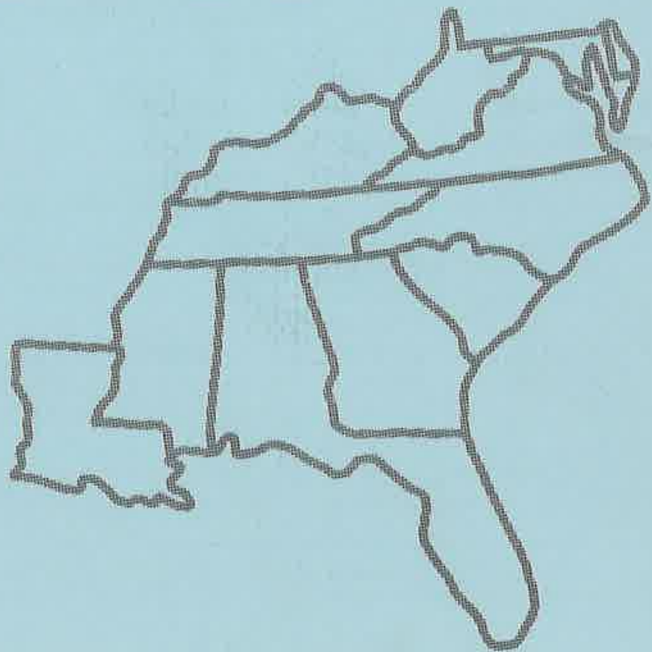
SOUTHEASTERN GEOLOGY



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SOUTHEASTERN GEOLOGY

Table of Contents

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RECENT TRENDS IN THOUGHT AND RESEARCH ON SOUTHERN APPALACHIAN TECTONICS

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ABSTRACT

The southern Appalachians may be divided into two distinct tectonic realms, the largely unmetamorphosed, thrusted and folded Valley and Ridge and Plateau and the complexly deformed and metamorphosed Blue Ridge and Piedmont. In the Plateau and Valley and Ridge the question of basement involvement still exists. Although geophysical studies and drilling favor a thin-skinned interpretation in many structures, some are best interpreted as thick-skinned.

Study of the polyphase tectonic history of the Blue Ridge and Piedmont is receiving considerable emphasis. Investigation of the complex relationships between regional metamorphism, mobilization and folding in the Blue Ridge and Piedmont has just begun. Relations between retrogressive metamorphism and complex faulting in the Brevard Zone have resulted in a multitude of interpretations. The importance and extent of the Towaliga and Goat Rock faults is just coming to light. Stratigraphic sequences in highly metamorphosed rocks are being recognized and utilized to help reconstruct the tectonic history of this region. Two schools of thought are developing regarding the structure of the Inner Piedmont. One interprets it as an autochthonous mass which was mobilized and folded into several large recumbent nappes. The other maintains that mobilization and internal folding of this mass occurred southeast of the Kings Mountain (or Pine Mountain) belt and it was then transported northwestward as a meganappe to its present position. Comparisons have been made between the crystalline Appalachians and the Alps and Greenland Caledonides.

The concept of plate tectonics has provided a possible mechanism for the generation of orogenic belts. At least one model for the southern Appalachians incorporates this concept.

INTRODUCTION

This paper is a progress report and summary and also an advertisement for anyone who might be interested in working on the problems of Appalachian tectonics. For, although we know something about the southern Appalachians and more about some portions than others, large areas of this belt remain as some of the most complex, least known geologic terrane in the world. The purpose of this paper is to present the ideas that exist and point out where research and thought on southern Appalachian tectonics is heading, but I will not attempt to discuss all the works on the tectonics of the southern Appalachians that have been published in recent years, nor to undertake to present detailed arguments for or against the many concepts that exist today. I wish to acquant those not familiar with, or working directly in, problems of Appalachian tectonics with the problems and the orientation of research in this area. Those interested in a detailed presentation should refer to John Rodgers' comprehensive work on the Tectonics of the Appalachians (Rodgers, 1970) and the more recent papers published since Rodgers' book. The southern Appalachians is here taken to include that portion of the Appalachians lying south of Roanoke, Virginia.

At one time I held the opinion that precise answers could be provided for the major questions of Appalachian tectonics and that these answers are readily obtainable if one could probe and gather enough data. I still hold this opinion but realize that we simply do not have access to all the data necessary for precise solutions to the problems. But through the means at our disposal, including detailed geologic mapping, stratigraphic and petrologic studies and more and better radiometric ages, we are rapidly accumulating urgently needed data for more precise solutions. We also have at our disposal a new concept for explaining and genetically connecting the features of mountain belts, continents and ocean basins on a global scale, that of plate tectonics.

The southern Appalachians may be divided into two major zones: the essentially unmetamorphosed folded and thrusted Valley and Ridge and Cumberland Plateau; and the multiply deformed, metamorphosed and partially mobilized Blue Ridge and Piedmont (Figure 1). Discussion will proceed from the unmetamorphosed to the more complexly deformed areas and then to some regional syntheses.

Acknowledgments

This paper was presented at the NAGT Symposium on the "Southeast as a Geological Laboratory" at the Southeastern Section Meeting of the Geological Society of America in Blacksburg, Virginia, May, 1971. The symposium was convened by H. R. Cramer of Emory University. I am grateful to Dr. Cramer for inviting me to present this review. Critical review by P. B. King and several unnamed reviewers have been quite helpful in pointing out items that need clarification and inconsistencies in presentation. For this I am also grateful.

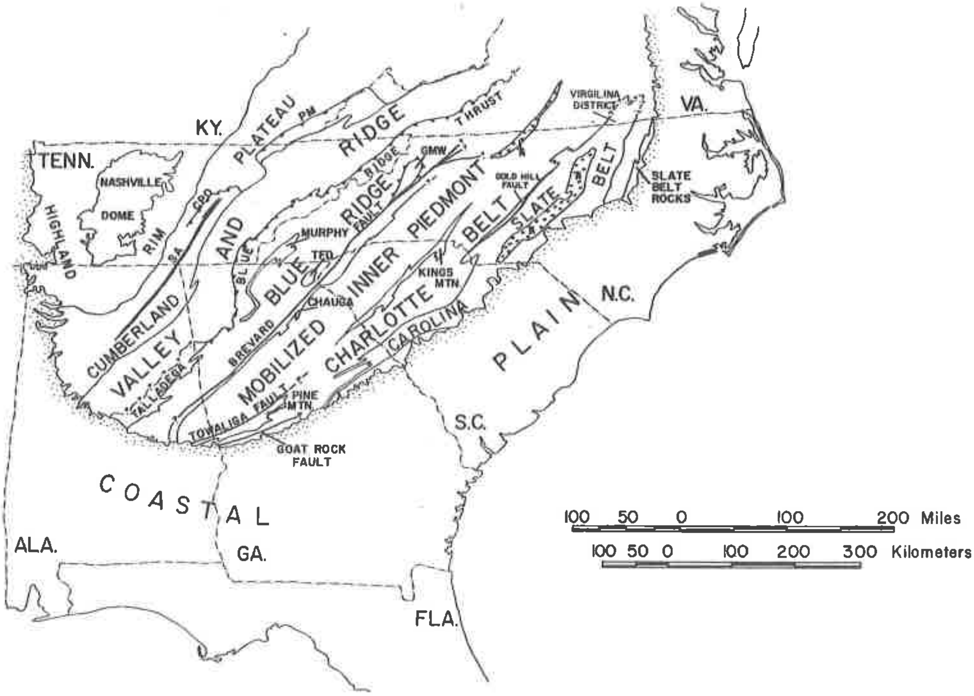

Figure 1. Map of the southeastern states illustrating the major geologic belts and several other major geologic features in the southern Appalachians. SA - Sequatchie anticline. CPO - Cumberland Plateau overthrust. PM - Pine Mountain fault. GMW - Grandfather Mountain Window. TFD - Tallulah Falls Dome. TR - Triassic basins.

VALLEY AND RIDGE PROVINCE AND CUMBERLAND PLATEAU

Central to interpretation of Valley and Ridge or Plateau structure is the question of basement involvement. The so called "thick-skinned" and "thin-skinned" theories have evolved around this question. The thick-skinned concept, maintains that all folds and faults extend into basement and that the sedimentary cover is deformed passively in response to active basement deformation (Figure 2).

The thin-skinned school maintains that the Valley and Ridge structures were developed as marginal features to the principal area of deformation and are the result of tangential stresses directed from the southeast which act only upon the sedimentary prism. These stresses produced large bedding thrusts and rootless folds without involvement of basement (Figure 3).

Considerable impetus has been given to the thin-skinned concept by the work of Gwinn (1964) and his interpretation of well data from the

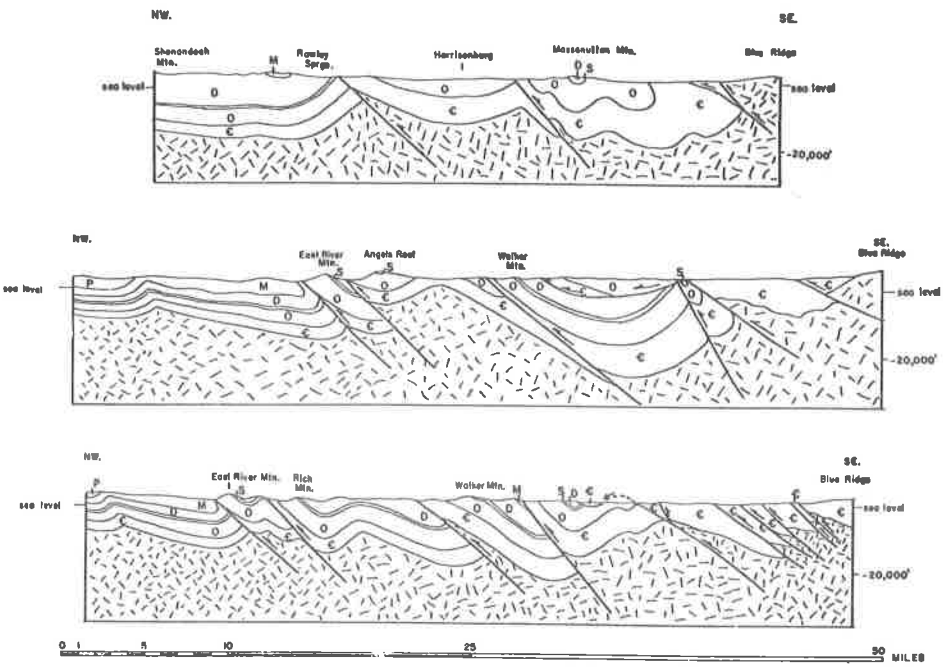

Figure 2. Geologic cross-sections across the Valley and Ridge of Virginia illustrating the thick-skinned and depositional synclines concepts (after Cooper, 1961).

Plateau and Valley and Ridge of Pennsylvania and West Virginia. This work has strongly influenced the thinking concerning thin-skinned tectonics in the southern Appalachians. Gwinn demonstrated that faults in the Plateau of the Central Appalachians rise from major sole thrusts and form branching splay thrusts in the cores of Plateau anticlines which are unfaulted at the surface. Several zones of decollement may exist along a single thrust sheet at different levels as faulting refracts from a lower level to one higher in the section along a more steeply inclined segment of the fault.

Harris (1970) concluded from a study of the window area of the Pine Mountain block and data from the Bales well that folded thrusts are not the result of folding operating within the sedimentary prism, rather that arching has resulted from subsurface duplication of beds. Closure of the arch is equal to the amount of duplication.

Harris (1970, p. 170) proposed that there is no abrupt change in structural style coinciding with physiographic boundary between the Valley and Ridge and Plateau. However, Rodgers (1970) and Milici (1970) maintain that there are structural changes associated with the Valley and Ridge-Plateau boundary. Milici (1970, p. 134) states that the

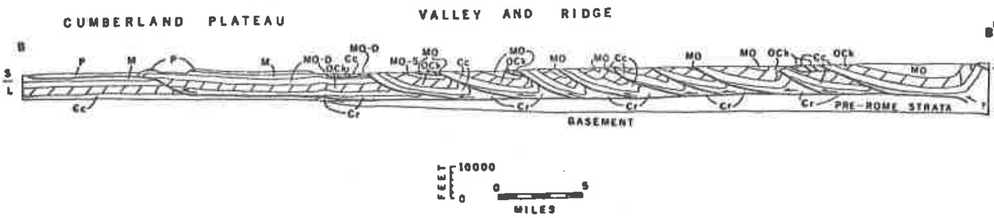

Figure 3. Geologic cross-section through the Valley and Ridge of Tennessee illustrating the thin-skinned interpretation (from Milici, 1970, Figure 3).

Allegheny front in Tennessee separates thrust faults of different types. Those in the Plateau are shear thrusts while those in the Valley and Ridge are break thrusts and override previously folded terrain. Rodgers has recognized several other structural fronts to the southeast within and on the southeast edge of the Valley and Ridge.

Cooper (1961, 1964, 1968) has pointed out that in the Valley and Ridge thicker sedimentation in depositional synclines persisted in the Appalachian region from the Early Paleozoic through the Late Paleozoic (Figure 2). Cooper (1961) maintained that certain portions of the geosyncline were subsiding faster than others. Compression during subsidence resulted in development of a thrust on the short (southeast) limb of the depositional syncline. The thrust flattens as it passes upward from the basement rocks into the overlying sediments and overrides the trough of the depositional syncline. Folding of the thrust surface results from continued subsidence of the depositional syncline.

Cooper (1961, 1964, 1968) has interpreted Valley and Ridge structure and stratigraphy as interrelated and continuous over a long period of time. Certainly there is evidence from other studies that folding preceded thrusting. Milici (1970, p. 134) has concluded that the thrusts of the Valley and Ridge rode over already folded terrane. Whether folding occurred immediately prior to thrusting or preceded it by some time is not clearly established.

The question of basement involvement remains with us. To some extent the scales seem tilted toward the thin-skinned advocates. Some geophysical studies support the thin-skinned concept (Watkins, 1964) while others (Sears and Robinson, 1971) support Cooper's concepts. Studies by Thomas and Drahovzal (1971) in the Valley and Ridge of Alabama indicate that the best interpretation of the Birmingham anticlinorium may be thick-skinned. Moreover, structure sections constructed in other areas with a thin-skinned interpretation occasionally unavoidably involve basement and the pre-Rome clastics that overlie the basement (Figure 4). Milici (1970, Figure 6) has constructed such sections across the Valley and Ridge of Virginia.

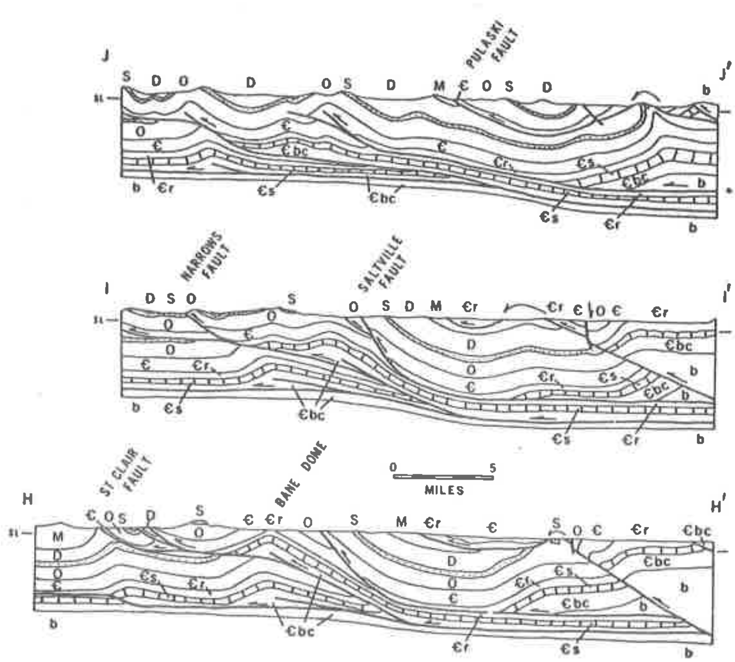

Figure 4. Thin-skinned interpretive cross-sections through a portion of the Virginia Valley and Ridge. The surface geologic relations here require that basement (b in the sections) be involved in faulting (from Milici, 1970, Figure 6).

CRYSTALLINE APPALACHIANS

In recent years there has been a significant increase in the number of geologic studies in the Blue Ridge and Piedmont of the southern Appalachians. Before examining some of the conclusions of a few of those studies, a review of the divisions of that area may be in order (Figure 1). The Blue Ridge consists of a western belt of relatively unmetamorphosed to moderately metamorphosed Late Precambrian and Cambrian sedimentary and volcanic rocks. To the southeast lies the Murphy belt, then the less well known areas of perhaps basement and Late Precambrian metasedimentary rocks equivalent to the Ocoee Series in the central and southeastern Blue Ridge. Interrupting this complexly folded and metamorphosed mass is the Grandfather Mountain window in North Carolina. A structure in northeast Georgia thought by some (Livingston and McKniff, 1967) to be similar to the Grandfather Mountain window is the Tallulah Falls Dome (Figure 1). The Blue Ridge and its southwestward extension into the physiographic Piedmont of Georgia and Alabama is bordered on the southeast by the Brevard Zone. Southeast of the Brevard Zone is the Inner Piedmont which is divisible into a high grade mobilized portion and a low to moderate grade belt, the Chauga Belt (properly the Chauga belt is a belt separate from the Inner Piedmont). Southeast of the Inner Piedmont are the Kings Mountain and Pine Mountain belts. These are flanked to the southeast by the Charlotte belt and then the Carolina slate belt.

Mapping by several groups of U. S. Geological Survey geologists in the Great Smoky Mountains, northeast Tennessee, the Grandfather Mountain window area and to the northwest has proved a number of large thrust sheets in the western Blue Ridge of North Carolina, Tennessee and southwestern Virginia (King and Ferguson, 1960; Hamilton, 1961; Hadley and Goldsmith, 1963; King, 1964; Neuman and Nelson, 1965; Rankin, 1970; Bryant and Reed, 1970a, 1970b). Mapping of several window areas in the Smokies and particularly the Grandfather Mountain window has led to the general acceptance of the allochthonous character of the Blue Ridge from southwest Virginia to Georgia. Geologists working in the Grandfather Mountain area have concluded that the rocks inside the window are also allochthonous (Bryant and Reed, 1970a, Figure 3). This conclusion is based upon positions of different thrust sheets to the west and the nature and metamorphic grade of the rocks contained within each thrust sheet. The Ocoee Series has been subdivided into several stratigraphic units by King, Hadley, Neuman and others. Several premetamorphic thrusts were recognized in the Smokies, of which the most extensive is the Greenbrier fault, which was generated during the Early Paleozoic and was later metamorphosed and broken by other faults (Hadley and Goldsmith, 1963; King, 1964).

