

FROM: William H. McLemore
State Geologist

SUBJECT: Bulletin 93 reprint

Please note the loose reprint inserted into your copy of this publication. The reprint of "The Geology and Ground Water of the Gulf Trough" by Carol Gelbaum corrects a printing error made in her paper.

William H. McLemore

THE GEOLOGY AND GROUND WATER OF THE GULF TROUGH

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ABSTRACT

The Gulf Trough is a subsurface geologic feature that affected the deposition of sediments from probably as early as upper Eocene, Jackson time through Miocene time. The trough can be found parallel to a hydrologic anomaly trending N53°E. from northern Thomas, Colquitt and southern Tift Counties and is hypothesized to extend into northern Effingham County. The Gulf Trough is distinguished from an earlier defined feature, the Suwannee Strait, by location, stratigraphic position, lithologic character and structure.

Ground-water availability is adversely affected by the presence of the Gulf Trough. Yields of 50 gallons per minute are typical, whereas in areas outside the trough's influence, 500-1000 gallons per minute are common. Lithologic parameters of the aquifer affect the overall porosity and it is postulated that they are responsible for the reduced yields. Samples taken from within the axial zone of the trough show a finer microcrystalline texture with fewer bioclasts than samples from the aquifer outside the trough. The fine texture can inhibit ground-water flow. Geophysical well logs provide further evidence of low porosity. Other possibilities such as reduced aquifer thickness and multi-aquifer wells may be considered partly responsible for the reduced yields. Faulting has been considered by other investigators as the cause for the reduced yields.

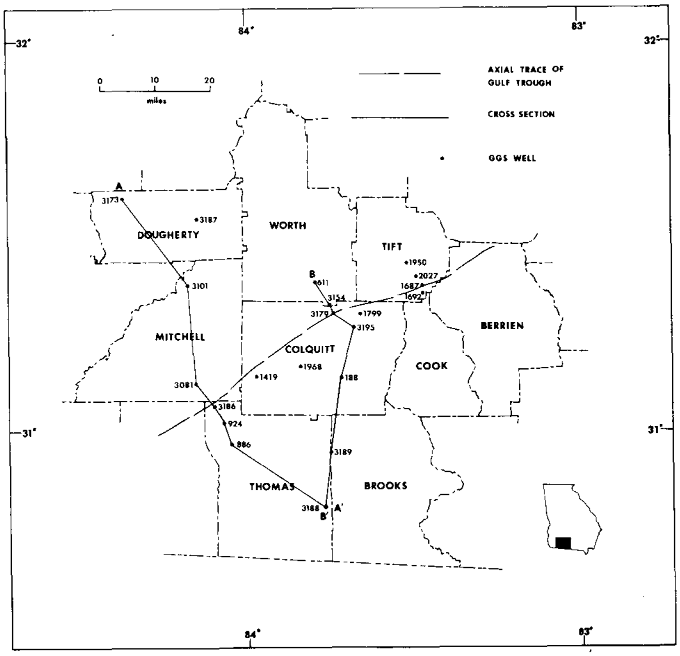

FIGURE 1 Location map of GGS wells used for this report. Location of cross sections A-A′ and B-B′. Location of axial trace of the Gulf Trough.

INTRODUCTION

The subsurface geologic feature known as the Gulf Trough is a relatively recent discovery. The name "Gulf Trough of Georgia" was first used by Herrick and Vorhis in 1963. (Hendry and Sproul, 1966, dropped the "of Georgia" and used "Gulf Trough" so that the feature wouldn't be restrictive to Georgia). The Gulf Trough is believed to be an extension of the Apalachicola Embayment located in the southwestern portion of Georgia and in the panhandle of Florida. The Apalachicola Embayment is recognized as an area of Late Mesozoic (Jurassic and Cretaceous) and Cenozoic sediment thickening. The greatest thickening occurs in the southwestern portion of the embayment; here the Tertiary is about 5,000 feet (1525 m) thick. The sediments thin in a northeasterly direction as the embayment narrows (Murray 1961). The trough extends in a northeasterly direction from the narrow end of the embayment into the Coastal Plain of Georgia (Fig. 1) and is postulated as being an area of concomitant Tertiary (possibly upper Eocene, Oligocene, Miocene and Pliocene) sediment thickening.

A geologic feature called the Suwannee Strait[1] was recognized by Coastal Plain investigators long before the Gulf Trough was first described. The Suwannee Strait was first described in Dall and Harris (1892) as a water passage extending from Savannah, Georgia to Tallahassee, Florida.

Since that time many investigators have delineated the subareal extent and geologic time span, speculated upon the cause and function and have given different names to the feature first called the Suwannee Strait. Some investigators have tried to show that the Suwannee Strait of Dall and others, and the Gulf Trough of Herrick and Vorhis and of this report are the same feature. After considering some of the data and descriptions of previous investigators (summarized below) as well as additional data acquired from the present study, the author believes that they are two independent subsurface geologic features with different geometries and geologic time spans. It is also believed that the geologic origin and function of these two features may indeed have been very different. The effect that the Gulf Trough exerts upon the present ground-water conditions of the principal artesian aquifer will be shown to be unique, while the Suwannee Strait has no apparent effect upon the ground-water conditions. The

[1]Original 1892 spelling was with one "n". Subsequent authors used two "n's" which is considered the correct spelling.

cause and effect relationship of the subsurface geology and the resulting ground-water behavior are responsible for the interest shown in this study of the Gulf Trough.

HISTORICAL BACKGROUND

The pertinent literature regarding the description of the Gulf Trough and its distinction from the Suwannee Strait is discussed below. The confusion of these features in recent literature has resulted from misinterpretation of the early literature and from postulates of origin and function based on insufficient data.

Dall and Harris (1892, pg. 121) proposed the name Suwannee Strait for a "passage between Florida and the mainland..." that "in Miocene time was a moderately deep body of water, the general trend of which did not differ much from that of a line drawn from Savannah to Tallahassee and which had a probable width of more than 50 miles." They state that this area is now occupied by the Okefenokee and Suwannee Swamps, and the trough of the Suwannee River, and is composed of deposits of clays and silicious material of Miocene age.

Applin and Applin (1964, pg. 1727) discussed a "major structural feature" that became evident from their stratigraphic study, but failed to completely identify the feature. They noted that a "channel or trough extended southwestward across Georgia through the Tallahassee area of Florida to the Gulf of Mexico". They point out that on the southeast side of the feature, from the top of the Oligocene down through the base of the late middle Eocene, the carbonate rocks form a continuous limestone sequence throughout southern Georgia into peninsular Florida. On the northwest side of the feature, from the middle Eocene down through the base of the Upper Cretaceous clastics predominate in southern Georgia and the Florida panhandle. The carbonates are laterally gradational within the "channel or trough" with the clastics, and are thinner in the gradational zone than they are on either side. They go on to state that with the passage of time the limestones of the peninsula overlapped western Florida and southern Georgia. They imply that the "structural feature" behaved as a natural barrier to carbonate sedimentation north of the channel beginning in the Upper Cretaceous and ceased to function after early middle Eocene time.

Jordan (1954) had a slightly different approach to this feature. She suggested that the Suwannee Strait is an erosional feature resulting from regional movement at the end of Cretaceous time, causing a channel to be cut between clastic rocks in Georgia and carbonate rocks in Florida. She stated that the location of the Suwannee Strait is between the Peninsular and the Chattahoochee "uplifts", but specifically drew the feature on a map extending from Tallahassee, Florida, through Lowndes County and the Okefenokee Swamp, Georgia.

Herrick and Vorhis (1963) proposed the term "Gulf Trough of Georgia" for a northeastern trending belt of anomolously thick sediments of Miocene age in northern Thomas, Colquitt and southeastern Tift Counties. These same investigators noted, in an unpublished report (1973), that the area of greatest thickening occurred along the same trend as an anomalous gradient phenomenon recorded on the potentiometric map of the principal artesian aquifer (Fig. 2). On the basis of this corroborating information the investigators extended the possible subareal extent of the Gulf Trough through Bulloch and Screven Counties. They lacked deep subsurface well data to further substantiate the coincidental trend.

Rainwater (1956) suggested that the Suwannee Strait existed along a line trending eastward from Jackson County, Florida, into Georgia. He placed the Suwannee Strait in the same geographic area as the Apalachicola Embayment and the Gulf Trough (as later defined). From his description of the strait it can be concluded that he borrowed the origin and function of the feature from the explanations of Applin and Applin (1944) and Jordan (1954), but shifted its position westward approximately forty miles (64 km). He used no geologic data to support this move. It seems evident that the confusion concerning the Gulf Trough and Suwannee Strait began here.

Chen (1965) further perpetuated the confusion when he showed a series of paleogeographic maps of southern Georgia and Florida and tried to demonstrate that the Suwannee Strait existed from at least Midwayan time through Jacksonian time, (early Paleocene through late Eocene), and migrated, during this time period to the northwest. In other words, it started out in the area delineated by Applin and Applin, and Jordan and it ended up where the Gulf Trough is now believed to exist. Chen supported the idea that the strait continued to act as a barrier and waterway separating carbonate and clastic depositional environments through late Eocene time. The barrier was previously postulated by Applin and Applin and Jordan only for sediments through middle Eocene time. Unfortunately, Chen used few data points to support his theory.

To emphasize the premise that the Suwannee Strait and the Gulf Trough are two separate features, the major conceptual differences noted prior to the present study are summarized below.

The Gulf Trough as described by Herrick and Vorhis (1963) existed in sediments of Miocene time. They noted a marked thickening of Miocene strata. They also implied an apparent thickening of strata deposited after middle Eocene time. The existence of the trough is further supported by the potentiometric map of the principal artesian aquifer, especially along the line of Miocene sediment thickening.

The Suwannee Strait was recognized and defined by Applin and Applin (1944) and Jordan (1954) as being active from Late Cretaceous through middle Eocene time. It was presumed to have acted as a barrier to sedimentation, separating carbonates on the south side of the feature from terrigenous clastics on the north side. The well cuttings they used for their studies tend to support their hypothesis of a facies change across the zone where the axis of the Suwannee Strait is located. Unfortunately, the subsurface data relied upon by these investigators was inadequate for a thorough determination of the geometry, structure, and origin of the strait, therefore much liberty was taken by these early investigators regarding the nature of the feature.

The Suwannee Strait as defined by Dall and Harris (1892), Applin and Applin (1944) and Jordan (1954) was located

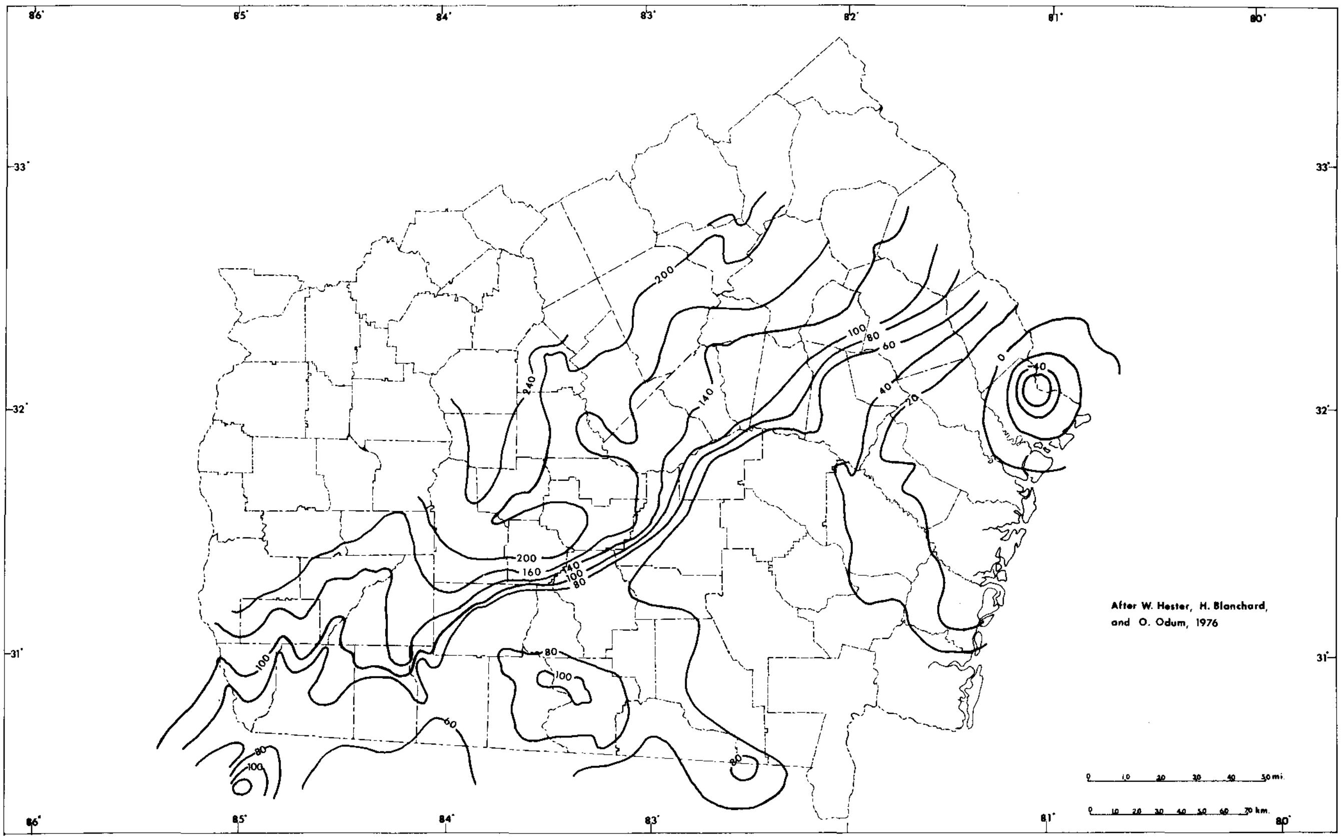

FIGURE 2 **Potentiometric map of the principal artesian aquifer, January-May 1976; simplified after Hester, Blanchard, and Odum, U.S.G.S.**