The stratigraphy of the Late Precambrian Mount Rogers formation has been delineated by Rankin (1967, 1970) in the area north of the Grandfather Mountain window. Rankin also recognized and named the Late Precambrian Ashe Formation as an assemblage of medium to high grade metasedimentary and metavolcanic rocks presently thought to be in part equivalent to the Ocoee Series and Mount Rogers Formation. These rocks were formerly called the Carolina and the Roan Gneiss and thought to be part of the basement assemblage of the Blue Ridge (Keith, 1903, 1905, 1907a). The rocks of the Spruce Pine synclinorium probably in part belong to the Ashe Formation (Rankin, 1971, Figure 3).

The rocks of the Murphy belt form a syncline overlying the Great Smoky Group and are thought by several geologists to be Cambrian in age (Keith, 1907b; Hurst, 1955; Fairley, 1965; Forrest, 1969; Power and Forrest, 1971). However, Hadley (1970, p. 256) presents the possibility that the Murphy belt rocks may be equivalent to part of the Walden Creek Group. The synclinal structure of the Murphy belt was originally discerned by Keith (1907b) and LaForge and Phalen (1913) and has been confirmed by Hurst (1955), Power and Reade (1962), Fairley (1965), Forrest (1969) and Power and Forrest (1971).

Recent work by J. B. Hadley and Arthur Nelson in the Blue Ridge of North Carolina has shown that the Ocoee Series is more extensive than has previously been known, and the area underlain by basement rocks is considerably smaller (Hadley, 1970; Hadley and Nelson, 1971). Similar metasedimentary assemblages have been recognized by several geologists working in the same belt in Georgia (Higgins, 1966; Bentley and Neathery, 1970; Hatcher, 1971a) and South Carolina (Hatcher, 1969, 1971b).

The Tallulah Falls Dome in northeast Georgia has been interpreted as a window like the Grandfather Mountain window because of its structural position, configuration and its lithologic assemblages (Livingston and McKniff, 1967). My preliminary studies (Hatcher, 1971a) indicate that the dome may be part of a large nappe which is cored by basement rocks and flanked by Late Precambrian Ocoee equivalent metasedimentary and metavolcanic rocks. Much work remains before this problem will be solved.

Many geologists who have worked in the Blue Ridge of the southern Appalachians have noted the polyphase character of the deformation (Hamilton, 1957; Hadley and Goldsmith, 1963; King, 1964; Reed and Bryant, 1964; Higgins, 1966; Butler and Dunn, 1968; Hatcher, 1969, 1971a, 1971b). Periods of Middle Precambrian (Grenville) and Middle Paleozoic folding, intrusive activity and progressive metamorphism and Late Paleozoic folding, thrusting and retrogressive metamorphism have been recognized.

Brevard Zone

The Brevard Zone is a structure to which considerable thought has been devoted for a number of years and is presently being intensively studied by several geologists. It is a 375 mile linear belt that contains rocks that are very similar throughout its length. This similarity is partly due to the penetrative cataclastic nature of these rocks and probably also due to an original similarity of the sedimentary rocks (silty shale, black shale, impure limestone, fine sandstone and graywacke) which are involved in faulting along a portion of its extent (Hatcher, 1969; 1971b). Many interpretations of the Brevard Zone have been proposed in recent years. It has been interpreted as a strike-slip fault (Reed and Bryant, 1964; Reed and others, 1970), a root zone (Burchfiel and Livingston, 1967); a tightly appressed isoclinal syncline (Dunn and others, 1968), a thrust (Hatcher, 1969, 1971b; Roper and Dunn, 1970; Bentley and Neathery, 1970; Rankin and others, 1971) and most recently a paleosubduction zone (Watkins, 1970, 1971). Few geologists today question that it is a fault zone and it is reasonably established that a mappable, unbroken stratigraphy exists along a sizeable portion of the fault zone (Hatcher, 1970a). It is possible that all the proposed interpretations are in part correct since, according to Butler (1971), the Brevard Zone is a complex and fundamental structure which has undergone a long history of movement. It will undoubtedly be a number of years before we completely understand it.

Piedmont

The Inner Piedmont has been divided into two belts in North Carolina, South Carolina and part of Georgia: the Chauga belt, previously termed the Low Rank belt by Hatcher (1969) and non-migmatitic belt by Griffin (1969a, 1971b), of low to moderate metamorphic grade containing the Poor Mountain, Henderson and Chauga River (Brevard) rocks, and the mobilized high grade portion of the Inner Piedmont consisting of biotite, hornblende and granitic gneisses, amphibolites, schists and quartzites (Figure 1). Actually the Chauga belt is a division of the Piedmont of the same order as the Inner Piedmont and Kings Mountain belts. Griffin (1967) first proposed that the mobilized Inner Piedmont consists of a series of large nappes. He (Griffin, 1967, 1969a, 1969b, 1970, 1971a, 1971b) has compared this area to a stockwork folding model of the East Greenland Caledonides and has proposed that the mobilized Inner Piedmont comprises an infrastructure, while the Chauga belt, Brevard Zone and Kings Mountain belt make up the detachment zone in the stockwork model (Figure 5). Bentley and Neathery (1970) have found evidence in the Inner Piedmont of Alabama to support the stockwork model, although Griffin's nappes are rooted in the Inner Piedmont while those of Bentley and Neathery are rooted to the southeast (Figures 6 and 7).

The Charlotte belt is the zone of maximum intrusive activity in the southern Appalachians. Plutons ranging from ultramafic to felsic have intruded the gneiss and schist country rock. Many plutons have been individually studied and Butler and Ragland (1969) made a petrologic and chemical study of a large number of these plutons. They concluded there were three periods of intrusive activity: pre-, syn-, and postmetamorphic.

Butler (1966), in a study of the geology of York County, South Carolina, found that the country rocks into which the Charlotte belt plutons are intruded have been folded into a series of near vertical isoclinal folds. Other studies indicate that folding in the Carolina slate belt is more open but axial surfaces remain near the vertical (Stuckey, 1958, 1965; Parker, 1968; Sundelius, 1970; Tobisch and Glover, 1971). Tobisch and Glover (1971) interpret the boundary between the Charlotte and Carolina slate belts as a metamorphic gradient that has produced differences in structural style which they believe to be an infrastructure-superstructure relationship.

Several studies in recent years have revealed stratigraphic successions in different areas of the Carolina slate belt. Conley and Bain (1965) first demonstrated that the rocks of the Carolina slate belt may be divided into mappable units. Thicknesses of many thousands of meters have been estimated (Overstreet and Bell, 1965; Stromquist, 1966; Secor and Wagener, 1968; Sundelius, 1970). The units of the slate belt are predominately volcanic sandstones, mudstones, and volcanic flows (Crickmay, 1952; Overstreet and Bell, 1965; Sundelius,

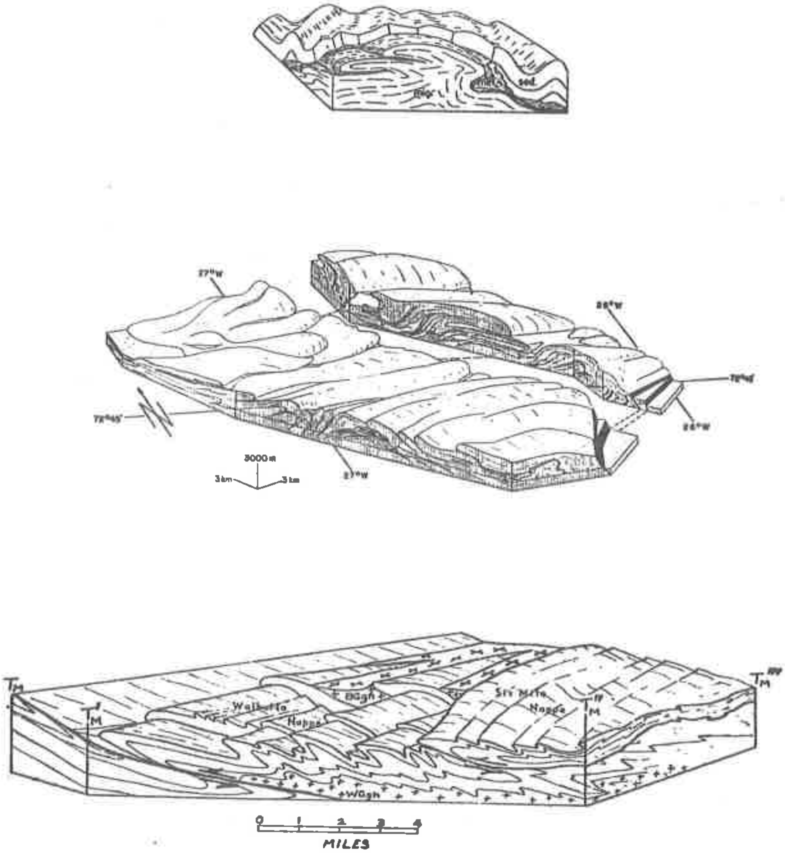

Figure 5. Stockwork tectonics. The top block diagram is a generalized model showing the relationships between the infrastructure (mig.), superstructure (sed.) and the detachment zone (met.) in East Greenland according to Haller. The central diagram illustrates the structural pattern across a portion of East Greenland while the lower diagram expresses that in a portion of the Inner Piedmont of northwestern South Carolina (after Griffin, 1969a, Figures 9 and 10, 1969b, Figure 6).

1970). All of these rocks have been slightly to moderately metamorphosed. Several geologists believe that the lithologic units of the slate belt are also recognizable in the higher grade portions of the Charlotte, Kings Mountain and Inner Piedmont belts (Overstreet and Bell, 1965; Secor and Wagener, 1968). Overstreet and Bell (1965) have recognized higher grade slate belt rocks in the Charlotte belt. Other geologists (Stuckey, 1958, 1965; McCauley, 1961), however, have interpreted the higher grade portions of the Charlotte belt as anticlinoria of older rocks

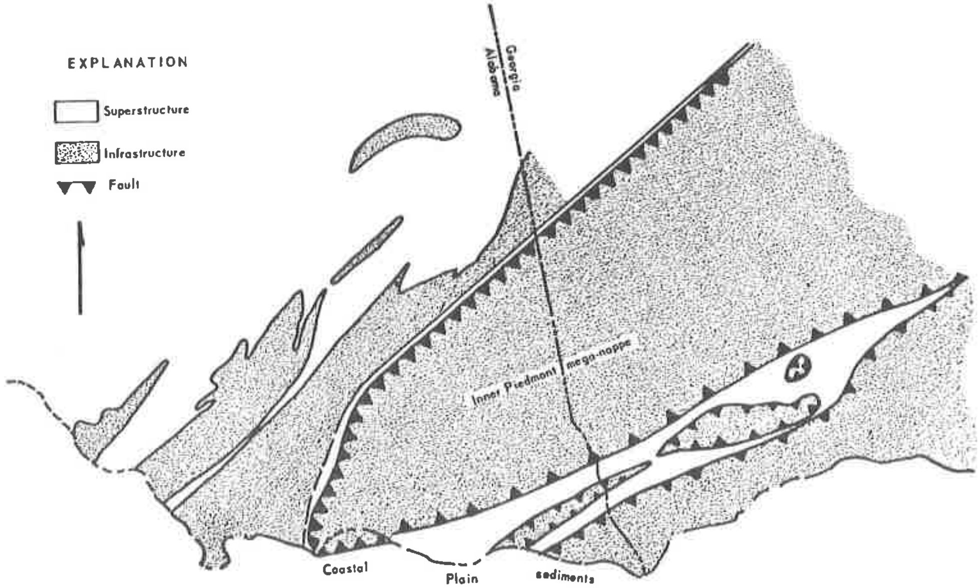

Figure 6. Map of a portion of the Piedmont of Alabama and Georgia showing the relationships between major tectonic units as interpreted by Bentley and Neathery (1970, Figure 18).

and the Kings Mountain and Carolina slate belts as flanking synclinoria. Much more work is needed before this question can be resolved.

Bentley and Neathery (1970) have revived an idea proposed originally by Clarke (1952) that the Pine Mountain belt in Georgia and Alabama is a window and that the Inner Piedmont and Charlotte belts are both allochthonous (Figures 6 and 7). They connect the Goat Rock, Towaliga and Brevard faults as a single large thrust that transported the Inner Piedmont and Charlotte belt rocks from the southeast. An alternative interpretation treats the Inner Piedmont and Charlotte belts as autochthonous units (Griffin, 1971b), except for tectonic transport during folding, and the respective faults to have developed separately but perhaps during the same but post-folding series of movement events (Hatcher, in press).

PLATE TECTONICS

Studies of the structures of modern island arcs, ocean basins and recent mountain belts in the past few years by many geologists and geophysicists have provided us with the concept that the lithosphere is in a state of continuous motion and major parts of it are constantly being generated and destroyed (see for example Isacks and others, 1968). Plate generation occurs at oceanic ridges and destruction of heavier (oceanic) crustal material occurs as it is engulfed in treach areas or

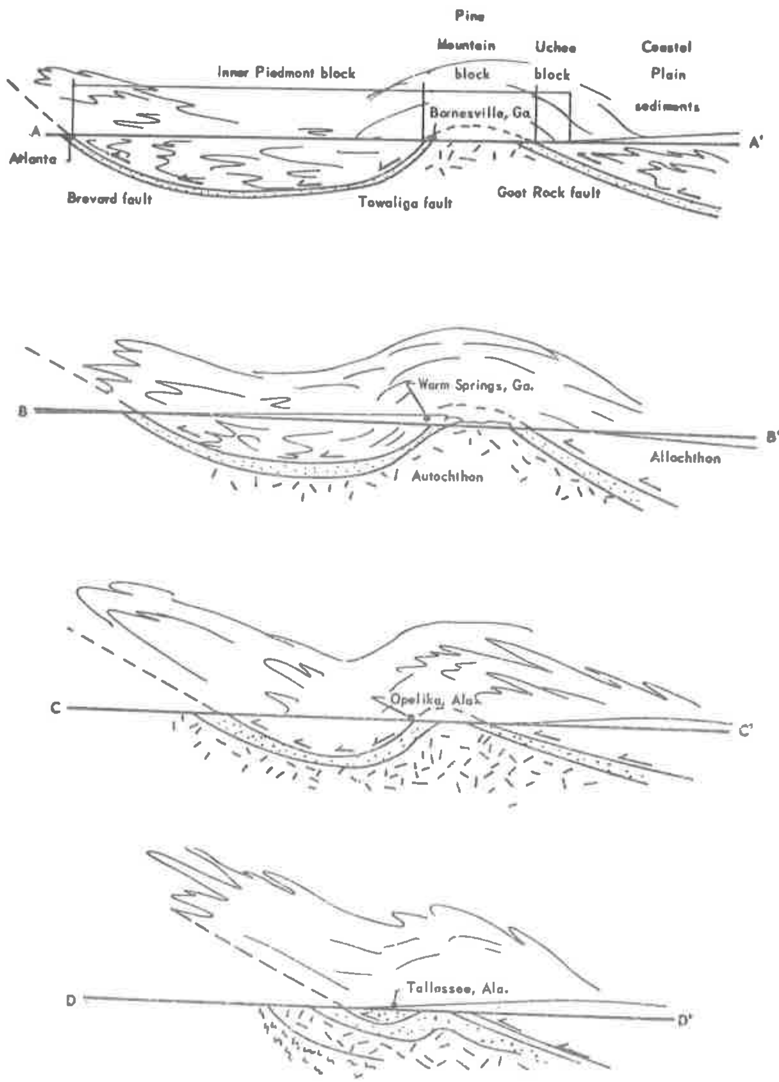

Figure 7. Cross-sections through portions of the Piedmont of Alabama and western Georgia showing the inferred relationships between the Brevard, Towaliga and Goat Rock faults (from Bentley and Neathery, 1970, Figure 17).

subduction zones. Continental material is prevented from being swallowed by its lower density. Tectonic events from successive orogenic periods are recorded in the continents while oceanic crust remains relatively young. Recent mountain belts were apparently generated by crumpling along a leading edge of an advancing plate or by collision of two continental masses (Dewey and Bird, 1970).

The question must be asked whether or not the concept of plate tectonics can be applied to all mountain systems of all ages or whether it should be restricted to recently formed orogenic belts. Due to the similarities of structures, metamorphism and rock types found in both recent and ancient mountain systems and their linear nature, many believe that the plate tectonics theory must also explain the more ancient mountain belts. Some have even called for rewriting all our textbooks with this new orientation and feel that this concept will be for geology what the atomic theory is to chemistry and physics and evolution is to biology (Wilson, 1968).

A detailed plate tectonic synthesis of the northern Appalachians and the Caledonides of western Europe has been presented by Bird and Dewey (1970). Brown (1970) has related the Late Precambrian to Late Ordovician history of the Virginia Piedmont to a downflowing oceanic crustal block (inclined northwestward) which is being overridden by continental material (Figure 8). Watkins and Huggett (1970) and Watkins (1971) have presented geophysical evidence favoring interpretation of the Brevard Zone and Kings Mountain belt as ancient southeastward dipping subduction zones.

Brown's account resembles that which could be proposed for the southern Appalachians (Hatcher, 1970b; in press). Such a model could consist of a Late Precambrian to Middle Ordovician phase of continental margin sedimentation and formation of a trench-island arc system followed by a Middle Ordovician to Late Devonian phase of compression derived from westward underflow of oceanic crust and producing isoclinal folding of the core, progressive regional metamorphism and intrusive activity with deposition from a rising tectonic land. The second phase could have resulted from westward underflow of a lithospheric plate beneath and eastward moving plate, of which the North American continent was a part. Destruction of the oceanic portion of the westward advancing plate occurred in a northwest dipping subduction zone which generated compression in the continent. This subduction zone was probably located west of the Carolina slate belt. Mid to Late Paleozoic compression and continued deposition resulting from North America-Africa collision produced the folds and faults of the Valley and Ridge, the Blue Ridge thrust sheet and the Brevard zone, Towaliga, Goat Rock and Gold Hill faults.