approximately forty miles (64 km) southeast of the position of the Gulf Trough as defined by Herrick and Vorhis (1963) and by this author. In addition, the Suwannee Strait is not known to be associated with any phenomenon affecting the hydraulic characteristics of the principal artesian aquifer.

ANALYTICAL PROCEDURES

A total of 21 wells were selected for this study. Their locations are shown on Figure 1 and they are listed on Table 1. Among these wells 15 are water wells, five GGS wells are cores, four with geophysical well logs. GGS #3154 is an oil test well with geophysical well logs. Almost all of these wells have complete sets of core or cuttings.

Samples were obtained from the sample library of the Georgia Geologic and Water Resources Division of the Department of Natural Resources. They were examined with a binocular microscope and in hand specimen where applicable. Samples were studied from the surface downward. In well cuttings and in core, formation boundaries were most often picked by the first appearance of diagnostic foraminifera or other fossil. Geophysical logs were used to help determine boundaries and relative porosities. Lithology was not generally considered a good criterion for determining the formation boundary between the Suwannee and Ocala Limestones.

The Suwannee Limestone is a bioclastic limestone with occasional dolomite layers. It is free of quartz sand, major clay layers and phosphate pellets. Its relative monomineralogy lithologically distinguishes the unit from the overlying Miocene strata which consists of major sand, clay and common dolomite but sparse limestone layers with occasional fossiliferous lenses and phosphate rich horizons. The Suwannee's textural and mineralogical similarity to the Ocala make it necessary to separate the two limestones by the first appearance of diagnostic foraminifera, however, in some core, lithologic differences were prominent and were used as well. *Paraotalia mexicana*, many species of *Lepidocyclina*, and/or *Dictyoconus sp.* distinguish the Suwannee Limestone from other formations. *Lepidocyclina ocalana*, *Nummulities (Operculina) marianensis*, *N. (Operculina) floridensis* or *N. (Operculina) ocalana*, *Pseudofragmina sp.* and *Asterocyclina sp.* are representative Ocala Limestone foraminifera.

The middle Eocene Lisbon Formation can be distinguished from the overlying Ocala Limestone by lithology as well as diagnostic fossils. The Lisbon Formation is commonly a fine grained, crystalline limestone that is arenaceous, argillaceous, glauconitic and micaceous.

The upper Paleocene formations, the lower Paleocene Clayton Limestone and the Cretaceous age formations were encountered in four wells. The units are disconformable and are distinguished by their lithologies.

Cross sections A-A′ (Fig. 3) and B-B′ (Fig. 4) are located on Figure 1 and were constructed to show the subsurface in a simplified manner. The fence diagram (Fig. 5) shows the relative thickness and subareal relationships of the strata in

TABLE 1 List of GGS wells selected for this study. Unless otherwise stated, well samples are cuttings.

Brooks County

GGS 3189* (core)
Location: 30° 56′ 26″ N
83° 44′ 06″ W
Altitude: 200′ (61m)
T. D. (Total Depth): 84-335′ (25.6 - 102.2m)
Depth to Suwannee Lm.: 148′ (45.1m)
Depth to Ocala Lm.: 292′ (89m)

Colquitt County

GGS 188
Location: 31° 08′ 15″ N
83° 42′ 30″ W
Altitude: 288′ (87.8m) (above sea level)
T. D.: 760′ (231.8m) (below land surface)
Depth to Suwannee Lm.: 245′ (74.7m)
(below land surface)
Depth to Ocala Lm.: 545′ (166.2m)
(below land surface)

GGS 1419
Location: 31° 08′ 15″ N
83° 57′ 30″ W
Altitude: 307′ (93.6m)
T. D.: 820′ (in Miocene) (250m)

GGS 1799
Location: 31° 18′ 00″ N
83° 38′ 45″ W
Altitude: 285′ (86.9m)
T. D.: 660′ (in Miocene) (201.3m)

GGS 1968
Location: 31° 09′ 33″ N
83° 49′ 55″ W
Altitude: 320′ (97.6m)
T. D.: 800′ (244m)
Depth to Suwannee Lm.: 480′ (146.4m)
Depth to Ocala Lm.: 670′ (204.4m)
(estimated)

GGS 3195*
Location: 31° 15′ 13″ N
83° 40′ 22″ W
Altitude: 330′ (100.6m)
T. D.: 1200′ (366m)
Depth to Suwannee Lm.: 470′ (143.4m)
Depth to Ocala Lm.: 450′ (137.3m)
Depth to Lisbon Fm.: 1080′ (329.4m)

GGS 3179 (core)
Location: 31° 17′ 33″ N
83° 43′ 24″ W
Altitude: 370′ (112.8m)
T. D.: 705′ (in Miocene) (215m)

Dougherty County

GGS 3173* (core)
Location: 31° 35′ 29″ N
84° 20′ 24″ W
Altitude: 210′ (64m)
T. D.: 675′ (205.8m)
Depth to Lisbon Fm.: 94′ (28.6m)
Depth to upper Paleocene: 351′ (107m)
Depth to lower Paleocene: 500′ (152.5m)
Depth to Cretaceous: 661′ (201.6m)

GGS 3187 (core)
Location: 30°31′ 05″ N
84° 06′ 44″ W
Altitude: 195′ (59.5m)
T. D.: 1515′ (462m)
Depth of well: 79.3-1401.3 (24.2-427.9m) at
79.3′ (24.1m) in Ocala Lm.
Depth to Lisbon: 243.8′ (74.4m)
Depth to upper Paleocene: 417.1 (127.2)
Depth to lower Paleocene: 709.8′ (216.5m)
Depth to Cretaceous: 928′ (283m)

Mitchell County

GGS 3101
Location: 31° 22′ 40″ N
84° 09′ 52″ W
Altitude: 176′ (53.7m)
T. D.: 973′ (296.7m)
Depth to Ocala Lm.: 50′ (15.25m)
Depth to Lisbon Fm.: 308′ (93.9m)
Depth to upper Paleocene: 664′ (202.5m)
Depth to Cretqceous: 912′ (278.2m)