DISCUSSION AND SYNTHESIS

The southern Appalachians are a paradox. On the one hand part of this mountain system served as the cradle of some of our basic geologic concepts, while on the other there are large areas into which the geologist concerned with detailed mapping has never set foot. There are some controversies that have raged for years and other aspects that

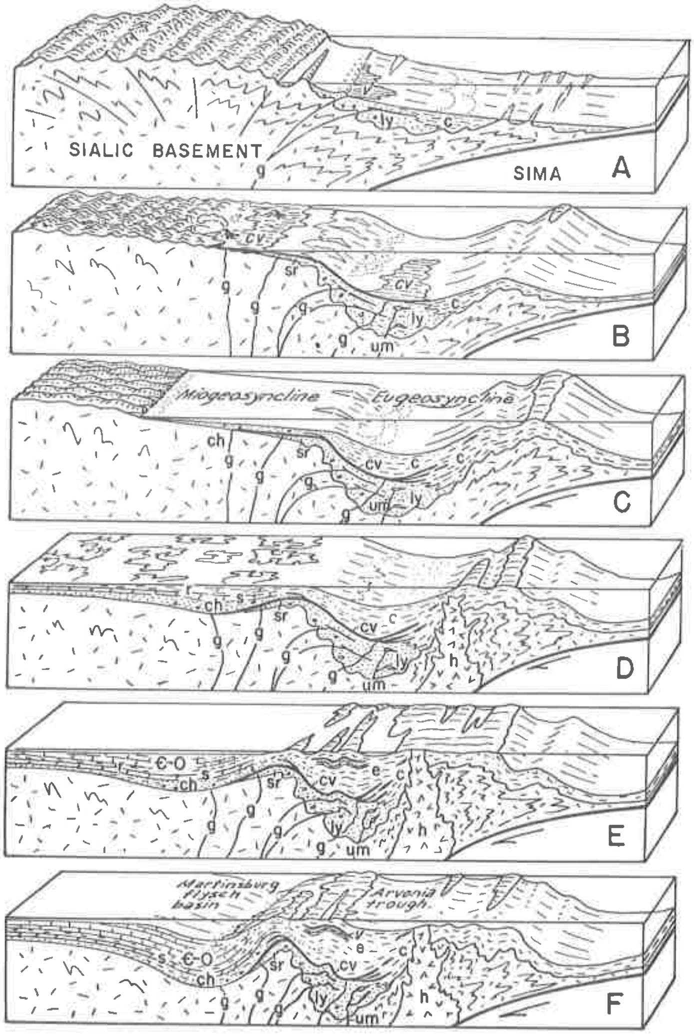

Figure 8. Late Precambrian to Late Ordovician history of the Virginia Appalachians: A. near the start of the Late Precambrian. B. near the end of the Late Precambrian. C. Early Cambrian. D. near the end of the Early Cambrian (Rome tidal flats). E. Middle Ordovician. F. Late Ordovician (from Brown, 1970, Figure 8).

we are just learning enough about to be able to discuss and argue.

In the Valley and Ridge the meticulous stratigraphic work of Byron Cooper has demonstrated that the miogeosynclinal sediments were not deposited in sheets but as masses of different thickness due to syndepositional vertical tectonic movements (Cooper, 1968). Cooper cites this as evidence for tectonism and mountain building throughout the Paleozoic, not just at the end. This latter work has resulted in a rebirth of the thick-skinned concept. However, others, such as King (1964, p. 24), state that differential subsidence during the Paleozoic is of the same type that occurs on the craton to the west resulting in a regional basin and dome structural pattern. Cooper (1968, p. 33) states that geophysical (gravity) studies are of limited value in solving the problems of basement involvement. Thus far the results of such geophysical studies are contradictory (see Watkins, 1964, and Sears and Robinson, 1970). The final solution to end all arguments regarding basement involvement in the Valley and Ridge would be provided by several basement holes drilled in critical areas. Thin-skinned structure has been proved by drilling in the Allegheny Plateau of Pennsylvania and West Virginia (Gwinn, 1964). More detailed mapping and sedimentological-stratigraphic studies in this area would be fruitful.

The tectonic history of the southern Appalachians is closely associated with the different episodes of regional metamorphism. In the Virgilina synclinorium in the Piedmont (along the North Carolina-Virginia border), metamorphism, folding, faulting and plutonic activity began as early as the Late Precambrian (Avalonian), some 620 to 570 m.y. ago (Glover and others, 1971). Major regional metamorphism occurred in the Blue Ridge more than 430 m.y. ago, some 430 to 410 m.y. ago in the Inner Piedmont and about 420 to 380 m.y. ago in the belts to the southeast. Major folding accompanied or slightly post-dated regional metamorphism. This Paleozoic event in the southern Appalachians confirms the occurrence of the Taconic Orogeny here (Butler, 1972).

The folding and thrusting in the Valley and Ridge and emplacement of the Blue Ridge thrust sheet may be dated as post-Carboniferous by fossils. The large faults (Brevard, Towaliga, Goat Rock, Gold Hill) to the southeast and the other structures associated with them are more difficult to date, although their principal movement history appears to be post- mid-Paleozoic (Hatcher, 1971b; Roper and Dunn, 1971). However, some geologists, such as Bentley and Neathery (1970) and Griffin (1969a, 1971b), believe these faults formed with the early structures. But there is presently a diversity of opinion as to the actual nature of these structures so there is no reason to expect agreement at present as to the timing of movement either.

Portions of the crystalline southern Appalachians have been the subject of several reconnaissance studies in recent years (for example, Overstreet and Bell, 1965; Cazeau, 1967; Bentley and Neathery, 1970; Hatcher, 1971a; and Hadley and Nelson, 1971). Such studies serve to either provide knowledge where none previously existed, to put forth working hypotheses to serve as ideas for future studies, to refine earlier reconnaissance studies by greater detail or by application of knowledge from other areas or new techniques. Therefore, they are of considerable value. However, for a precise solution to the problems of the crystalline Appalachians, detailed mapping on a scale of 1/24,000 or larger will be necessary. Data gathered by this tedious, painstaking and time-consuming method should be subjected to stratigraphic and petrologic as well as structural analysis. Where such integrated investigations have been made, the results have shed considerable light on both local and regional problems. But until a number of such studies have been made over large areas and the results published, we shall remain in the speculative realm of tectonic thought in the southern Appalachians.

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STRATIGRAPHY OF THE JACKSON GROUP IN EASTERN GEORGIA

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ABSTRACT

The Jackson Group in eastern Georgia is predominantly Upper Eocene in age. It is a transgressive-regressive sequence with a thin, extensively developed transgressive sand and a much thicker, more complex fine- to coarse-clastic regressive phase. In down-dip areas the group is represented by the Ocala Limestone; in up-dip areas, by fluviatile sediments indistinguishable from the late Cretaceous to possibly Middle Eocene Middendorf Formation. Between the down-dip marine limestone facies and the up-dip fluviatile facies occurs a lithologically complex near-shore facies, the Barnwell Formation. While general patterns of lithologic distribution can be recognized and the formation roughly divided into members, individual lithologic units are lenticular or deeply channeled and can not be traced over distances of more than a very few miles.

The transgression and regression represented by the Jackson Group in Georgia began much earlier in South Carolina, probably as the result of differential warping down to the northeast. Therefore, stratigraphic markers associated with the transgression, the basal zone of kaolin boulders in the coarse-clastic facies for instance, transgress time between central South Carolina and eastern to central Georgia. Recognition of this relationship permits correlations between South Carolina and Georgia that have previously been viewed as illogical or impossible.

INTRODUCTION

This report is an extension, based on several years of field and laboratory work, of an earlier paper on the stratigraphy of the Jackson Group in central Georgia. The history of nomenclature of the Jackson Group was presented in the earlier paper (Carver, 1966), and in several more-recent papers by others, and will not be extensively discussed in this report. The stratigraphic terminology employed in this paper is outlined in Figure 1.

The Jackson Group in eastern Georgia consists of clastics and

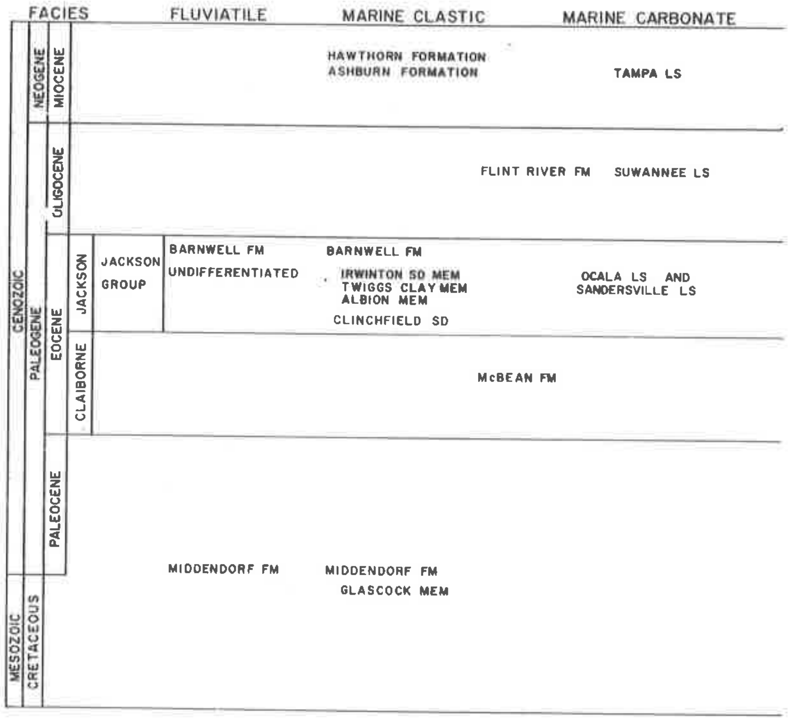

Figure 1. Stratigraphic terminology employed in this report. Time lines shown are approximate and vary along strike, as explained in text.

poorly indurated limestones in a classic transgressive-regressive vertical and lateral sequence. The basal, transgressive facies includes spiculitic clays and opal-cemented sandstones (Carver, 1968; Sandy, Carver and Crawford, 1966) in the extreme eastern part of the area, but more typically consists of cross-bedded sands with chips or boulders of kaolin. The overlying marine-clastic facies includes laminated to cross-bedded sands, but is characterized by opal claystone (Heron, Robinson, and Johnson, 1965, p. 59) of the Twiggs Clay Member of the Barnwell Formation. Lenses of opal-replaced, opal-cemented, sandy shell hash occur in the Twiggs Clay Member (Carver, 1969). Cross-bedded, gravelly sands of the Irwinton Sand Member typify the up-dip regressive phases of the sequence. Exposed parts of the carbonate facies consist dominantly of poorly indurated biomicrites.

The depositional strike of the Jackson Group, based on the location of known occurrences of spiculites (Figure 2), an apparent shoreline indicator (Cavaroc and Ferm, 1968), is approximately N80°E. The trend of the Fall Line between Macon and Augusta, Georgia, is N65°E, so that sections near the Fall Line at Macon represent more marine

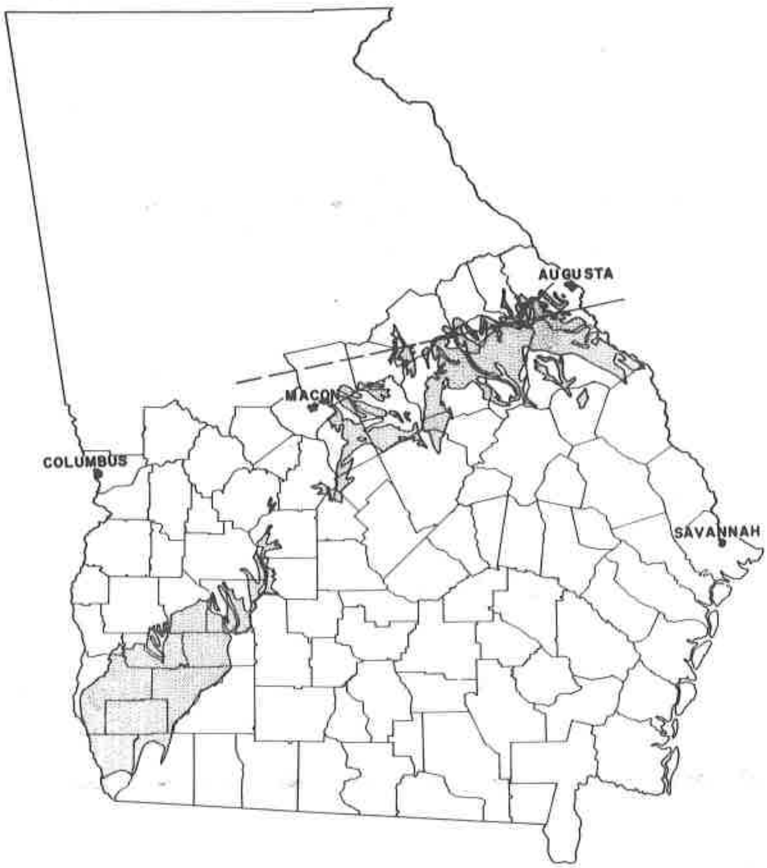

Figure 2. Outcrop area of the Jackson Group in Georgia (after Cooke, 1939). The three black dots near Augusta represent outcrops of the Albion Member and the line joining them is assumed to be the depositional strike of the basal Jackson Group in eastern Georgia.

facies (that is, originally deposited further off shore) than sections near the Fall Line at Augusta. A line of section extending from Section 1 (Figure 3) in central Georgia to Section 14 in western South Carolina therefore extends from a near-shore marine facies to a littoral or fluviatile facies of the Jackson Group.

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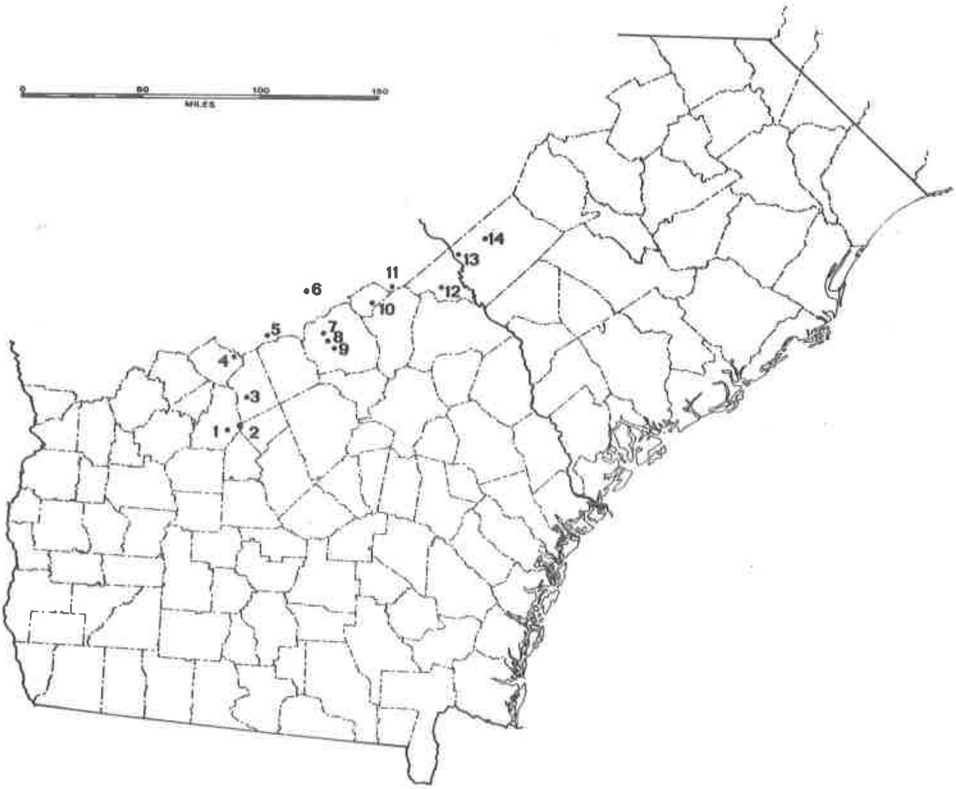

Figure 3. Coastal Plain areas of Georgia and South Carolina showing location of sections described in the text and appendix.

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LIMITS

The Stratigraphy of the Jackson Group is complex in detail and can be most directly and clearly approached through consideration of stratigraphic limits. Limits of a rock unit may include upper and lower contacts with other rock units, lateral pinchouts, facies changes (in the case of strictly defined lithologic units) and changes in terminology across administrative or philosophical boundaries. Limits of the Jackson Group include all of these.

Basal Contacts

Middendorf Formation: In the northern part of the outcrop area in eastern Georgia (Figure 2) the Jackson Group lies unconformably on igneous and metamorphic rocks of the Piedmont province or on the Middendorf Formation. LaMoreaux (1946) reported extensive channel sands in the east-central Georgia area and mapped them as Barnwell Formation units. The channels are cut into metamorphic and deeply weathered igneous rocks, or into the richly kaolinitic sediments of the Middendorf Formation. LaMoreaux states that kaolin pebbles and boulders up to 4 feet in diameter are characteristic of the channel sands. LeGrand (1956) noted the occurrence of a sandy gravel with kaolin pebbles on granites to the west and southwest of Warrenton, Warren County, and suggested that they represented outliers of the Barnwell Formation.

The criteria for distinguishing Middendorf Formation from Jackson Group sediments employed in this study were that massive kaolin and kaolin-matrix sands were uncontestably Middendorf; channel sands free of kaolin, or with kaolin chips, pebbles, or boulders only in the base, lying on richly kaolinitic sediment were uncontestably Jackson Group. Unfortunately, in up-dip sections the Middendorf and Jackson Group sands closely resemble each other and the criteria can not always be applied. A typical example involves the Albion Kaolin Mine (Section 12) and the Windsor Spring Road sections of Richmond County. Sandy, Carver, and Crawford (1966) noted that the contact of the Barnwell and Middendorf (Tuscaloosa) Formations at the Albion Mine was clearly definable as the contact between the flint kaolin (Glascock Member) and a spiculitic clay and opal-cemented sandy spiculite (Albion Member) that occupies depressions in an erosion surface developed on the flint kaolin, even though the spiculitic clay was kaolinitic. They therefore put the contact of the formations at the Windsor Spring Road section, about 8 miles northeast, at the base of a series of kaolinitic to opaline, spiculitic siltstones, 40 feet above a bed of cross-bedded sand containing large kaolin boulders. Herrick and Counts (1968), however, retained the kaolin boulder criterion and listed the sands below the siltstones in the Windsor Spring Road section as "Lower Jackson Sand (?)" and "Channel Sands (?)".

The first criterion, that the end of massive kaolin occurrence marks the top of the Middendorf Formation, works well only in kaolin mines. Buie and Fountain (1968) note the massive kaolin in the Middendorf Formation occurs at the tops of two sequences of medium- to coarse-grained, cross-bedded sand. The lower sequence is though to be of Cretaceous age, the upper sequence of Eocene, possible middle Eocene, age. The Eocene age of parts of the Middendorf Formation has been supported by Scrudato (1969) and Scrudato and Bond (1970). Where massive kaolin is not present, Middendorf sands commonly contain chips, pebbles, boulders, or thin stringers of kaolin and appear to be lithologically indistinguishable from channel sands of the Jackson Group.

The second criterion, that fragments of kaolin occur only in the lower part of the basal Jackson Group sands, appears to be broadly applicable to the marine-clastic facies of the group, but probably does not apply to the fluviatile facies (see Up-Dip Facies).

One additional criterion for recognition of the contact between the Barnwell and Middendorf Formation is proposed: Where the Albion Member, as defined in this report, is present, the base of the member should be considered the base of the Barnwell Formation and Jackson Group.