GGS 3081
Location: 31° 07′ 11″ N
84° 08′ 39″ W
Altitude: 348′ (106.2m)
T. D.: 822′ (250.7m). No samples 274-422′
(83.6-128.7m) in Ocala Lm. at
422 (128.7m)
Depth to Lisbon Fm.: 722′ (220.2m)

Thomas County

GGS 886
Location: 30° 58′ 00″ N
84° 02′ 35″ W
Altitude: 255′ (77.8m)
T. D.: 530′ (161.7m)
Depth to Suwannee Lm.: 395′ (120.5m)

GGS 924
Location: 31° 01′ 25″ N
84° 03′ 40″ W
Altitude: 205′ (62.5m)
T. D.: 530′ (161.7m)
Depth to Suwannee Lm.: 500′ (152.5m)

Thomas County (Cont'd)

GGS 3186*
Location: 31° 03′ 53″ N
84° 05′ 12″ W
Altitude: 330′ (100.6m)
Depth: 810′ (247m)
Depth to Suwannee Lm.: 470′ (143.4m)
Depth to Ocala Lm.: 780′ (238m) (estimate)

GGS 3188* (core)
Location: 30° 48′ 39″ N
83° 45′ 23″ W
Altitude: 200′ (61m)
Depth: 70-904′ (21.4-275.7m)
Depth to Suwannee Lm.: 162′ (49.4m)
Depth to Ocala Lm.: 289′ (88.1m)

Tift County

GGS 1692
Location: 31° 20′ 55″
83° 27′ 17″ W
Altitude: 329′ (100.3m)
T. D.: 900′ (274.5m) (in Miocene)

GGS 1687
Location: 31° 22′ 10″ N
83° 27′ 15″ W
Altitude: 324′ (98.8m)
T. D.: 700′ (213.5m)
Depth to Suwannee Lm.: 640′ (195.2m)

GGS 2027
Location: 31° 23′ 40″ N
83° 27′ 50″ W
Altitude: 328′ (100m)
T. D.: 605′ (184.5m)
Depth to Suwannee Lm.: 575′ (175.4m)

GGS 1950
Location: 31° 25′ 10″ N
83° 30′ 00″ W
Altitude: 335′ (102.2m)
T. D.: 500′ (152.5m)
Depth to Suwannee Lm.: 390′ (118.95m)

Worth County

GGS 611
Location: 31° 23′ 40″ N
83° 49′ 00″ W
Altitude: 330′ (100.6m)
Depth: 55-802′ (16.8-244.6m)
Depth to Suwannee Lm.: 252′ (76.86m)
Depth to Ocala Lm.: 638′ (194.6m)

GGS 3154*
Location: 31° 19′ 03″ N
83° 44′ 15″ W
Altitude: 325′ (99.1m)
T. D.: 5568′ (1698m). Depth completed 1780′ (542.9m) - Cretaceous
Depth to Suwannee Lm. surface: 480′ (146.4m)
Depth to Ocala Lm.: 690′ (210.45m)
Depth to Lisbon: 1060′ (323.3m)
Depth to upper Paleocene: 1190′ (362.95m)

*Geophysical logs are available for these wells.

the study area. Solid unconformity lines are drawn where the the contacts were picked by diagnostic foraminifera, or by a prominant lithology change. Dashed lines are drawn where a contact is suspected because of the presence of poorly preserved fauna.

DESCRIPTION OF THE GULF TROUGH

In the present study the portion of the Gulf Trough that will be emphasized is that segment restricted to northern Thomas, Colquitt and southeastern Tift Counties. This portion of the Gulf Trough might be representative of the trough farther to the northeast, further study is in progress to extend the subsurface control in that direction.

Features of the Gulf Trough relating to ground water.

The existence and location of the trough is based partly on the potentiometric map pattern of the principal artesian aquifer (Fig. 2). The proposed extension of the trough into the northeastern section of the State is based solely on this map. In the central section of the Coastal Plain there is a series of closely spaced potentiometric contour lines representing an area of anomalous hydrologic characteristics in which there is a marked increase in gradient in the direction of flow. For example, north of the trough, in the segment from Tifton to Omega the gradient is 13 feet per mile (2.5 m per km) and measured from Omega to a point on a contour 2.5 miles (4 km) south is 28 feet per mile (5.3 m per km). In this zone of increased gradient, water wells have low yields compared to wells drilled outside of the trough. The Gulf Trough is hypothesized to have existed, and to continue to influence the geohydrology of the area located within this zone of closely spaced contours.

There are two characteristics of the trough evident on the map. First, the trough is a linear feature which appears to extend across the State. Although the trend of the axis of the trough has not been completely defined by geological methods, the anomalous potentiometric surface has a strike of N 53°E. The anomaly varies less than 10 to 12 miles (16 to 19.2 km) from a straight line extending from the southwestern corner of Grady County to the northern Effingham County line.

The second characteristic is that the feature is quite narrow. The trough widens where it joins the Apalachicola Embayment but in the central portion of the Georgia Coastal Plain it is only 8 to 14 miles (12.8 to 22.4 km) wide.

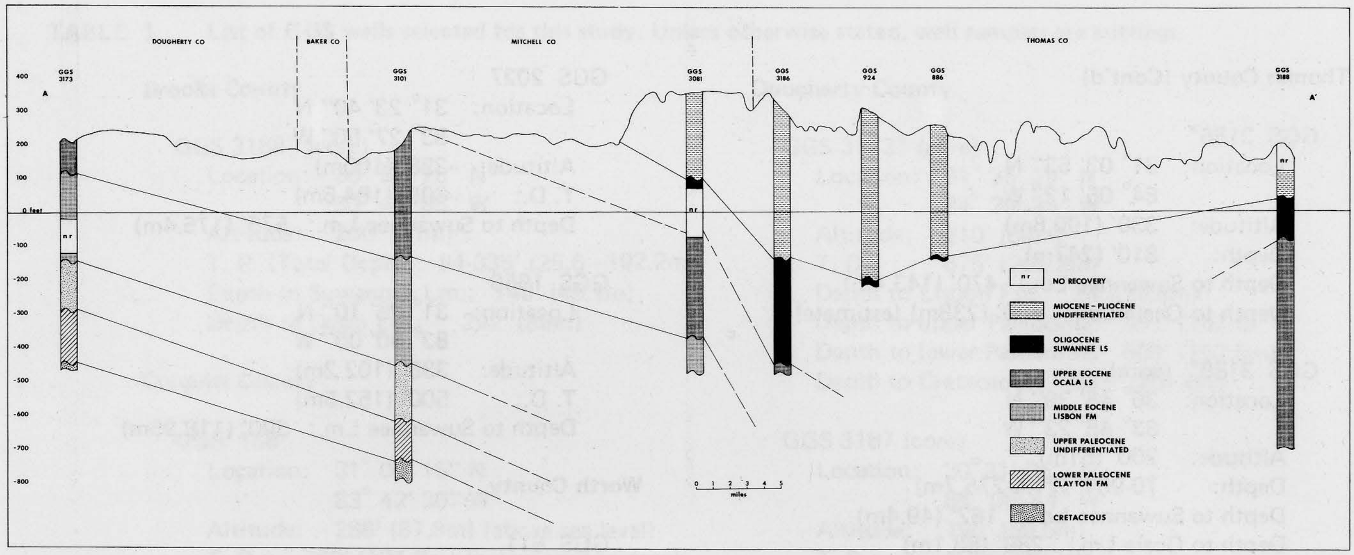

FIGURE 3 Cross section A-A′ across trend of the Gulf Trough, through Dougherty, Mitchell and Thomas Counties.

Features of the Gulf Trough relating to geology.

The zone of axis of the Gulf Trough is considered to lie within the hydraulic anomaly described above, with the axial trace approximating the mid-line of the anomaly. In the following paragraphs the subareal extent of the trough will be discussed using the data collected for this study.

An oil test well, GGS #3154, drilled near the Worth County - Colquitt County border penetrated 460 feet (140.3 m) of Miocene sediments before the Suwannee Limestone was reached. The Georgia Geologic and Water Resources Division drilled a core hole, GGS #3179, one mile (1.6 km) southeast of the oil test well to a depth of 705 feet (215 m) below land surface without reaching the Oligocene surface. Approximately five miles (8 km) to the southeast of the oil test well, or two miles (3.2 km) southeast of the city of Norman Park, a city water supply test well, GGS #3195, penetrated the Suwannee Limestone at a depth of 470 feet (143.4 m). The Ocala Limestone and Lisbon Formation were probably penetrated, but the fossil fauna was sparse and poorly preserved so that the boundaries are speculative. The three wells GGS #3154, 3179 and 3195 can be seen on the cross section B-B′ (Fig. 4) and on the fence diagram (Fig. 5).

If it is assumed that the geologic data from these wells represent the typically disconformable Miocene - Oligocene boundary relationship and have complete formations, that GGS #3154 lies on the northwest flank of the trough, GGS #3179 was drilled in the zone of axis, and that GGS #3195 lies on the southeast flank, it can be concluded that within a distance of approximately 7 miles (11.2 km) the minimum subsurface dip on top of the Oligocene changes from approximately 90 feet per mile (17.2 m per km) to the southeast to approximately 40 feet per mile (7.6 m per km) to the northwest. This subsurface structure may indicate a local depocenter where Oligocene deposits are complete but thin due to a slow deposition rate, or are complete but thick and found deeper in the sediment basin. However, GGS #3179 does not reach the Oligocene surface, therefore the thickness of the Miocene and Oligocene is still unknown for this area of the trough. The possibility exists that the Miocene sediments filled in a locally eroded portion of the Oligocene, producing a local angular unconformity. If the attitudes were calculated on the complete Oligocene unit the dips would be less steep.

Several wells that lie in the trough have thick Miocene deposits. In southeastern Tift County GGS #1962 was drilled to a depth of 900 feet (274.5 m) and does not penetrate the Oligocene surface. One mile (1.6 km) to the northwest of this well, GGS #1687 was drilled to a depth of 700 feet (213.5 m) with the bottom 20 feet (6.1 m) lying within the Oligocene (Fig. 5). In the northern part of Colquitt and southern Tift Counties it is evident that the Miocene sediments are found to be in excess of 900 feet (274.5 m). Indeed, local water well drillers have reported encountering Miocene-type sediments in wells drilled to more than 1000 feet (305 m) deep in the Omega area. GGS well #1419 can be found along the strike of the trough axis in the southwest portion of Colquitt County. In this well the cuttings show that 820 feet (250.1 m) of sediments were drilled and the Oligocene surface was reached at 640 feet (195.2 m).

In the northwest corner of Thomas County, the Miocene units are thinner than they are to the northeast along the trough axis. The Meigs city test well, GGS #3186, appears to be situated close to the trough axis (Fig. 3). In this well the Miocene is 470 feet (143.4 m) thick. GGS wells #3081 and #3101, drilled on the northwestern flank of the trough, show that the Miocene unit thins and disappears, exposing the Suwannee Limestone along the Pelham Escarpment near the town of Baconton, in Mitchell County. On the southeastern flank of the trough the Miocene thins to approximately 150 feet (45.75 m) thick in some areas of central Thomas County. For example, in GGS well #3188, the Miocene is 160 feet (48.8 m) thick.

The dip calculated on the Oligocene surface from GGS #3081 to #3186 is approximately 45 feet per mile (8.57 m per km) to the southeast. The dip calculated from GGS # 886 to #924 is approximately 15 feet per mile (2.88 m per km) to the northwest. The regional dip of the Oligocene surface for the study area outside of the trough is approximately 10 feet per mile (1.9 m per km) to the southeast.

In GGS #3186, at Meigs, the thickness of the Suwannee Limestone is 320 feet (97.6 m) and is assumed to be a complete unit in a disconformable relationship with the overlying Miocene sediments. The Oligocene is somewhat thicker near the trough axis than it is in GGS #3081, northwest of the trough, near the town of Pelham, where the Suwannee is approximately 200 feet (61 m) thick. In the northern Colquitt County area the Oligocene is approximately 200 feet (61 m) thick at the site of GGS #3154, it is 350 feet (106.75 m) at #3195, near Norman Park and 300 feet (91.5 m) at the site of #188 southeast of Moultrie. The Oligocene sediments are thickest along the trough axis and they thin to approximately 130 feet (39.65 m) in southern Thomas County towards the site of GGS well # 3188. If the excessive thickness of the Miocene sediments in the northern Colquitt County area is considered to occur locally then the two areas (northern Colquitt and northwestern Thomas Counties) reflect similar information regarding attitudes and thickness of the Oligocene.

In the above discussion of the subareal extent of the trough, a mental picture of the change in thickness of the various units becomes clear. Along the strike of the trough axis, from southeastern Tift to southwestern Colquitt Counties, the Miocene is thickest. It gets progressively thinner to the southwest. The Oligocene is generally thicker in the trough. Both of these units thin out on the adjacent sides of the trough.

Generally, the Ocala Limestone thickens just southeastward of the area of thickening of the Miocene and Oligocene units. However, some of the top contacts used for the Ocala Limestone in the zone of the trough are estimated due to insufficient faunal data, and most wells never penetrate the underlying formations, therefore, the actual thicknesses still remain to be accurately defined.