McBean Formation: The McBean Formation, in eastern Georgia, is overlapped by the Jackson Group and crops out only in northern Burke County, primarily along McBean Creek and the Savannah River. At the type section, a quarter mile southeast of McBean, Georgia, it consists of calcareous and carbonaceous sands, green clays and marls. Veatch and Stephenson (1911) originally defined the formation and considered that all of the Shell Bluff section below the upper sand was McBean. Cooke and Shearer (1919) redefined the McBean Formation to include only the beds below the *Ostrea georgiana* Conrad beds at Shell Bluff. The *Crassostrea gigantissima* (Finch), as *Ostrea georgiana* are now called (Palmer and Brann, 1965) normally occur in a matrix of gray clay typical of the Twiggs Clay Member, and Cooke and Shearer's correlation of the oyster beds with the Twiggs Clay appears reasonable.

Cooke and MacNeil (1952) further restricted the McBean Formation to equivalents of the Cook Mountain Formation (Mississippi) and the *Ostrea sellaeformis* zone of the Lisbon Formation (Alabama) noting that *Ostrea sellaeformis* Conrad occurs at Shell Bluff. While this is useful for establishing the age of the McBean Formation as late middle Claiborne, it does little to clarify the lithologic relationship between the Jackson Group and McBean Formation. In the area of the McBean type section the formation appears to be characterized by carbonaceous to lignitic and glauconitic sands, calcareous sands, sandy limestones, and true marls. On a purely lithologic basis, the difference between the Jackson Group and the McBean Formation sediments appears to be the presence, in the McBean, of abundant carbonaceous matter, glauconite, or rocks composed of mixed clastic and carbonate materials, none of which are common in the Jackson Group. A definition of the McBean based on these characters would results in the same division of the type section, the Shell Bluff section, and Griffins Landing section made by Cooke and Shearer (1919), LeGrand (1956), and Herrick and Counts (1968). If, in addition, the oyster beds are considered as part of the Twiggs Clay Member, a reasonable precise division of the section can be made anywhere in the general outcrop area of the McBean Formation in eastern Georgia. Pickering (1970) based his identification of the lower few feet of section (at the Penn-Dixie cement quarry, Clinchfield, Georgia) as McBean Formation essentially on these criteria.

Upper Contacts

Where it is not exposed to erosion, the Jackson Group is overlain by the Upper Oligocene Flint River Formation, or its down-dip equivalent, the Suwannee Limestone, or the overlapping Hawthorn Formation. As both LaMoreaux (1946) and LeGrand (1956) have remarked, it is difficult to distinguish between basal sands of the Flint River or Hawthorn Formations and upper sands of the Jackson Group in areas where they are heavily weathered. However, distinctive lithologies occur in each of the overlying formations and it is possible to determine the upper limit of the zone of uncertain correlation in most areas. The basic problem is that good sections including the contact with the Jackson Group and significant thicknesses of the Flint River or Hawthorn Formations appear to be rare or non-existent, and the exact position of distinctive lithologies within the two formations is not now known.

Flint River Formation: The Flint River was named for large boulders of flint that nearly fill the river channel in some areas of southwest Georgia. The upper Oligocene Flint River Formation was named by Cooke (1935) for exposures along the river near Bainbridge, Georgia, and large flint boulders are everywhere characteristic of the formation. The boulders commonly accumulate in alluvium and colluvium below their original stratigraphic position, but in the upland areas they are excellent stratigraphic markers. There seems to be little question that the chert boulders are residuum from the weathering of a cherty limestone, but it is difficult to say whether they originally developed as irregular, boulder-shaped masses in the limestone, or are fragments of thick beds of chert. At Stony Bluff on the Savannah River, in extreme southeastern Burke County, exposed beds of chert are not more than 2 feet thick and are interbedded with sand. Herrick and Counts (1968) find that a 1 to 2 foot thick bed of chert occurs at the top of the formation in many places, but many of the chert boulders are greater than 2 feet in minimum dimension.

Given the uncertainty about the stratigraphic distribution of chert in the Flint River Formation, chert boulders, in place, mark a horizon that is 0 to 200 feet above the top of the Jackson Group, based on the range of thickness of the Flint River Formation given by Stringfield (1966). However, observation of chert boulders and cobbles on the upland a short distance above Sections 1 and 2 (appendix) in Houston County indicates that the chert occurs well down in the Flint River Formation in the central Georgia area, and the first occurrence of chert in place is probably within 30 feet of the top of the Jackson Group.

Hawthorn Formation: The Hawthorn Formation, of Miocene age, overlaps the Flint River Formation and lies directly on the Jackson Group over a wide area between the Oconee and Savannah Rivers.

Typically, the formation consists of irregularly interbedded sandy to silty clays and gravelly sands that are not easily distinguished from upper sands of the Jackson Group. However, a distinctive hard, sandy claystone occurring at the base of the Hawthorn Formation appears to be a reliable stratigraphic marker over much of the area. Olson (1967) informally named this unit the Ashburn Formation, for outcrops along the West Fork of Deep Creek at Interstate Highway 75 four miles north of Ashburn, Turner County, Georgia, and considered it a fluviatile facies of the early Miocene Tampa Limestone.

The Ashburn Formation appears to be an excellent stratigraphic marker for the top of the Jackson Group in the area where the Ashburn Formation overlaps the Flint River Formation, especially in the area around Millen, Georgia. However, it is possible that the Hawthorn Formation overlaps the Ashburn unit in updip areas, or that typical Hawthorn lithologies underlie the Ashburn Formation in some areas. In this case, it would be difficult to distinguish Hawthorn Formation sands from upper sands of the Jackson Group.

Up-Dip Facies

The up-dip equivalents of the Middendorf Formation, McBean Formation, Jackson Group, and Hawthorn Formation probably are fluviatile gravelly sands and gravels which are indistinguishable from each other. LeGrand (1956, p. 39) described gravels with a sand matrix in a discontinuous belt extending from near Barnett in Warren County south-southeast to Glascock County. LeGrand considered the gravels, which contain kaolin balls, as a facies of the Barnwell Formation because the Middendorf Formation is extensively overlapped in the area. Based on the locations of entirely marine sections, it does appear that the most extensive transgression of the Coastal Plain of Georgia occurred during deposition of the Jackson Group (late Eocene time). The greater part of the up-dip coarse clastics may well correlate with the Jackson Group.

Down-Dip Facies

Clastic facies of the Jackson Group grade into carbonates in the downdip direction, as do the clastic facies of the younger and older formations. The Ocala Limestone is the carbonate facies of the Jackson Group and in eastern Georgia consists primarily of micritic limestones or biomicrites composed primarily of bryozoan fragments and echinoids. Outcrops of the Ocala Formation were mapped by Cooke (1939) as the Cooper Marl, but it can be clearly demonstrated that the limestones in question correlate with the Barnwell Formation and Ocala Limestone in east central (Carver, 1966) and east Georgia. Time correlatives of the lower Oligocene Cooper Marl, of the Charleston, South Carolina, area do not appear to occur in the upper Coastal Plain of Georgia, but may be present along the Ocmulgee River in the area south of Sections

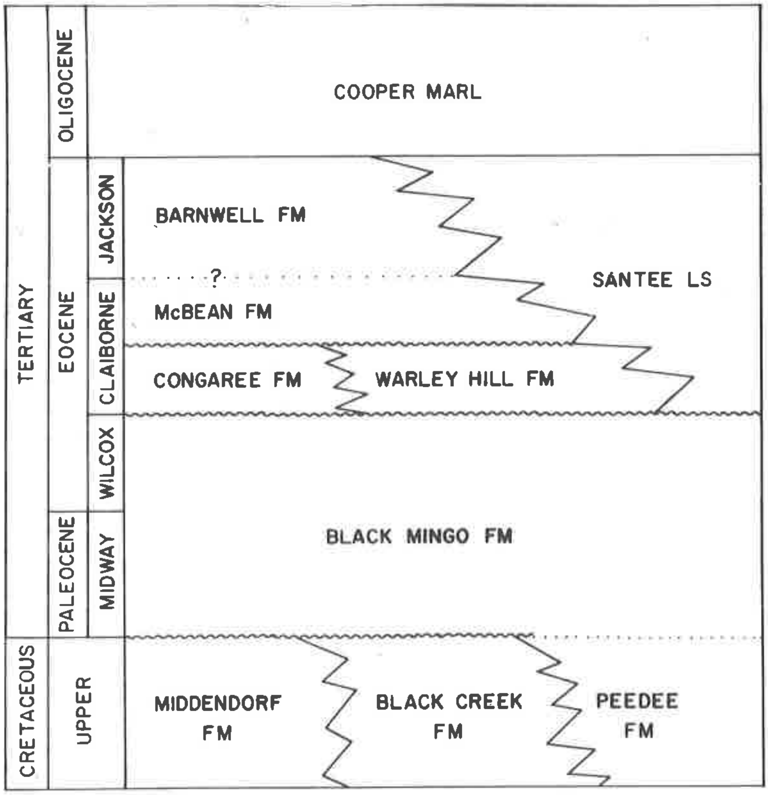

Figure 4. South Carolina Cretaceous to Oligocene stratigraphic terminology as presented by Colquhoun et al. (1969).

1 and 2, as suggested by Pickering (1970).

In the Savannah area the Upper Miocene Hawthorn Formation is present and is underlain by a hard, sandy limestone with abundant fossil molds, the Suwannee Limestone. The upper part of the Ocala Limestone in this area is soft limestone consisting primarily of bryozoan fragments in a sparse micritic matrix.

South Carolina Nomenclature

The Paleogene stratigraphy of South Carolina has recently been reviewed by Colquhoun, et al. (1969), and is briefly summarized in Figure 4. Based on the opal claystone and kaolin boulder criteria used in this study, the upper part of the Black Mingo Formation and the Santee Limestone and all of its up-dip equivalents are physically correlative with the Jackson Group in Georgia. Heron, Robinson, and Johnson (1965) and Heron (1969) have reported the extensive occurrence of opaline sediments in the upper part of the Black Mingo Formation, and a persistent coarse-clastic unit containing pisolitic kaolin cobbles and boulders has long been recognized as a marker for the base of the Congaree Formation (see Cooke and McNeil, 1952, p. 23).

Correlation of the Jackson Group in Georgia with the upper part of the Black Mingo Formation and the Santee Limestone is, at first glance, inconsistent with the ages assigned to the units in South Carolina and Georgia. If the Middendorf Formation in eastern Georgia ranges up to Claiborne age, it is correlative, in time, with the Middendorf, Black Mingo, and Congaree Formations of South Carolina, but lithologically correlative only with the Middendorf Formation of South Carolina. In addition, the McBean Formation in South Carolina would lie above the base of the Jackson Group, rather than below the Group, as in eastern Georgia.

The paradox is easily resolved. It is recognized that the Jackson Group in Georgia is the result of a major transgression and regression (Carver, 1966, 1968), as are the middle Eocene to Middle Miocene sediments of South Carolina (Colquhoun, et al., 1969, p. 12). The stratigraphic markers used to define lithologic units are related more to shoreline position than to times or events and are time transgressive. The Paleogene transgression began earlier in South Caroinla than in Georgia and coastal-zone sediments of the Black Mingo Formation were being deposited in South Carolina while deposition of the fluviatile Middendorf Formation continued in Georgia. Eastern Georgia was a positive area, relative to the areas to the east and west, at least until late Eocene time, and the Paleogene transgression is represented, in Georgia, by a much thinner, less complex, series of lithologic units. This relationship is roughly indicated by the correlation chart, Figure 5.

PRINCIPAL LITHOLOGIC UNITS

Clinchfield Sand

The name "Clinchfield sand" was informally introduced by Vorhis (1965) and was used in an informal sense by Carver (1966) in reference to the basal sand of the Jackson Group in the marine, or calcareous, facies. The unit is exposed in Sections 1, 2, and 3 of this report. In Section 3 the full thickness (about 19 feet) of white to yellow-brown, calcareous sand is exposed in contact with massive white kaolin of the Middendorf Formation. Sand dollars, solitary corals (*Flabellum cuneiforme* Lonsdale), molluscs and microfossils are common to very abundant in the section, indicating a near-shore marine environment of deposition.

The exposures in the cement-rock quarry of the Penn-Dixie Cement Company at Clinchfield, Georgia probably will be designated the type section of the Clinchfield Sand. In a drainage ditch in the quarry

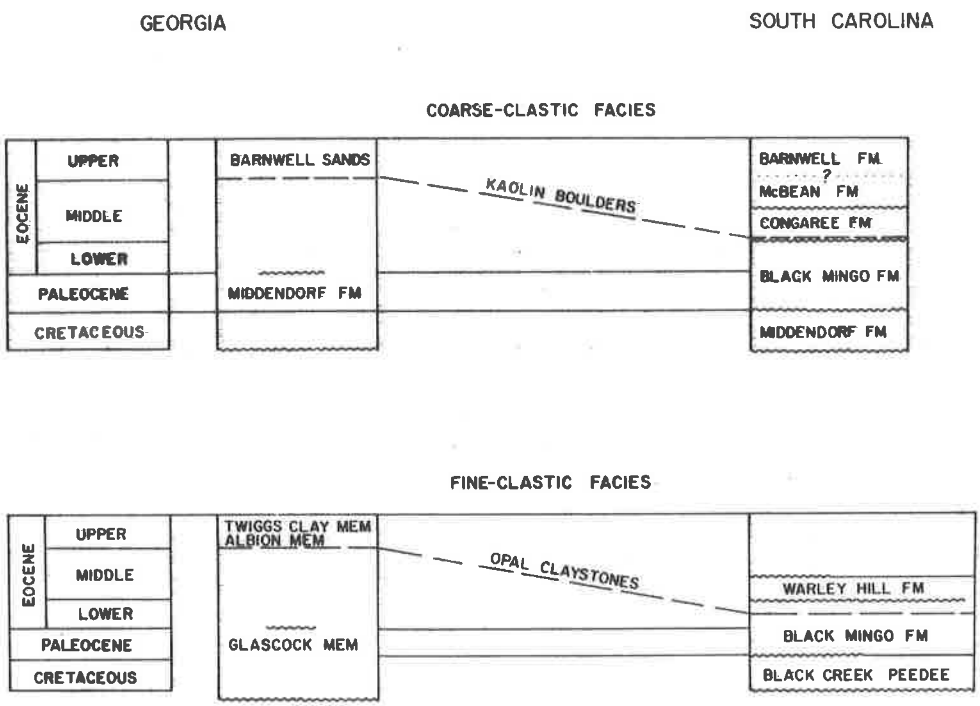

Figure 5. Correlation between Georgia and South Carolina Eocene stratigraphy based on important stratigraphic markers in eastern Georgia and western South Carolina. The first occurrence of opal claystones in the fine-clastic facies and the kaolin-boulder zone both transgress time toward the west.

pit west of U. S. Highway 341, about 10 feet of structureless brown sand is exposed.

In Section 2, on Georgia Highway 247 northeast of Hawkinsville, the Clinchfield Sand is poorly exposed along Big Indian Creek.

Albion Member

The Albion Member of the Barnwell Formation was informally defined in Sandy, Carver and Crawford (1966). The type section of the Albion Member is here designated as the section exposed in the Albion Kaolin Mine approximately 1 mile west of Hepzibah, Richmond County, Georgia (Section 12 of this report). The Albion Member consists of discontinuous lenses of spiculite, spiculitic clay and mudstone and opal-cemented sandy spiculite and spiculitic sandstone (Buie and Oman, 1963; Carver, 1968, 1966). Typical exposures of spiculite and spiculitic clays occur in the Albion Mine and in the section along Windsor Spring Road a mile north of Tobacco Road in Richmond County (Sandy, Carver and Crawford, 1966; Herrick and Counts, 1968). Opal-cemented rocks occur in the Albion Mine, in the abandoned Harbison-Walker Co. flint-kaolin pit near Gibson, Glascock County (Sandy, Carver and Crawford, 1966), and as float along Brier Creek at the crossing of U. S. Highway 1 on the west boundary of Richmond County.

Fine-grained sediments of the Albion Member normally are massive to poorly bedded and megascopically structureless, but intraformational conglomerates (with 1 to 4 mm diameter fragments) occur in the Windsor Spring and Gibson sections. X-ray diffraction indicates that the underlying Middendorf Formation consists of mixtures of kaolinite and quartz, but the spiculitic clays and mudstones of the Albion Member are mixtures, in varying proportions, of opal, montmorillonite, quartz and kaolinite, with opal and montmorillonite dominant in most cases. The change in clay mineralogy across the Middendorf-Barnwell Formation boundary in the Albion Member outcrop area is evidence for a significant change in source materials, perhaps the introduction of volcanic ash, or a change in weathering conditions in the source area, and substantiates the importance of the unit as a stratigraphic marker.

The mudstones and clays of the Albion Member contain abundant sponge spicules, which are one source of opal in the sediment. Separation and low-power microscopic examination of the coarser fractions of the spiculitic clays reveals that the tripartite spicule terminations of class Hyalospongia are common. Hyalospongia are an entirely marine class of sponges, and the presence of hyalospongia firmly establish the marine character of Albion Member sediments.

Opal-cemented units of the Albion Member are dominantly opal with varying amounts of clastic quartz and minor amounts of montmorillonite and kaolinite. One lens of opal-cemented sandstone in the Albion Mine grades laterally into marcasite-cemented gravel with montmorillonite fragments, pinching-out within a few feet of the lateral change in grains size. Silicified (opalized) plant fragments and carbonaceous matter are abundant in lower parts of the unit. Taken together, the thin, discontinuous nature of the units, the abundance of plant fossils, and the abrupt lateral transition to marcasite-cemented gravels strongly indicate an extreme near-shore, perhaps tidal pool, origin for the opal-cemented clastics and spiculites of the Albion Member of the Barnwell Formation. This interpretation confirms the observation of Cavaroc and Ferm (1968) that richly spiculitic sediments are indicators of shoreline environments.

Twiggs Clay Member

The Twiggs Clay Member of the Barnwell Formation was named by Cooke and Shearer (1919) for exposures in a fuller's earth mine near Pikes Peak Station in northern Twiggs County. Lamoreaux (1946) found it to be widespread in east-central Georgia, lying unconformably on the Middendorf Formation where basal channel sands were not present in the Barnwell Formation. LeGrand (1956) traced the member into

In Section 4 a fossiliferous sandstone cemented by opal occurs as lenticular beds up to 2 feet thick near the base of the Irwinton Sand. Examination of thin sections reveals the following paragenesis: (1) compaction of the sediment accompanied by fracture of shell fragments and Riecke-effect solution of shell carbonate at grain contacts. (2) Deposition of light-colored, isotropic opal cement and dark, optically fibrous, slightly birefringent opal cement in three or more episodes, before and after additional fracture of shell fragments by compaction. (3) Replacement of the greater part of the shell material by opal with preservation of conchiolin and, consequently, shell structure. Unfortunately, replacement of shell carbonate and cementation of grains by opal are non-interfering, or petrographically independent events and their relative order cannot be determined. (4) Replacement of remaining carbonate by microcrystalline quartz with preservation of shell ultrastructure, but without preservation of conchiolin and filling, or partial filling, of remaining voids by chalcedonic quartz.

The several episodes of cementation, occurring during and after final compaction of the rock, indicate that the opal was deposited over a long period of time and after burial of the sediment. Preservation of conchiolin during opal-replacement of the shell carbonate suggests reducing conditions. A change from reducing to oxidizing conditions during the final phase of shell-replacement, by microcrystalline quartz, is indicated by the absence of conchiolin in this material. While the opal is, in this case, late diagenetic and secondary, opal in the Albion Member and Ashburn Formation is primary (Carver, 1969) which leaves the question of the origin of opal in opal claystones unsettled. Heron (1969, p. 38-39) reviewed the reported distribution of opal in Tertiary sediments of the southeast and concluded, on the basis of his studies of zeolites and opal in the Black Mingo Formation, that the ultimate source was volcanic ash.

In most sections of the Irwinton Sand fossils are entirely absent. Pelecypods and gastropods are abundant, however, in lenses of silicified shell hash or fossiliferous sandstones like the one described above, and sponge spicules are not uncommon. Unit 13 of Section 3 contains sponge spicules, foraminifera, ostracods, bryozoans and abundant oysters. The absence of bioturbidation in the thinly interlaminated sand and clay lithology is remarkable. Sandy clays of the Twiggs Clay member appear to have been thoroughly worked by burrowing organisms, but the laminated sediments of the Irwinton Sand have not. The general scarcity of fossils, the absence of burrowing and the dominance of environmentally tolerant types where fossils do occur suggest a harsh or variable environment of deposition. This, along with the evidence of highly variable currents in channeled sands and interlaminated sands and clays indicate lagoonal to estuarine, or perhaps tidal-marsh sedimentation.

Upper Sand Member

LaMoreaux (1946) noted the presence of an upper member of the Barnwell Formation consisting of coarse, subangular to subrounded quartz sand with flat quartz pebbles 1/4 to 2 inches in greatest dimension scattered along the base of the unit. As LaMoreaux notes, the unit is so badly weathered in most areas, even in machine-made cuts, that it is difficult to identify or characterize over much of the area of east Georgia. LeGrand (1956, p. 37) recognized the unit in central-east Georgia and described it as being "... typical of the undifferentiated Barnwell Formation of Barnwell County, S. C."

Where freshly exposed, as in Section 2 in 1966, the unit can be seen to consist of white to yellow, coarse- to medium-grained, cross-bedded sand with abundant pebble lenses. In 1966, when Section 2 was originally measured, parts of the machine-made cuts were covered by 12 to 18 inches of weakly iron-oxide-cemented crust that effectively obscured the primary clastic sedimentary structures. At that time the crust was soft enough, and thin enough, to be cut with a trenching tool, but at this writing, only four years later, the deep-red crust is to thick and tough that a major effort would be required to cut through it with hand tools. Rapid development of tough iron-oxide cement is common in all permeable rocks in near-surface exposures in the eastern Georgia Piedmont and Coastal Plain and it occurs on vertical as well as horizontal surfaces. The process is essentially the development of the "B" soil horizon.

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.

Barnwell Formation Undifferentiated

Definable members of the Barnwell Formation all pinch out in the up-dip direction and are replaced by medium- to coarse-grained, gravelly, cross-bedded sands containing kaolin balls, chips or stringers; or medium- to coarse-grained, structureless to cross-bedded sand with pebble stringers. The upper members extend farthest up-dip, the eastern Georgia, where it is thinner and more lenticular than in east-central Georgia. The discontinuous nature of the unit is indicated by the fact that LeGrand did not observe the Twiggs Clay in the Albion Kaolin Mine (Section 12 of this report), where later mining revealed a lens of the typical lithology over 40 feet thick at its maximum exposure.

In the area of its type section, the Twiggs Clay is a gray to green, silty to sandy, hackly clay, weathering to shades of light brown and gray. Disseminated sand, thin stringers of fine-grained white sand, and thin beds of opal-replaced sandy shell hash are common. In road cuts and pit walls the unit normally appears massive to indistinctly bedded. The tough, hackly chips produced on weathering of machine-made cuts are distinctive. Where cuts are covered by vegetation, as along Interstate Highway I-75 south of Perry, Houston County, Georgia, and at Section 2 of this report, presence of the Twiggs Clay Member is revealed by large slumps.

Mineralogically, the Twiggs Clay consists of varying amounts of clastic quartz and biogenic calcite in a matrix of montmorillonite and opal with minor amounts of kaolinite. The sediments are opal claystone as described by Heron, Robinson, and Johnson (1965). Their analyses of clay from a now inactive fuller's earth mine near Jeffersonville, Twiggs County, Georgia, (Table 1) are typical of the Twiggs Clay. X-ray diffraction patterns made in connection with this study indicate that the claystones range in composition from nearly pure opal to nearly pure montmorillonite, as observed by Heron (1969) for the Black Mingo Formation of South Carolina. In general, up-dip sections tend to be sandy and opaline, down-dip sections calcareous and montmorillonitic. Samples of commercial claystone from the Georgia-Tennessee Mining and Chemical Co. pit at Wrens, Jefferson County, Georgia are nearly pure opal with minor amounts of montmorillonite, kaolinite and quartz. Clays from the upper part of Section 1 and 3 consist of montmorillonite and calcite, with minor amounts of opal and kaolinite.

Sponge spicules, echinoid spines, foraminifera and ostracods are abundant in some sections. Larger fossils include small, simple, thin-shelled pelecypods that appear to be ubiquitous, and less commonly, the solitary coral <u>Flabellum cuneiforme</u>, bryozoans, high-spired gastropods, pectenoid pelecypods, oysters and shark teeth. Carbonaceous matter, probably finely divided plant material, is common. The presence of corals, bryozoans and echinoids indicates a normal marine environment of deposition for much of the sediment. However, a few units (i. e. unit 5, Section 8) with a restricted fauna, commonly consisting only of the small pelecypods mentioned above, may represent estuarine or tidal-flat environments of deposition.

Irwinton Sand Member

LaMoreaux (1946) named the Irwinton Sand Member of the Barnwell Formation for exposures in gullies 0.3 miles south of the

Table 1. Mineralogical Analysis of Opal Claystones From Georgia. With the exception of analyses 1 through 3, all analyses were based on peak-height ratios of the minerals tabulated. Analyses 1 through 3 were of the less than 2 micron fraction (see note below).

Sample Number	Opal	Montmorillonite	Kaolinite	Illite	Quartz
1	25	71	4	0	
2	38	56	3	3	
3	41	53	2	4	
4	55	10	6	0	30
5	20	30	4	0	46
6	32	41	0	0	27
7	39	31	0	0	30
8	27	23	0	0	50

Samples 1 through 3. General Reduction Co. pit near Jeffersonville, Twiggs County, Georgia, from Heron, Robinson and Johnson (1965, p. 59).

Sample 4. Georgia-Tennessee Clay and Chemical Co. pit at Wrens, Jefferson County, Georgia.

Sample 5. Twiggs Cl. from near Section 4 of this report.

Sample 6. Twiggs Cl. from near Fitzgerald, Twiggs County, Georgia.

Sample 7. Fuller's earth pit 1 mile north of Pikes Peak, Twiggs County, Georgia.

Sample 8. Twiggs Cl. from borrow pit 6 miles east of the Ocmulgee River on Georgia Highway 96, Twiggs County, Georgia.

courthouse at Irwinton, Wilkinson County, Georgia. Typically it consists of cross-bedded, channeled, white to yellow, fine-to medium-grained sand with carbonaceous matter and tough clay partings. A common and distinctive lithology is thinly interlaminated sand and clay with individual laminae averaging as little as 0.25 inch in thickness. LeGrand (1956) described this lithology in exposures along U. S. Highway 1 on the north side of Brush Creek in Wrens, Jefferson County, Georgia.

The clays are mineralogically similar to the Twiggs Clay, but with higher average percentages of montmorillonite and much lower percentages of opal. An exception occurs in Section 11, an up-dip section in which the typical lithology lies near the contact with the Middendorf Formation and the Twiggs Clay is entirely absent. Here the clays are kaolinite with minor amounts of illite. This is a typical mineralogical facies change for the Jackson Group as a whole, with kaolinite replacing montmorillonite in the up-dip direction.

Twiggs Clay overlapping the Albion Member and the Clinchfield Sand and the Irwinton Sand overlapping the Twiggs Clay. In extreme up-dip sections, members of the Barnwell Formation cannot be distinguished and the undifferentiated Barnwell Formation is indistinguishable from the Middendorf Formation, which in Jones, Baldwin, Hancock, Warren and Columbia Counties consists of sands bearing kaolin fragments and stringers.

Fossils are absent in the undifferentiated Barnwell Formation. The prevalence of high-angle cross-bedding, channeling and very poor sorting suggest a fluviatile origin for most, or all, of the units in Sections 6, 11, and 14, which are typical.

Sandersville Limestone and "Cooper Marl"

The Sandersville Limestone was named by Cooke (1943) for exposures in a sinkhole southwest of Sandersville, Washington County, Georgia. LaMoreaux (1946) considered the Sandersville Limestone an equivalent of the Irwinton Sand, and identified outcropping limestones in Twiggs County as Ocala Limestone. Cooke (1939) mapped extensive areas of outcrop of Cooper Marl west and east of the Ocmulgee River and an inlier of Cooper Marl in the Hawthorn Formation at Magnolia Springs, Jenkins County. Carver (1966) correlated the limestones, identified as Cooper Marl by Cooke, with the Ocala Limestone and suggested that the name Cooper Marl be reserved for stratigraphic and lithologic equivalents of the Cooper Marl at Charleston, South Carolina.

The Sandersville Limestone, the limestone cropping out at Magnolia Springs, and limestones cropping out along the Ocmulgee River north of Hawkinsville, Pulaski County, are sparsely fossiliferous micritic limestones that do not closely resemble the biomicrite of the Ocala Limestone in Section 1. Herrick (1961) reports marl (soft, micritic limestone) over an extensive area of the subsurface Jackson Group of eastern Georgia. Herrick's data, summarized in Figure 6, indicate that the micrites and micritic limestones form a continuous and consistent unit which probably is identical to the Sandersville Limestone and correlative with the entire thickness of both the Barnwell Formation and the Ocala Limestone. It is therefore suggested that the name Sandersville Limestone be applied to all subdivisions of the Jackson Group which consist predominantly of soft, yellow to white, micritic limestone.

The Sandersville Limestone southwest of Sandersville contains sand dollars, pectenoid pelecypods and abundant molds of pelecypods and gastropods. At Magnolia Springs the limestone is sparsely fossiliferous, with foraminifera, bryozoans, ostracods and pectenoid pelecypods. The fossil suite and micritic texture suggest a normal marine environment with gentle currents, beyond the reach of fine clastics from the Piedmont source area.

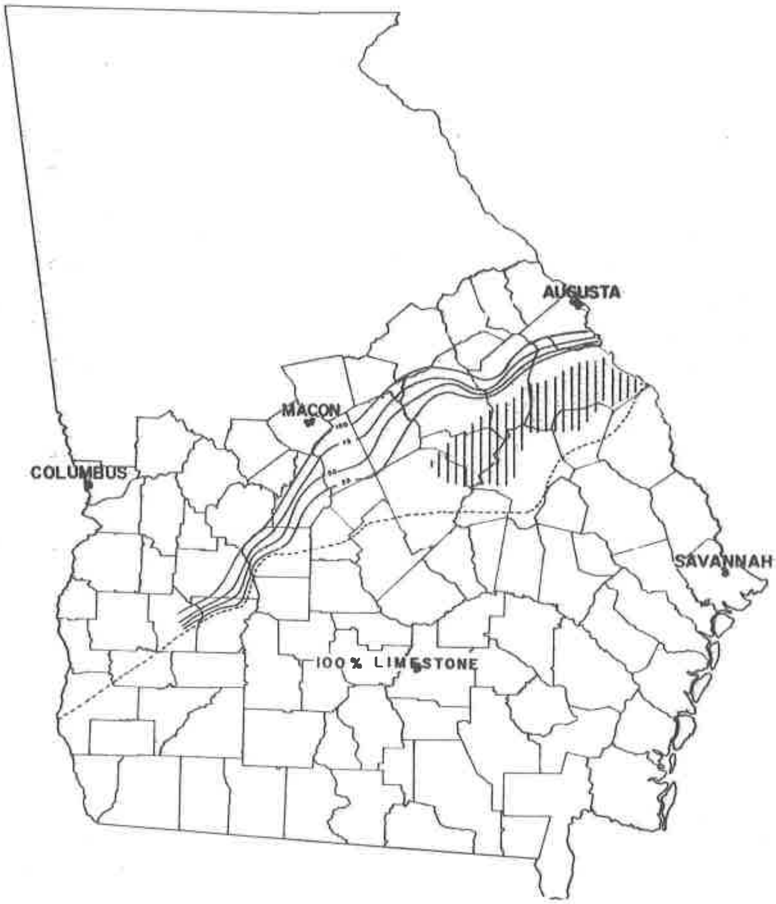

Figure 6. Lithofacies map of the Jackson Group based on subsurface data from Herrick (1961). Solid lines represent clastics by 25 percent intervals. Vertical hatched area indicates marl (micritic limestone) dominant in section. In areas south and southeast of the dashed line the Jackson Group is composed entirely of limestone.

Ocala Limestone

The Ocala Limestone named by Dall and Harris (1892) for exposures around Ocala, Florida, is the dominant unit of the Jackson Group in the subsurface of eastern Georgia and the major unit of the Coastal Plain artesian aquifer. In Section 1 it consists of soft biomicrite composed principally of bryozoan fragments with abundant sand dollars. In a core hole at Petit Chou Island, near Savannah and 165 miles ESE of Section 1, the lithology appears identical. Limestone consisting predominantly of bryozoan fragments may be characteristic of the Ocala Limestone in eastern Georgia, but outcrops are so rare and core-hole data so limited that sweeping generalizations about the Ocala Limestone are not warranted at this time.

The dominance of fossils belonging to stenohyaline groups indicates a normal marine environment of deposition for the Ocala Limestone. Fragmentation of bryozoans and scarcity of micrite matrix suggests moderately high currents and shallow water sufficiently far offshore to prevent transportation of terrigenous clastics into the area.

GUIDE FOSSILS

Ostrea gigantissima (Finch)

The long narrow oyster _Ostrea gigantissima_ (Finch), or _Crassostrea gigantissima_ (see Howe, 1937, and Palmer and Brann, 1965 for the history of nomenclature and synonomy) has long been considered, at least informally, as guide fossil for the Jackson Group in Georgia. However Lawrence (1968) reports the species in the Oligocene of North Carolina, and the species may be long ranging, through the Paleogene and perhaps into the Neogene of the Atlantic Coastal Plain. _Ostrea gigantissima_, therefore, probably is not a valid guide fossil for the Jackson Group.

Flabellum cuneiforme (Lonsdale)

The scleractinian coral _Flabellum cuneiforme_ is widely distributed in marine lithostratigraphic units of the Jackson Group (Sections 1, 3, 7 and 12). The fossil was originally reported from the Upper Eocene of Georgia and North and South Carolina (Lonsdale, 1845) and may be restricted to sediments of Upper Eocene age. However, sediments younger or older than the Jackson Group in the area of study appear to have been deposited in fluviatile to estuarine environments that would have been hostile to the presumably stenohyaline _F. cuneiforme_. The question of the absolute range of _F. cuneiforme_ in eastern Georgia therefore is open to question, but the existing evidence indicates its restriction to the Jackson Group and Upper Eocene Stage.

SUMMARY

The depositional strike of the Jackson Group in eastern Georgia, based on locations of known outcrops of the Albion Member of the Barnwell Formation, is approximately N80°E. Because the trend of the Fall Line between Macon and Augusta is about N65°E, exposures of the Jackson Group near Macon represent more typically marine facies than do outcrops near Augusta. Therefore, a line joining Sections 1 and 13 of Figure 3 runs diagonally up-dip from an entirely marine to an entirely fluviatile, or at least littoral, facies of the Jackson Group.

In down-dip sections in eastern Georgia the Jackson Group lies on calcareous, glauconitic, carbonaceous clastics of the McBean Formation (Claiborne Stage). Up-dip the McBean Formation is overlapped and the Jackson Group lies unconformably on the late Cretaceous to Paleocene Middendorf Formation, becoming indistinguishable from the Middendorf in the fluviatile facies. Over most of the area of outcrop of the Jackson Group the Middendorf-Barnwell Formation contact is marked by a change from richly kaolinitic sands to sands that contain kaolin clasts and boulders in the basal part only, grading upward into montmorillonitic or opaline sands and clays. In the marine facies the basal, transgressive sand is recognized as the Clinchfield Sand Formation (as named by Vorhis, 1965).

In eastern Georgia the Jackson Group is overlain, perhaps conformably in the southernmost part of the area of outcrop, by progressively overlapping formations of upper Oligocene and Miocene age. The upper Oligocene Flint River Formation, characterized by large boulders of residual chert, overlies the Jackson Group and crops out along the Savannah and Ocmulgee Rivers but in the area between the rivers is extensively overlapped by the late Miocene Ashburn (Olson, 1967) and Hawthorn Formations. Where the Flint River Formation is absent, the hard, sandy kaolinitic claystones and associated lenses of sandy opal of the Ashburn Formation appear to be an excellent marker for the top of the Jackson Group.

To the northwest, in South Carolina, the lateral equivalents of the Jackson Group in Georgia are the entire Middle and Upper (Claiborne and Jackson) Eocene stages. That is, approximately the same sequence of sedimentary events occurred in Georgia and South Carolina, but while this sequence began in middle or early Eocene time in South Carolina, it did not begin until late Eocene time in Georgia. Thus important stratigraphic markers associated with the Eocene transgression (kaolin boulders at the base of the coarse-clastic faces, a shift from kaolinite to montmorillonite as the dominant clay mineral, and deposition of primary and secondary opal) cross time lines as they are traced along strike. The transgressive-regressive sequence represented by the Jackson Group in Georgia is the time equivalent of the Barnwell Formation and upper Santee Limestone in South Carolina; but the lithologic equivalent of the upper part of the Black Mingo Formation, all of the Congaree, McBean and Barnwell Formations and the entire thickness of the Santee Limestone.

The foregoing correlation of time and events implies that central Georgia was a structurally positive area, relative to South Carolina, during most of Eocene time, and this probably holds true for Alabama also.