INFLUENCE OF THE GULF TROUGH UPON GROUND WATER

The Gulf Trough has a profound effect upon the groundwater availability in south Georgia. It has been previously pointed out that there is a steep increase in gradient of the potentiometric surface in the direction of flow across the axis of the Gulf Trough. This steep gradient indicates that the transmissivity of the principal artesian aquifer has been inhibited.

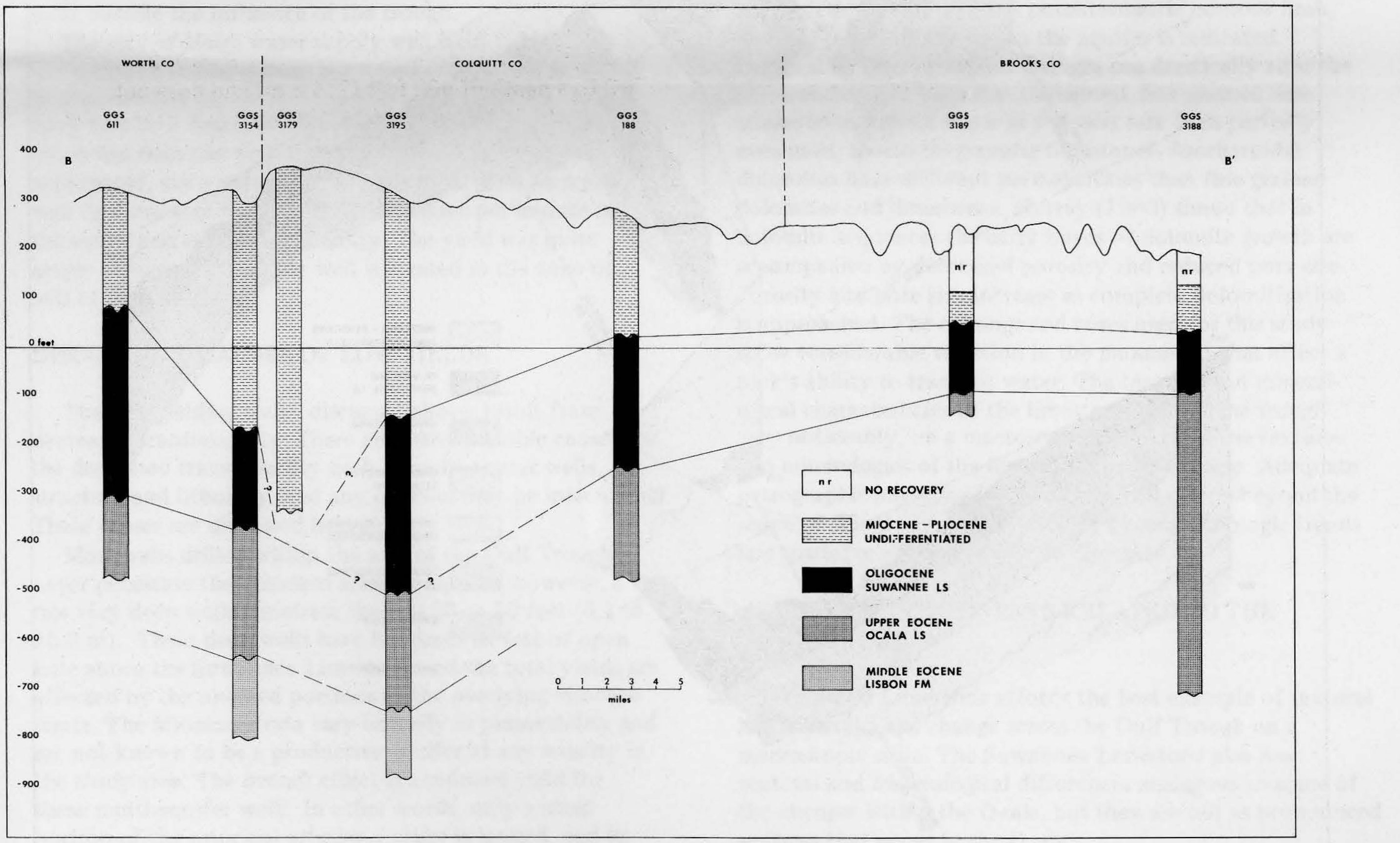

FIGURE 4 Cross section B-B′ across trend of the Gulf Trough, through Worth, Colquitt, Brooks and Thomas Counties.