The Jackson Group in eastern and central Georgia includes the

Ocala Limestone, Sandersville Limestone, Clinchfield Sand and Barnwell Formation. The marine carbonate facies includes the Ocala Limestone, typically a poorly cemented hash of bryozoan fragments with abundant sand dollars, and the Sandersville Limestone, typically micritic with abundant fossil molds. In the near-shore marine facies the basal, transgressive Clinchfield Sand underlies the members of the Barnwell Formation. The Albion Member consists of spiculite, spiculitic clays and opal-cemented sandstones and occurs at the base of the Barnwell Formation in the eastern part of the area of outcrop. The Twiggs Clay Member is characterized by hard, opaline, montmorillonitic clays that are commonly referred to as "fuller's earth". The unit is lenticular and massively interbedded with cross-bedded sands. The Irwinton Sand Member represents the regressive phase of Jackson Group deposition. It consists of complexly channeled, cross-bedded sand with tough clay partings, and finely interlaminated sand and clay. Lenses of opal-cemented sand and opal-replaced shell hash occur in some sections. In the up-dip direction coarse clastics of the Barnwell Formation become coarser and kaolinite replaces montmorillonite as the dominant clay mineral. The undifferentiated formation is indistinguishable from the Middendorf in extreme up-dip sections, and probably is fluviatile in origin.

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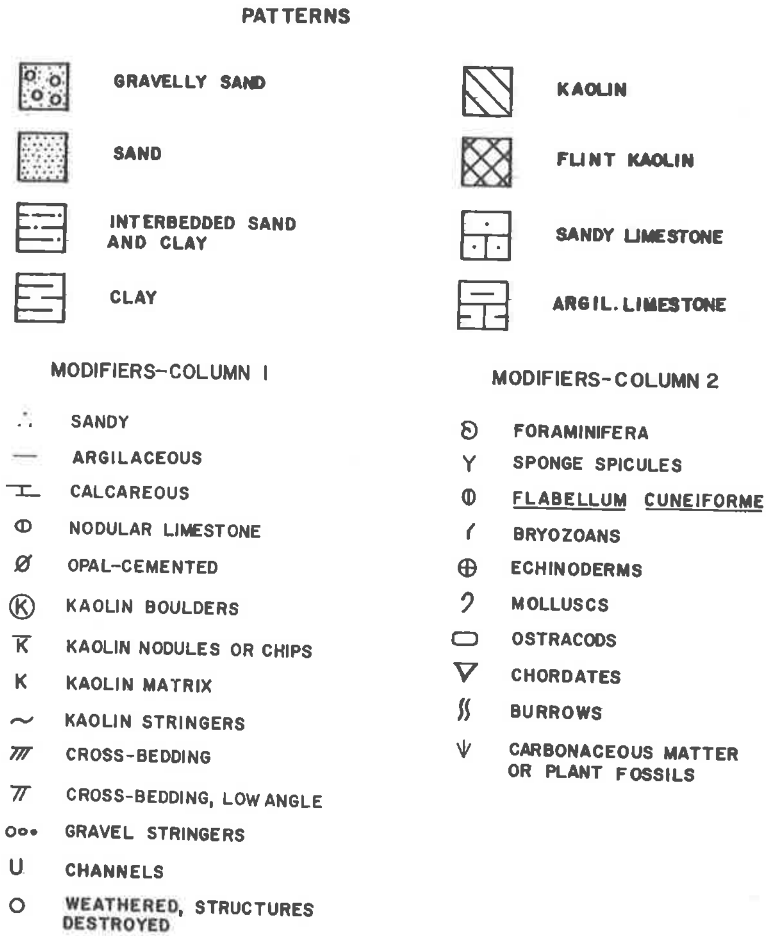
APPENDIX

Graphic Sections and Notes on Sections

Section 1. Quarry of the Penn-Dixie Cement Company on the west side of U. S. Highway 341 about 0.25 miles south of Clinchfield, Houston County, Georgia. Section measured in a vertical cut trending northeast-southwest.

Numbers in the first column to the right of the section are percentage of acid insoluble residue in grab samples from various units. Numbers immediately left of the column, in this and all other sections, are stratigraphic unit numbers from the original field notes. Clinchfield Sd., units 1 through 3; Ocala Ls., unit 4 grading upward into, and interfingering with, the Twiggs Cl. Member. Chert boulders from the Flint River Formation occur on the first drainage divide to the south.

Section 2. Section exposed in road cuts along Georgia State Highway 247 between Big Indian Creek and the top of the first drainage divide to the south, southeastern Houston County, Georgia. This section was originally described by Carver (1966, p. 90-91) as including 90 feet of Twiggs Clay Member with two thin (2 and 4 feet) limestone beds near the base. The Georgia Department of Mines, Mining and Geology drilled

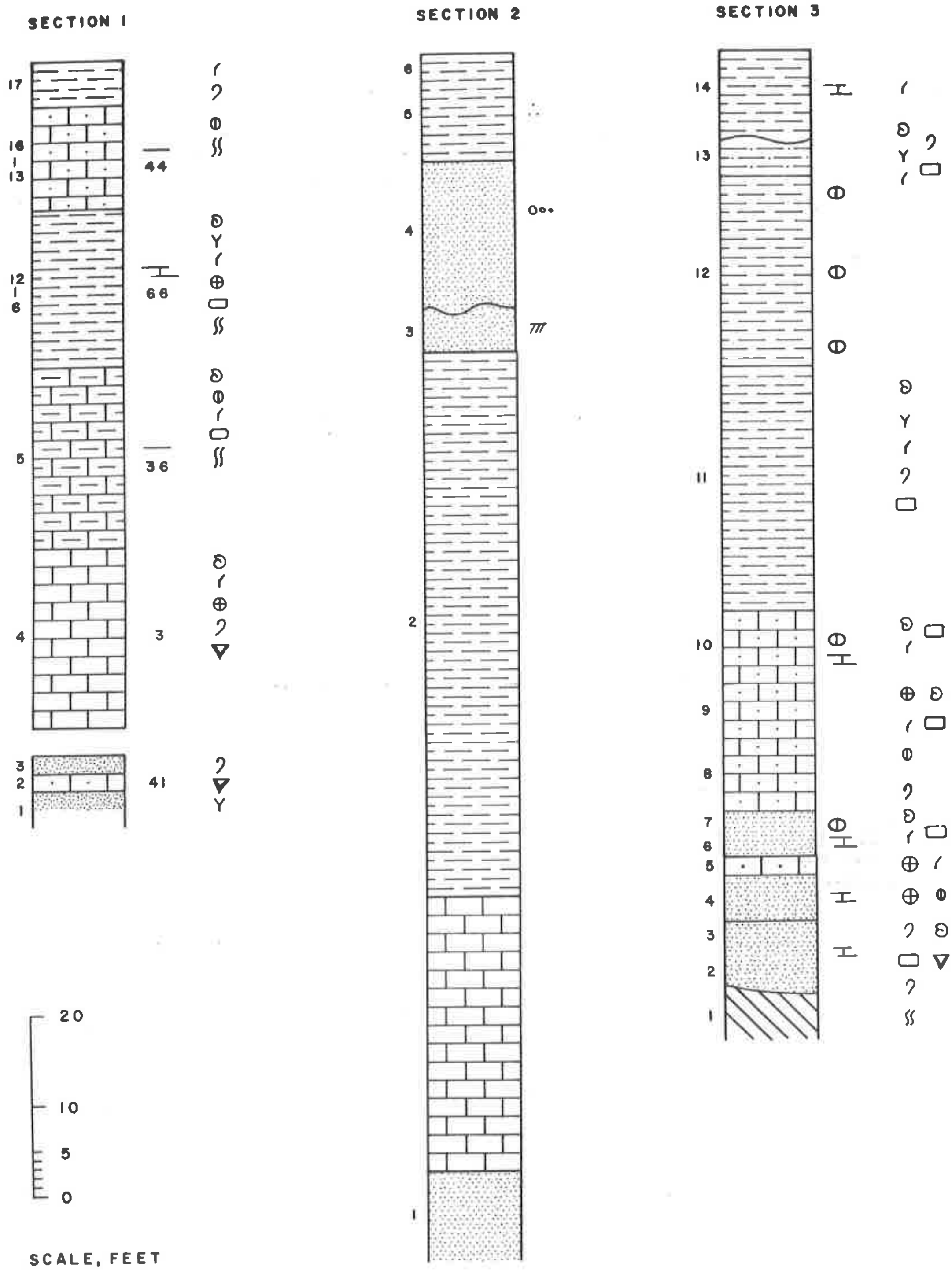


this site and found 30 feet of limestone at the base of the unit described by Carver as Twiggs Clay (S. M. Pickering, oral communication, 1970). Since 1966, sedimentary structures in the upper sand units have been completely obscured by iron oxide cementation.

Clinchfield Sd., unit 1; Ocala Ls., limestone at base of unit 2; Twiggs Cl. Mem., upper unit 2; Irwinton Sd. Mem., units 3 and 4; Upper Sd. Mem. of LaMoreaux (1946), unit 4; Flint River Fm., unit 5.

Section 3. J. M. Huber Corp. Pit No. 22, approximately 4 miles ENE of Huber, Twiggs County, Georgia. Teeth of elasmobranchs are very abundant in parts of the spoil bank bordering the pit. The teeth appear to be from unit 2 or 3.

Middendorf Formation, unit 1, abundant and varied borings in the upper part of the unit; Clinchfield Sd., grading irregularly upward into Ocala Ls., units 2 through 7; sandy nodular limestones of the Ocala


SECTION 1
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SECTION 2
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SECTION 3
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SCALE, FEET

Ls. grading upward into Twiggs Cl., units 8 through 10; calcareous Twiggs Cl. with hard limy sub-units and nodular lenses in upper part, units 11 through 14; unit 13 is interlaminated sand and clay of the Irwinton Sd. Mm. interfingering with the Twiggs Cl.

Section 4. Section in road cuts at Mattie Wells School on Georgia State Highway 49, NE of Macon, 2.2 mile NE of the Bibb County-Jones County line. Unit 3 is composed of opal-replaced shell hash and opal-cemented sandstone (Carver, 1969).

Twiggs Cl. Mem. grading upward into Irwinton Sd. Mem., units 1 through 3; Irwinton Sd. Mem., interbedded-sand-and-clay facies, individual beds 1 to 18 inches thick, units 4 and 5.

Section 5. General Refractory Co. Stevens Pottery pit, approximately 5.5 miles NE of Gordon, Wilkinson County, Georgia. This section is sufficiently far up dip that the boundary between the Middendorf Formation and Jackson Group is not clearly defined.

Middendorf Fm., units 1 and 2A; Barnwell Formation undifferentiated, unit 2B; Twiggs Cl. Mem., unit 3; Irwinton Sd. Mem., units 4 and 5.

Section 6. Old Atlantic Refractory Co. pit on the Georgia Railroad 4.0 miles S. W. of Devereaux, Hancock Co., Georgia.

Middendorf Formation, unit 1; Barnwell Formation undifferentiated (or Middendorf Formation), unit 2.

Section 7. Abandoned kaolin pit on Brush Creek aporoximately 2 miles northwest of Deepstep (33°2.8'N, 82°55'W), Washington County, Georgia. Two Zones of limestone nodules, with nodules up to 5 feet long and 1 foot thick, in unit 1 probably represent tongues of the Sandersville Ls., Unit 5 contains very abundant horizontal, sand-filled burrows.

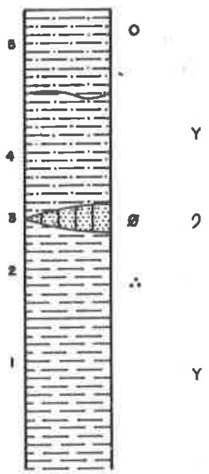
Twiggs Cl. Mem., unit 1; Irwinton Sd. Mem., units 2 through 6.

Section 8. Thiel Kaolin Co. Hall Mine, between Buffalo Creek and Brush Creek on the north side of Georgia Road S689 approximately 1 mile east of Deepstep, Washington County, Georgia.

Middendorf Fm., units 1 through 3, Clinchfield Sd., unit 4; Twiggs Cl. Mem., unit 5, Irwinton Sd. Mem., units 6 through 8.

Section 9. Minerals and Chemical Co. Phillip kaolin pit, north of U. S. Highway 24 and west of Bluff Creek approximately 15 miles due west of Sandersville, Washington County, Georgia. Section 7 is 8.75 miles NE of Section 9, and Section 8 is 6 miles ENE. These three closely spaced sections illustrate the lenticular nature of lithologic units in the Middendorf Formation and Jackson Group. The sections are similar in overall distribution of lithologies, but no single unit can be traced from one section to any other.

SECTION 4



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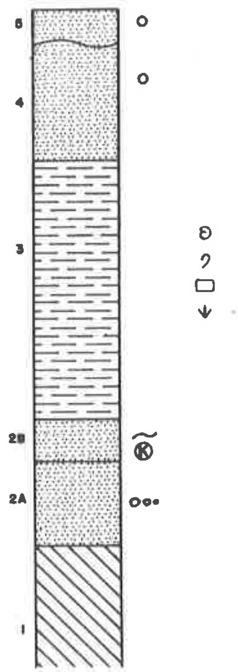

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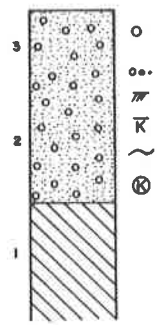

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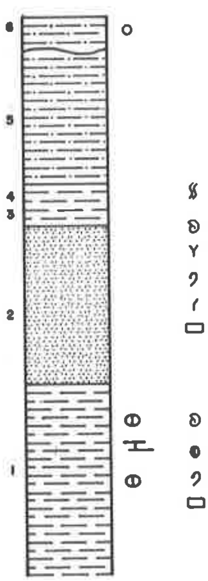

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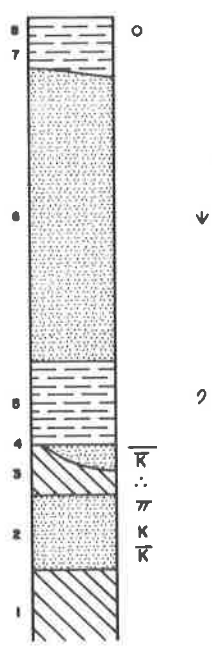

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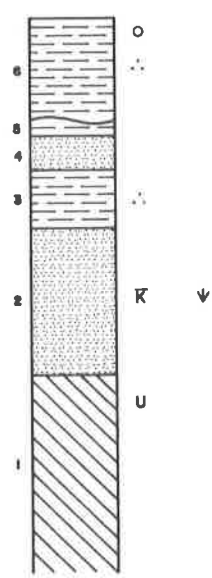


Middendorf Fm., unit 1; Clinchfield Sd., unit 2; Twiggs Cl. Mem., units 3 through 6.

Section 10. The left column is a section measured along Georgia Road S1454, 2 miles ESE of Gibson, Glascock County, Georgia. The right column is the section at the abandoned Harbison-Walker flint kaolin pit 2.5 miles SW, as described by Sandy, Carver and Crawford (1966, p. 23-28). These two sections illustrate, again, the extreme variability of the clastic facies of the Jackson Group. While the sequence in the right-hand column can be subdivided as Glascock Mem., Albion Mem., Twiggs Cl. Mem., Irwinton Sd. Mem.; the left-hand column can be subdivided only as Middendorf Fm. (unit 1) and Irwinton Sd. Mem. (units 2 through 4), or Middendorf Fm. and Barnwell Fm. undifferentiated.

Section 11. J. M. Huber Corp. kaolin pit on the east side of Georgia Highway 17 approximately 4 miles north of the Jefferson County-Warren County line. The channel contact between the kaolin (unit 1A) and sand (1B) of the Middendorf Formation is sharp and suggests that the kaolin was deposited in an abandoned channel. The Jackson Group sediments are dominantly finely interlaminated sand and clay, complexly channeled. These probably are fluviatile sediments, assignable to the Barnwell Formation undifferentiated.

Section 12. Babcock and Wilcox Corp. Albion Kaolin Mine, 1.5 miles WNW of Hepzibah, Richmond County, Georgia. This is the type section of the Albion Member of the Barnwell Formation. All of the Barnwell Formation units are lenticular and vary greatly in thickness within the mined area. The section given here is a composite of three sections measured in the northwest corner of the pit between August 1965 and November 1966. The sands numbered 9 and 10 are lenses within unit 8, or vice versa, and the whole complex has been assigned to the Twiggs Clay Member. Units 4 through 7 occupy low areas along the Middendorf-Jackson Group contact and are absent where the contact is highest.

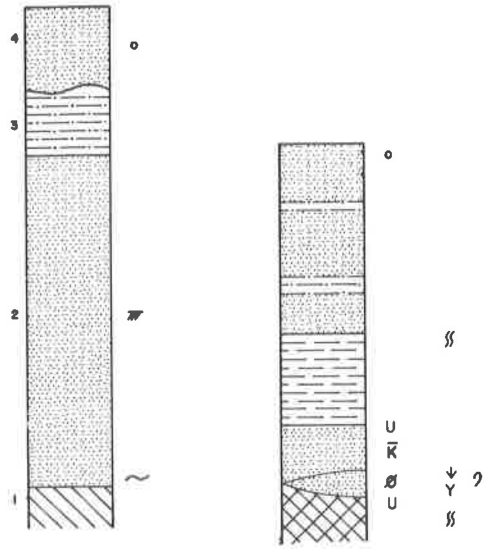
Middendorf Fm., units 1 through 3 (Glascock Mem., unit 3); Albion Mem., units 4 through 7; Twiggs Cl. Mem., units 9 through 10; Irwinton Sd. Mem., unit 11.

Section 13. Abandoned sand pit on west side of Augusta Concrete Block Co. at the Savannah River Valley wall 0.5 miles NE of the Georgia-South Carolina boundary on U. S. Highway 1/78; Aiken Co., South Carolina.

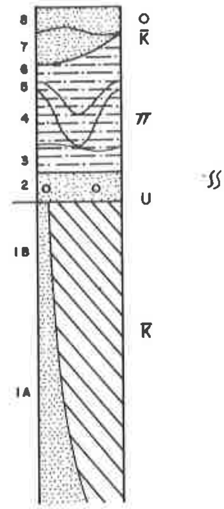
Middendorf Fm., units 1 and 2; Barnwell Fm. Undiff., units 3 through 10.

Section 14. Abandoned sand pit on the north side of U. S. Highway 1/78 on the east side of Graniteville, Aiken County, South Carolina. Massive apparently structureless sands of the Barnwell Formation undifferentiated (units 2 through 4) overly the Middendorf Formation (unit 1).

SECTION 10



SECTION 11



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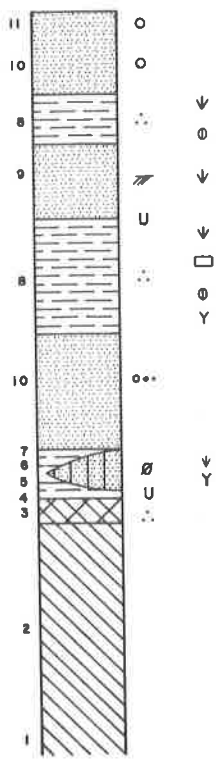

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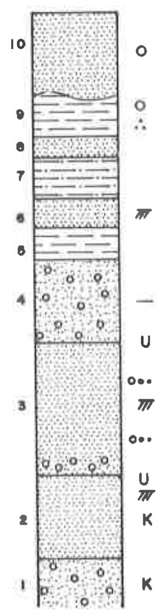
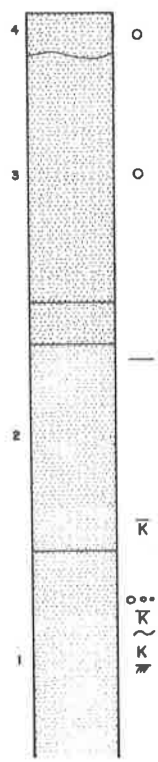


SECTION 14



A PETROFABRIC STUDY OF THE DARK RIDGE AND BALSAM GAP DUNITES, JACKSON COUNTY, NORTH CAROLINA

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ABSTRACT

These dunites, typical of the Southern Appalachian province, contain the mineral assemblage olivine-talc-chromite. Both have been partially serpentinized near the peripheries. Neither exhibits evidence of shear zones or foliation that can be related to the country rock. Optically there is no apparent dimension fabric or fabric to the extinction scheme. Only a few grains exhibit undulatory extinction or twinning; virtually none show kink bands or deformation lamellae. Optical petrofabric analysis revealed no preferred orientation of olivine a, b, or c axes from either body. The lack of both deformation features and preferred orientation seems incompatible with either the solid dunite intrusion or crystal mush intrusion hypotheses. The random orientation of olivine suggests crystallization or recrystallization in a low-strain environment, an environment which could be explained if the olivine-talc assemblage had been derived by the thermal dehydration of serpentinite.

INTRODUCTION

Statement of the Problem

Alpine-type periodotites generally exhibit tectonite fabrics, an indication that they have been subjected to severe deformations. The fabric of the olivine has been used in these cases to help deduce the conditions under which the peridotite was formed and emplaced, and the conditions under which the entire peridotite-country rock complex was deformed. It is the purpose of this paper to report the results of a fabric study of two alpine-type dunites of the Southern Appalachians. The investigation was carried out in an effort to ascertain first, whether a preferred orientation existed; and second, whether the fabric could be

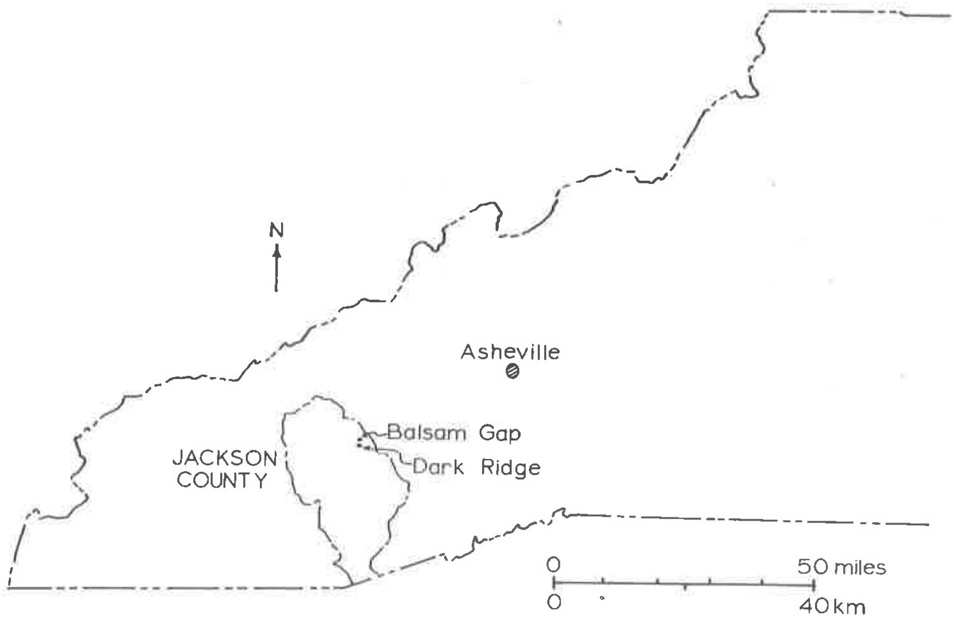

Figure 1. Index map of western North Carolina, showing location of Dark Ridge and Balsam Gap dunities.

related to the origin and mode of emplacement of these dunites and the post emplacement history of the entire dunite-country rock complex. The Balsam Gap and Dark Ridge deposits of Jackson County, North Carolina were selected for this study (Figure 1).

Acknowledgements

We greatly appreciate the comments and criticisms of several people who have read various drafts of this manuscript, especially Donald T. Secor of the University of South Carolina, C. B. Raleigh of the U. S. Geological Survey (Menlo Park), Hans G. Ave' Lallemant of Rice University and George W. DeVore of Florida State University. This does not imply acceptance of all of the interpretations presented here by any of the above. We further acknowledge financial support from several sources, N. S. F. grant GA-4518 (to J. R. C.), field expense support from the North Carolina Department of Conservation and Development, Stephen G. Conrad, State Geologist (to P. M. A.) and the University of South Carolina.

PETROLOGY AND STRUCTURE

Country Rocks

Only two major lithologic units were noted within the general area; a mafic gneissic country rock, part of the "Precambrian layered gneiss and migmatite" of Hadley and Nelson (1971) and three dunite deposits. The mineralogic composition of the country rock is highly variable on both a regional and local scale. The country rocks range in composition from an intermediate composition (plagioclase 40%, quartz 30%, biotite 20%, garnet 10%) to a very mafic composition (blue-green hornblende 65%, plagioclase 35%); plagioclase compositions range from An_{30} to An_{45}. No pyroxene was noted in any of the thin sections. These brackets, according to Turner (1968) place the gneiss in the amphibolite facies, indicating that it has been subjected to an overburden pressure of about 6 kb and a maximum temperature of 550°C.

The country rocks surrounding the Balsam Gap and Dark Ridge dunites, as in the case of many other dunites of the Southern Appalachians, exhibit a thin rind of alteration immediately adjacent to the dunites. The typical mineral assemblage in these alteration zones is talc-biotite-vermiculite. According to Hunter (1941) this alteration post-dates intrusion of the ultramafics and is probably related to the later intrusion of the alaskite and pegmatites so common in this area.

In contrast to the highly variable composition of the gneiss, its local structure is remarkably uniform. The strike of the foliation, measured in 30 locations, ranged from N35E to N70E and averaged to N45E. The dip, though slightly less regular, averaged to about 70° NW, (Figure 2). The only exceptions to these measurements were found in the area immediately adjacent to the Dark Ridge deposit. There the strike of the foliation appeared to be wrapped around the dunite body. Whether this warping can be related to the origin and emplacement history of the dunites is problematical.

Dunite Bodies

Macroscopic features. Both dunite deposits are roughly pod-shaped with an overall length of about 1600 feet and a width of approximately 900 feet. At Balsam Gap the rock is composed predominately of light green friable olivine and contains relatively little serpentine, talc, and spinels. The rock at Dark Ridge is darker green, fairly hard, and appears to contain a greater percentage of serpentine and spinels than Balsam Gap. The olivine is very coarse grained in places.

Neither body exhibits any evidence of structural features that can be related to the structure of the country rock. The bodies do exhibit a weak foliation defined by an occasional planar distribution of disseminated chromite, and there are folded seams of chromian spinel. There appears to be no relationship between the foliation of the dunite

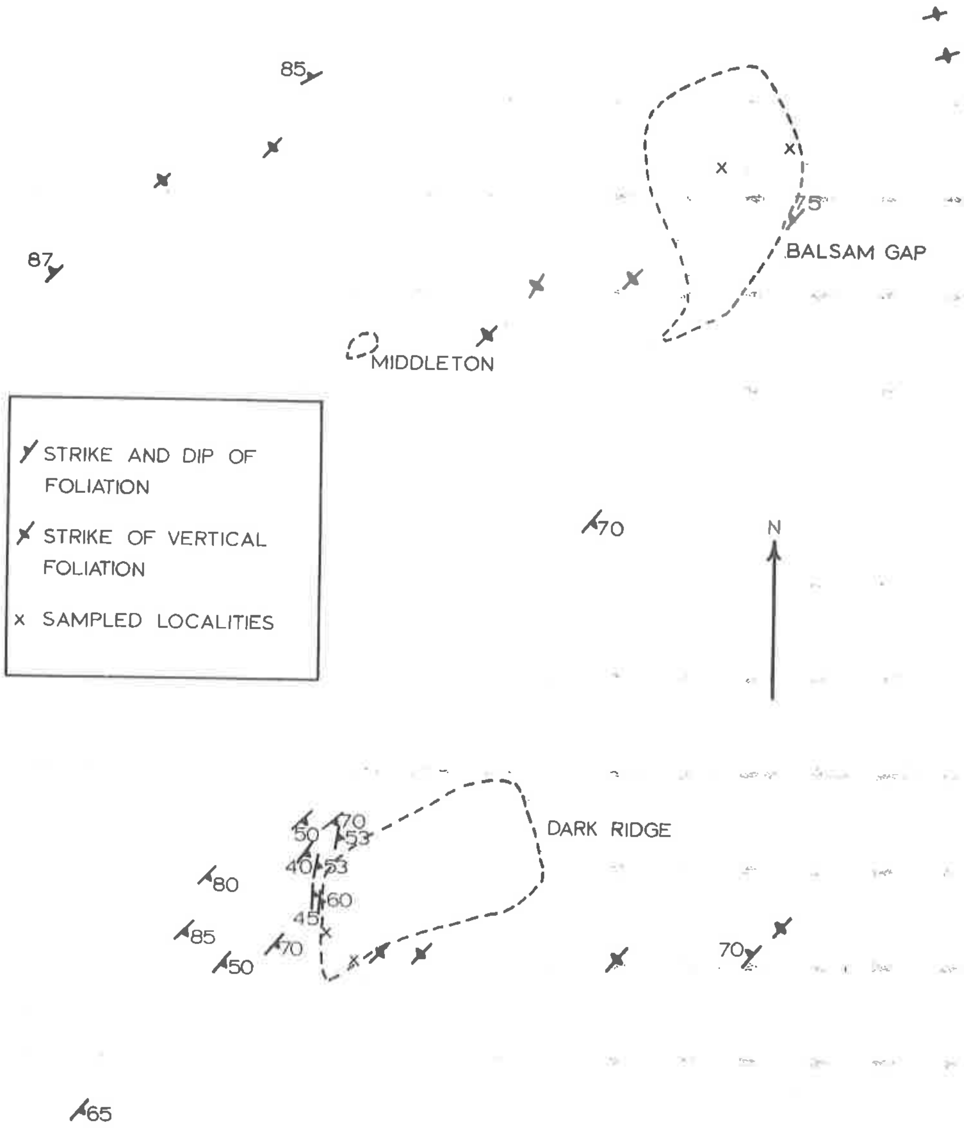

Figure 2. Generalized geologic map of area surrounding Dark Ridge and Balsam Gap dunities.

and the foliation of the country rock. Nor does there seem to be any relationship between the deformation of these chromite seams and the deformation of the country rock.

Petrography. Both bodies can be classified as dunites or serpentinized dunites. The olivine content decreases outward from about

90 percent in the relatively unaltered center of each body. The olivine composition, as determined by 2V, x-ray data, and partial spectrochemical data, ranges between $Fo_{92.0}$ and $Fo_{92.5}$. There is no evidence, optical or chemical, for compositional zonation. Serpentine content varies from less than 5 percent in the centers to a maximum of 75 percent at the periphery of the Dark Ridge deposits. The majority of this variation may be accounted for by post-emplacement hydrothermal alteration. Such peripheral serpentinization has been noted by Hunter (1941) in most of the alpine-type dunite-peridotite bodies of this province. Talc is a ubiquitous component making up about 10 percent of the rock and does not vary significantly within a deposit or between deposits. A chromian spinel is the only other mineral phase making up more than 1 percent of the total rock.

Texture. Olivine grains in each deposit are anhedral with diameters ranging from 0.1 mm to 2 cm but generally in the range 0.5 mm to 1.0 mm. Unlike other alpine-type peridotites such as the Cypress Island, Washington (Raleigh, 1963, 1965) and the Lherz, Pyrenees (Ave' Lallement, 1967), there appears to be no dimensional fabric to either of the rock bodies studied; virtually all olivine grains are equant and virtually none show any signs of shearing and there is no fabric to the extinction. Furthermore, no deformation lamellae or kink bands were noted. In general, grains were untwinned, although occasional grains were twinned on (100).

PETROFABRIC ANALYSIS

Method

Two oriented samples were taken from each of the dunites for petrofabric analysis. At Balsam Gap one sample was collected near the center of the body and one near the edge. At Dark Ridge the overburden cover made it necessary to take both samples near the perimeter. Three orthogonal thin-sections were cut from each sample. The thin sections were examined with a Leitz five axis universal stage using the method described by Emmons (1943). On the average 40 crystals were examined in each slide; the positions of the a, b, and c crystallographic axes were plotted on standard Schmidt nets with a diameter of 20 cm. Each of these plots was then "rotated" to the horizontal plane. The final step was to combine all of the data concerning each crystallographic axis for each dunite body onto one stereographic net. These final six projections were then contoured at levels of 1, 2, 4 and 8 percent.

Results

Contoured petrofabric diagrams of olivine a, b, and c crystallographic axes from the Dark Ridge and Balsam Gap deposits (Figures 3

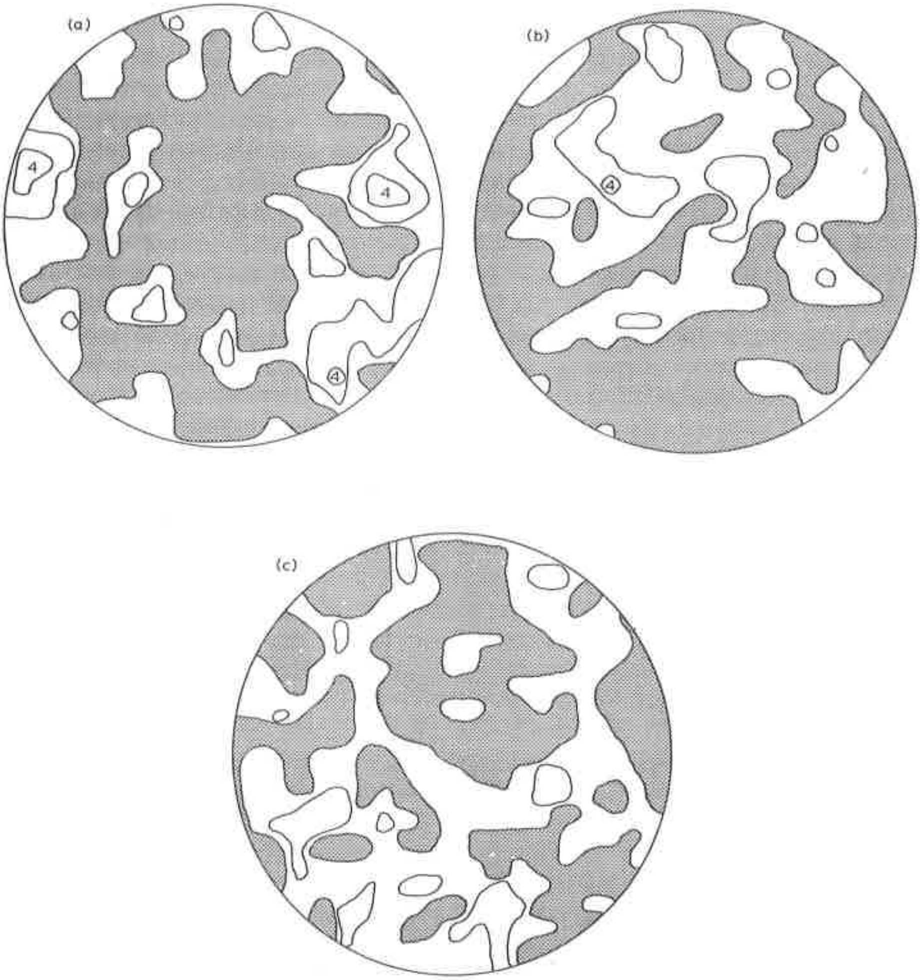

Figure 3. Petrofabric diagrams for Dark Ridge. Contours at 1%, 2%, 4%, 8%. a. olivine a axes, b. olivine b axes, c. olivine c axes.

and 4) show that point maxima concentrations on each are very low. In addition there is little evidence for girdle maxima.

For sake of comparison, a diagram based on 100 computer-selected points of pseudo-random azimuth and inclination has been prepared (Figure 5). Note that the point maximum concentration on the random diagram is higher than the point maximum concentrations on four of the six plots from the ultramafic rocks. It seems doubtful that diagrams so closely resembling plots of random data can indicate any significant preferred orientation.

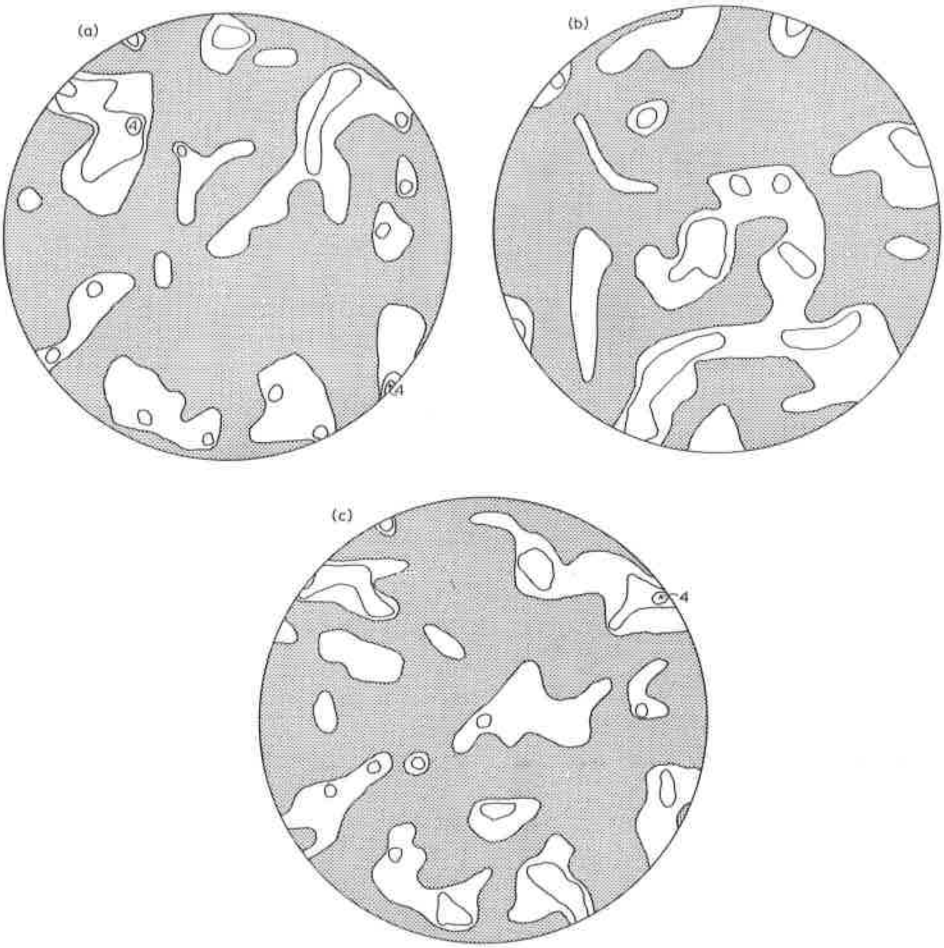

Figure 4. Petrofabric diagrams for Balsam Gap. Contours at 1%, 2%, 4%, 8%. a. olivine a axes, b. olivine b axes, c. olivine c axes.