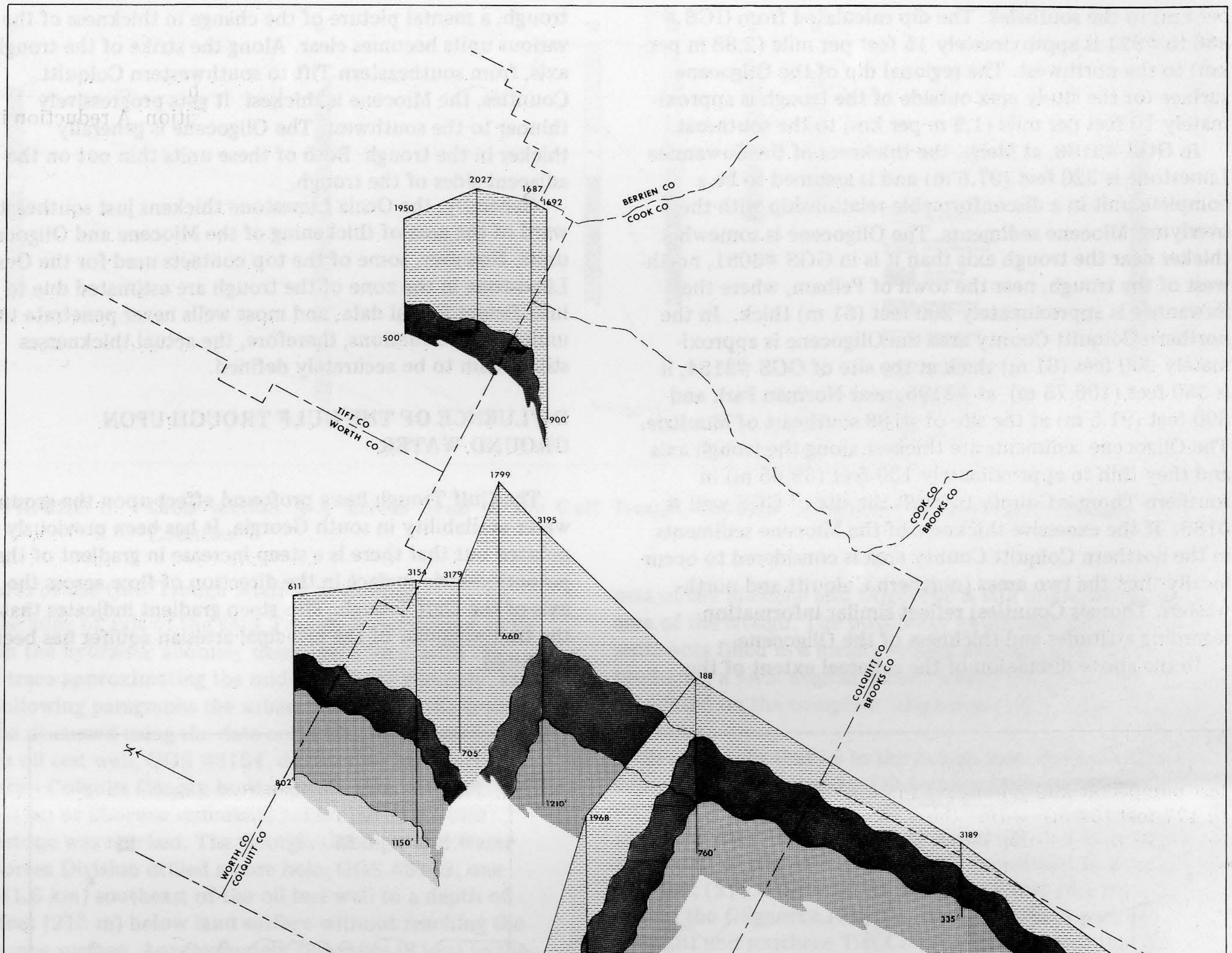

FIGURE 5 Fence diagram of the southwestern portion of the Gulf Trough.

The unique ground-water conditions within this narrow zone have a negative influence on the potential for local industrial development and community growth. Properly constructed wells in the major portion of Tift, and most of Colquitt, and Thomas Counties, adjacent to the trough, can yield as much as 1000 gallons per minute, whereas within the zone of axis of the trough, wells constructed in excess of 1000 feet (305 m) deep may yield water at a rate of only 50 gallons per minute. Some of these wells are constructed with as much as 500 feet (152.5 m) of open hole below the casing and penetrate only the upper portion of the principal artesian aquifer. The cause of the low yield is not related to the method of well construction but probably to the lithologic character of the subsurface. The expense of drilling wells in excess of 1000 feet (305 m) is prohibitive to potential community and industrial investors, especially if adequate quantities of ground water can be obtained in the areas adjacent to the trough at lesser depths.

Two city water supply test wells drilled near the axis of the trough, GGS #3186 near Meigs, and GGS #3195 near Norman Park, proved to be a surprise in the quantities of water yielded. GGS #3195 was cased to the Miocene-Oligocene contact at 470 feet (143.3 m), therefore water withdrawal is exclusively from the principal artesian aquifer. The well was pumped at the rate of 250 gallons per minute, which was considerably more than the 50 g.p.m. supplied to other wells in that area, but much less than wells outside the influence of the trough.

The city of Meigs water supply well GGS # 3186 was test pumped at 200 gallons per minute. The well was cased to 177 feet (54 m) (within the Miocene) and remained an open hole to a depth of 810 feet (247 m). The yield was much less than the well driller and consulting engineer anticipated, since they planned a principal artesian aquifer well that commonly supplies 1000 gallons per minute in the south part of Thomas County. The yield was quite ample considering that the well is located in the zone of axis of the trough.

DISCUSSION: CAUSES OF LOW YIELDS

The low yielding wells, discussed above, result from decreased transmissivity. There are many possible causes for the decreased transmissivity such as multi-aqufer wells, structure and lithology, and any of them may be interrelated. These causes are discussed below.

Most wells drilled within the axis of the Gulf Trough never penetrate the principal artesian aquifer, however, a few very deep wells penetrate the top 20 to 50 feet (6.1 to 15.3 m). These deep wells have hundreds of feet of open hole above the Suwannee Limestone and the total yields are affected by the uncased portions of the overlying Miocene strata. The Miocene strata vary laterally in permeability and are not known to be a productive aquifer at any locality in the study area. The overall effect is a reduced yield for these multi-aquifer wells. In other words, only a small portion of the principal artesian aquifer is tapped, and its percentage contribution to the combined Miocene and Oligocene aquifer may actually be large. Some of the wells from northern Colquitt and southeastern Tift Counties fall into this category.

A second possible explanation of reduced transmissivity is a reduced thickness of Oligocene strata. In the northern part of Colquitt and southeastern Tift Counties there is a possibility of a thinner aquifer which may be due to erosion, nondeposition, or slow rate of deposition. A reduction in aquifer thickness, regardless of the cause may lead to both the anomalous thickness of the overlying Miocene strata and a concomitant thinness of Oligocene limestone resulting in the development of low yielding wells in a multi-aquifer. However, the low yields that occur in this area are similar to yields from wells found southwestward along the trough axis where the Miocene is about half as thick and the Oligocene seems to be a complete unit; therefore, there is some additional factor responsible for the low yields found in the zone of axis of the Gulf Trough.

Faulting in a direction parallel to the trough axis could result in low permeability barriers and may reduce the thickness of the aquifer along the fault. This possibility has been proposed by several investigators (for example, see Sever 1966), however, their hypotheses will not be discussed here because of the lack of supporting geological data.

The last possibility is a change in lithology within the aquifer in the trough. The low yields from wells such as GGS #3195 and #3186 indicate that there is a possible lithologic control influencing the amount of water supplied to these wells from the principal artesian aquifer. Domenico (1972, pg. 168) states that when an increase in gradient occurs in the direction of flow across potentiometric contour lines, reduced permeability within the aquifer is indicated. Textural and mineralogical changes can drastically alter the permeability of an aquifer. Cemented, fine grained limestones transmit less water at a slower rate than partially cemented, bioclastic granular limestones. Saccharoidal dolomites have different permeabilities than fine grained dolomites and limestones. Murray (1960) shows that in dolomite sequences the early stages of dolomite growth are accompanied by decreased porosity and reduced pore size. Porosity and pore size increase as complete dolomitization is approached. The cuttings and cores used for this study show considerable variation in the parameters that affect a rock's ability to transmit water. The textural and mineralogical characteristics of the limestones within the trough vary noticeably, on a macroscopic scale, from the textures and mineralogies of the limestones on either side. Adequate petrographic analysis of the cuttings and core is beyond the scope of this study, therefore, only general lithologic trends and spatial relationships will be discussed.

LITHOLOGIC COMPONENTS RELATED TO THE GULF TROUGH

The Ocala Limestone affords the best example of textural and mineralogical change across the Gulf Trough on a macroscopic scale. The Suwannee Limestone also has textural and mineralogical differences analagous to some of the changes within the Ocala, but they are not as pronounced as those that occur in the Ocala.

On the northwestern side of the trough, particularly in Dougherty and Mitchell Counties, the Ocala is a grain-supported, framework limestone, with a great amount of visible porosity, both primary and secondary. The limestone has a similar lithology and texture throughout: bioclastic granular with coralline algae, bryozoan debris, foraminiferal remains and other kinds of bioclasts. Cementing varies haphazardly in amount and in zones of placement. Dolomitization is minor with dolomite occurring as euhedral rhombs, sparsely scattered throughout the section. Since the limestone is extensively recrystallized the foraminiferal control necessary to identify faunal zone boundaries within the Ocala is lacking.

In oil test well GGS #3154 from Worth County (Fig. 1) the cuttings show that the Ocala Limestone has a visibly lower porosity than the Ocala has in Dougherty and Mitchell Counties farther to the northwest. The geophysical well logs also indicate a lower porosity. The limestone is recrystallized, micritized, and less bioclastic in Worth County than in Dougherty and Mitchell Counties.

Within the trough, the final 30 feet of the Meigs city test well, GGS #3186, is in the top of the Ocala Limestone. This limestone is composed of fine-grained crystalline calcite to micrite, not visibly porous and not very bioclastic. The low porosity found in the cuttings is supported by the geophysical logs and by the low sustained yield mentioned earlier.

On the southeast side of the trough the Ocala Limestone is texturally and mineralogically more diverse than on the northwest side. The Ocala tends to have thin discrete lithologic beds within the whole. Some of these beds tend to be finer grained and less bioclastic, almost a lithographic limestone. The cuttings from the lower portions of some wells tend to be fine grained, well cemented, with sparse, small, delicate *Lepidocyclinas sp.* Dolomite beds occur in layers 50 (15.25 m) or more feet thick. The dolomite is most commonly saccharoidal, with euhedral crystals. This type of dolomitization usually obliterates the original texture and is generally very porous. These dolomite units cannot be traced laterally with existing controls. Most limestone layers are bioclastic granular, similar to the Ocala on the northwest side of the trough.

The most outstanding difference in the Ocala Limestone on the southeast side of the trough is the presence of gypsum in the lower portion of the limestone. The gypsum occurs in lenses and in intergranular pore spaces. The gypsum is found in close association with both the limestone and dolomite from cuttings in GGS #188 (Fig. 1).located near Moultrie, Colquitt County, and from core in GGS # 3188 located in Thomas County.

In a test hole drilled by the Army Corps of Engineers near Valdosta, Lowndes County, gypsum was encountered in vugs and in intergranular pore spaces in the Ocala Limestone within a zone 900-1000 feet (274.5-305 m) below land surface. The presence of gypsum with no other evaporite species present was determined by X-ray diffraction analyses of this core.

Gypsum has an adverse effect upon the ground water quality on the southeast side of the trough. Stiff diagrams drawn by Zimmerman (1977) show anomalously high sulfate concentrations in the area southeast of the trough and normal readings of sulfate north of and within the trough.

Lithologic characteristics such as texture, mineralogy and biota are ultimately determined by the geochemistry of the depositional environment. The geochemistry is affected by many physical parameters including water temperature, water depth, energy, currents, photic penetration levels and latitude. The prevailing conditions in the water mass in the area of the trough compared to conditions on either side during sediment deposition are responsible for the genetic differences found in the limestones discussed above. The marine depositional environment within the trough may have acted as a partial boundary or separation and the water may not have been free to completely mix across the entire depositional environment.

The original lithology however, is only partly responsible for the diagenetic changes. The flow rates, residence time and flow direction of ground water within the subareally exposed limestones influence diagenesis as well. Equilibrium conditions at the fresh water – salt water interface are also believed to influence the deposition of dolomite and perhaps gypsum as well (Folk and Land 1975).

The lithologic differences that have been found to occur in the aquifer rocks across the trough suggest possible conditions in the depositional environment. One lithologic factor of the limestones found within the trough is the higher micrite content with correspondingly lower amount of biota. This factor may indicate less active currents or other parameter affecting precipitation of microcrystalline calcite. However, the micrite may be diagenetic and in that case the postdepositional environment would be the causitive factor.

Another factor is the thickening of the Oligocene and especially Miocene strata formed in the trough. Assuming that the Miocene and Oligocene are disconformable in all areas of the trough in the study area, then it is possible that there may have been a faster rate of deposition within the trough due to subsidence. The thicker Miocene unit within the trough in Colquitt County may be evidence of more pronounced subsidence in this area of the trough producing local basins of deposition within the length of the trough.

The presence of gypsum on the southeast side of the trough is direct evidence that there is a difference in the postdepositional environments within and on either side of the trough. The limestones deposited within the trough are texturally finer grained and less bioclastic and porous than limestones on the southeast side, and may have acted as a barrier and restricted the postdepositional emplacement of the gypsum to the southeast side of the trough.

In the trough therefore, carbonate deposition may have taken place at a different rate and under dissimilar equilibrium conditions than the adjacent areas of carbonate deposition. These adjacent areas may have developed as a carbonate platform similar to that of the Bahama Banks today and the Gulf Trough may have acted as an area of incomplete separation of that platform with its own unique depositional environment.

SUMMARY AND CONCLUSIONS

The Gulf Trough is a subsurface geologic feature represented by a thickening of Miocene strata and possibly Oligocene and upper Eocene as well. The trough is located along a line trending in a northeast direction from northern Thomas County, Colquitt County, and southern Tift County. The Gulf Trough is distinguished from the feature called the Suwannee Strait by the age of the formations encountered, the lithologic characteristics of these formations and the location of the feature.

The presence of the Gulf Trough has an adverse effect upon the ground-water yields of the principal artesian aquifer. Low water yields may be due to:

1. The principal artesian aquifer being located much deeper in the subsurface than elsewhere so that few wells are drilled deep enough to benefit from the more permeable zones.
2. Multi-aquifer wells that tap the entire Miocene and only the top 20-50 feet (6.1-15.25 m) of the aquifer.
3. The principal artesian aquifer being thinner in the Norman Park area due to erosion, non deposition, or slow rate of deposition of Oligocene sediments.
4. Faulting in a direction parallel to the trough, which may result in a low permeability barrier, such as a reduced aquifer thickness across a fault plane.
5. A carbonate facies change across the feature altering the permeability of the aquifer limestones.

The origin of the thicker deposits within the Gulf Trough may be associated with subsidence related to the formation of the Apalachicola Embayment. The subsidence caused differences in the chemical equilibrium and physical conditions within the water mass across the trough at the time it was active. The geochemical and physical parameters influenced carbonate deposition which in turn affected the postdepositional diagenetic environment, altered the original limestone, caused dolomitization, and controlled the emplacement of gypsum. The resulting limestone types have affected the ground-water conditions that are unique to the central portion of the Coastal Plain of Georgia.

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