Evaluation of the Significance of the Results

An approach to testing the extent of preferred orientation or degree of disorder for petrofabric diagrams has been worked out by one of us (W. E. Sharp).

The maximum degree of disorder of the diagram was defined as:

$$S_{max} = - \log G_T$$

where G_T is the total number of grid points in the diagram. Next the degree of disorder (S_T) of the observed diagram was calculated using the formula:

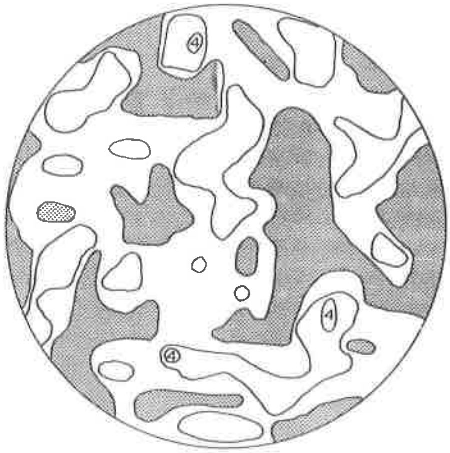

Figure 5. Petrofabric diagram for 100 random azimuths and inclinations. Contours at 1%, 2%, 4%, 8%.

$$S_T = \frac{1}{N_T} \sum_{i=1}^{k} N_i \log i - \log N_T$$

$$N_i = G_i \cdot i$$

$$N_T = \sum_{i=1}^{k} N_i$$

$$G_T = \sum_{i=1}^{k} G_i$$

where G_i represents the number of grid points on the stereo diagram which had a point count of i in a 1% area around the grid point. The deviation between the maximum degree of disorder possible and the calculated degree of disorder is then taken to be

$$D = ((S_{max} - S_T)/ S_{max}) \times 100.0$$

The values of D ranged from 0 for a perfectly uniform distribution up to a maximum of 100 when all points on the diagram are clustered on a single grid point (assuming a 1 cm square counting grid is used on a 20 cm equal area net).

In this instance, this formula is being used as a means by which the existing plots may be compared with that of a fictional plot which is completely uniform; that is, a plot on which 100 points have been so distributed that there will always be one point in any 1 percent area that may be selected. The degree of this fictional plot is represented by the figure 'S_{max}'. The degree of disorder of each real diagram, 'S_T', was then computed by the formula which has been given. The numbers which appear in Table 1 represent the percent to which 'S_T' deviates from 'S_{max}';

Table 1. Percent Deviation (D) of S_t From Uniformity.

Balsam Gap	Dark Ridge	Random Plots	Basalt Dike
b axes 4.5%	6.5%	8.9%	24.5%
c axes 8.5%	4.3%	10.1%	
a axes 7.6%	10.4%		

in effect they may be read as the percent to which the given diagram is more oriented than the fictional plot. Similar calculations were made using a stereonet diagram prepared by Brothers (1959) representing the orientation of olivine grains in a basalt dike.

Again no evidence of preferred orientation could be found and some of the plots based on petrofabric data appear to have a less preferred orientation than the random plots. At the same time it should be noted that both types of diagram differ significantly from Brothers' diagram. This information serves to strengthen the earlier contention in this paper that there is no sign of any significant orientation in these particular alpine-type ultramafic bodies.

Comparison with Other Alpine-type Dunite Peridotites

These results appear to be in direct contradiction to the results of other workers who have investigated olivine fabrics in other alpine-type peridotite-dunites. Ave' Lallemant (1967) concluded that olivine grains in an alpine-type lherzolite from the Pyrenees exhibited a preferred orientation with olivine a point maximum parallel to the fold axis of the country rocks and olivine b maxima spread into a partial or complete girdle perpendicular to the axial plane cleavage. He concluded that the fabric in this rock had been generated during metamorphism as a result of syntectonic recrystallization. It should be noted, however, that the maximum concentration was exhibited by olivine b and was only slightly over 6 percent per 1 percent area. The olivine a and c axes exhibited an even lower concentration. We would question whether 6 percent truly represents "preferred" orientation or perhaps is an expected value in a random distribution. Ave' Lallemant's (1967) point is strengthened, however, by the fact that some 18 samples yielded very similar petrofabric diagrams.

Raleigh (1963, 1965), in a study of the Cypress Island (Washington) peridotite, found that deformed mineral layering was the result of crystal settling and that the deformation of these layers was generated during solid emplacement. He further found that olivine grains exhibited a preferred orientation that could be related to deformation of the layering, olivine a maxima were parallel to fold axes of the deformed layers, and thus concluded that the olivine fabric was tectonic in origin, imprinted during the tectonic emplacement of the body. He also found evidence that suggested that the mechanism by which the olivine came to have a preferred orientation was recrystallization.

Turner (1942) and Battey (1960) found that the Dun Mountain (New Zealand) peridotite exhibited a strong preferred orientation of olivine, again with olivine a maxima parallel to the fold axes of the country rocks. Paulitsch (1953) found a similar fabric in a dunite from Greece. One other olivine tectonite fabric that has been described has olivine b point maxima (with partial girdles) perpendicular to the schistosity of the country rock (Andreatta, 1934; Ernst, 1935; Yoshino, 1961, 1964).

DISCUSSION

In contrast to the results reported for other alpine-type ultramafics, the olivine in the two bodies studied exhibited no preferred crystallographic orientation. This conflict may be resolved, in part, by the application of more rigorous statistical tests for randomness. We believe that such tests would reveal that the published data for some alpine-type peridotites are not statistically significant, and thus that the conclusions drawn concerning preferred orientation in those bodies are not justified. However, we do accept that there are other alpine-type ultramafics in which the olivine crystals do display a significant degree of preferred orientation. The lack of any such orientation in the Southern Appalachian dunites studied here suggests that these bodies differ from these other alpine-type ultramafic deposits in their origin, mode of emplacement or post-emplacement history.

The deformational fabric in the adjacent country rock suggests the existance of a medium-to-strong non-hydrostatic strain environment during deformation. The lack of any such fabric in the ultramafic bodies suggests that they crystallized under conditions of hydrostatic or very low non-hydrostatic strain. To reconcile this disparity we must establish the method by which the bodies were emplaced and the temporal relationship between that emplacement and the regional episodes of deformation.

If the ultramafics were emplaced after the regional deformation one would expect to find evidence of either tectonic or magmatic activity. It has been established by a number of studies that olivine that has been emplaced tectonically shows evidence of such in the form of foliation, deformation structures or textural fabric. None of these was observed. Magmatic emplacement should have created some sort of contact metamorphic aureole, but the only alteration of the wall rock that was noted around the Appalachian ultramafics is considered to be a low temperature phenomenon (less than 1000^{o}C).

If post-deformation emplacement seems thus unlikely, one must assume that the deposits were put in place before or during the regional deformation and were thus subjected to the same deformational episodes as the country rock. The total lack of fabric in the dunites may be explained in two ways. In one case it could be said that an earlier fabric was erased by a post-deformational thermal recrystallization, but the country rock shows no evidence of temperatures high enough ($\approx 1000^{o}C$) to achieve this. In the other case it could be suggested that some peculiar set of circumstances existed such that the ultramafics found themselves in a low-strain environment during deformation. Such would be the case if the ultramafics were emplaced early as serpentine and were then thermally dehydrated to an olivine-talc-(relict) chromite assemblage (Carpenter and Phyfer, 1969). This could have occured during a medium-to-high strain deformational event if the serpentine deformed continuously and the olivine grew as if subjected to an hydraulic medium. Once the equant olivine grains had formed, the rind of talc and serpentine on each grain would have permitted additional strain in the rock with little deformation of the olivine, providing all of the yield could be confined to the intergranular talc and serpentine. Additional pressure relief may have been provided by the volume decrease indicated by the serpentine dehydration model. One might also expect some local adjustment in the country rock. The fact that no direct evidence of such adjustment may be seen is not unreasonable since the country rock at about $500^{o}C$, the temperature of serpentine dehydration, would certainly be behaving as plastic (incompetent) bodies allowing most evidence of adjustment to be destroyed.

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THE LITTLE MOUNTAIN SYNCLINE IN THE SOUTH CAROLINA PIEDMONT

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ABSTRACT

Siliceous sediment not more than 250 feet thick, with subordinate aluminous material, was deposited as an original body about 4 miles long and 1.5 miles wide. Folding occurred with an axial direction of N 70° E, and metamorphism converted the siliceous sediment to kyanitic quartzite, clearly bedded and obscurely cross-bedded. The principal fold is a syncline exposed on Little Mountain. The kyanitic quartzite in the limbs closes around the northeast nose of the fold, but pinches out southwestward. This rock also occurs on the crest of Cannon Hill in a subordinate syncline, largely removed by erosion, which apparently plunges southwest at an angle of about 10 degrees.

The limited regional distribution of the kyanitic quartzite may be attributed to local hot-spring activity, and its silica and alumina to sinter and kaolin as hot-spring products. The associated sericite schists, requiring alkalies, were undoubtedly derived from regionally available felsic tuff with a minor addition of clay.

INTRODUCTION

Thirty miles northwest of Columbia, in central South Carolina, occur two low ridges owing their elevation to the erosional resistance of kyanitic quartzite. The larger of the two ridges is Little Mountain, near a town of the same name. It rises 150 feet above the Piedmont upland surface. The smaller ridge is Cannon Hill, rising 110 feet. The kyanitic quartzite was the subject of an excellent petrographic description by McKenzie and McCauley (1968). An earlier and shorter discussion of this locality was given by Espenshade and Potter (1960).

The Little Mountain area is in the western edge of the Carolina slate belt in a structural environment of folds overturned toward the southeast. From Little Mountain eastward to Columbia, Secor and Wagener (1968) divided the slate belt rocks into three formations originally consisting mostly of waterlaid tuffs and mudstones later converted to metamorphic rocks of greenschist facies. These rocks were found to grade upward stratigraphically into rocks of amphibolite facies in the Charlotte belt to the northwest, under conditions as discussed by Wagener in the Winnsboro quadrangle (1970). This interpretation differs from the earlier view of McCauley (1961a, b), who regarded the original rocks of the Charlotte belt as being older than those of the slate belt, which is essentially the relation shown on the map of Overstreet and Bell (1965).

The present study involves only the geologic structure of Little Mountain and Cannon Hill, and the possible origin of their rocks. McKenzie and McCauley (1968) showed the structure of these ridges as homoclinal, with the beds dipping southeast, and their base map is used in the preparation of Figure 1 herein.

LITHOLOGIC UNITS

The Little Mountain rocks may be grouped in three lithologic units, but stratigraphic correlations cannot be made with present knowledge. The units appear to be members of the undifferentiated Persimmon Fork and Wildhorse Branch Formations of Secor and Wagener (1968).

The most conspicuous unit is the kyanitic quartzite that has attracted both geologic and economic interest. The rock is dense and hard, consisting of quartz intercrystallized with kyanite, white mica, and pyrite in uneven proportions. McKenzie and McCauley (1968) state that the mica includes muscovite, paragonite, and pyrophyllite; and they also list rutile, ilmenite, hematite, lazulite, corundum, cassiterite, and garnet. Bedding planes are present, and are most sharply defined where the relatively coarse kyanite crystals are least abundant. The quartzite crops out prominently, and McCauley (1961b) has noted in drill cores that it is interbedded with muscovite and chlorite schists. The sequence has a maximum thickness of 250 feet, but close measurement cannot be made as the lower slopes have a forested soil cover.

Beneath the kyanitic quartzite occurs a fine-grained, quartzose muscovite schist containing widely uneven amounts of pyrite. McKenzie and McCauley (1968) also found paragonite, hematite, ilmenite, magnetite, and discrete layers of chlorite. Outcrops are too few and widely separated for the mapping of structural details, but in some of them the schist has a cleavage dipping steeply northwest and probably correlative with the axial-plane cleavage of Secor and Wagener (1968). Schistosity dips southeast at angles greater than 45 degrees.

Schist with apparently similar characteristics, containing thin kyanitic beds, occurs above the kyanitic quartzite. Only the lower part of this unit remains uneroded in the synclinal trough on the crest of Little Mountain, and outcrops are rare.

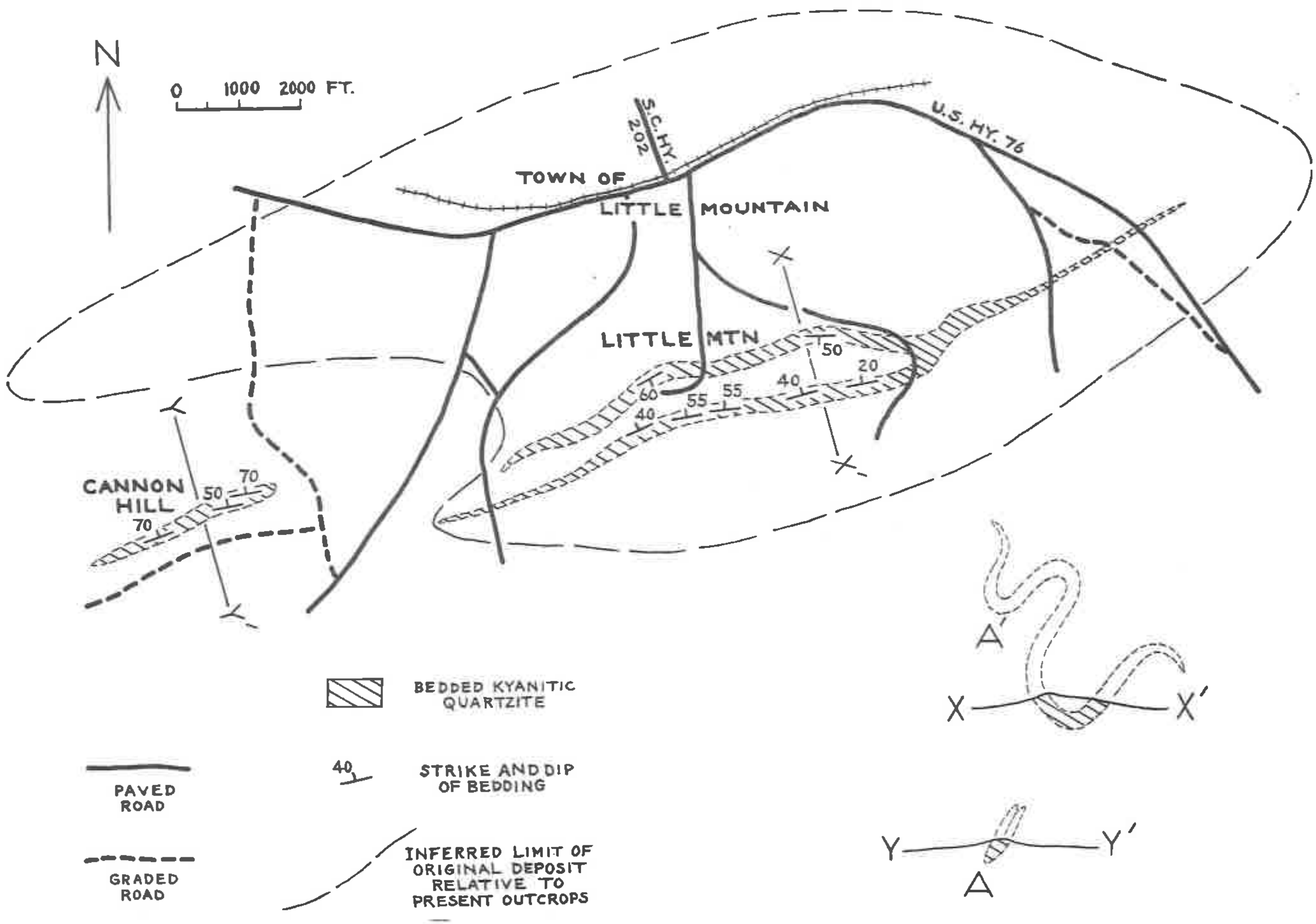

Figure 1. Geologic structure of the Little Mountain area, Newberry County, South Carolina.

FOLDS

On Little Mountain, two outcrop belts of the kyanitic quartzite converge at the eastern nose of a synclinal fold (Figure 1), a structural situation unusually convenient for open pit mining if the grade of the rock should be attractive. Outcrops are best on the south slope where bedding planes dip 20^{o}-55^{o} NW. Cross bedding in upright position is present in the outcrop showing a dip of 20^{o}. Outcrops are scarcer on the north slope where the dip is 50^{o}-60^{o} SE. The configuration of the limbs indicates an essentially horizontal axis, but the long, tapering nose suggests an eastward transition to the northeast plunge of folds recorded by Secor and Wagener (1968). The pinching out westward of the quartzite in the limbs marks the limit of deposition of its original sediment.

On Cannon Hill, the kyanitic quartzite occupies the entire crest and dips only to the northwest at 50 to 70 degrees. The outcrops suggest the compressed keel of a syncline as shown in section Y-Y' in Figure 1. The gap between Cannon Hill and Little Mountain is a shallow, headwater valley. Adequate outcrops show that it is underlain only by schist. Thus, the quartzite of the two ridges cannot be erosional remnants of a single fold, and the axis of the Cannon Hill fold must be parallel to and north of the Little Mountain axis.

The unique nature of the quartzite among the rocks of this region, however, suggests that all of it was originally a single deposit; and there could have been a connection providing that it had sufficient elevation for the erosional removal of its very resistant debris before the present time. This possibility would require a southwest plunge of the Cannon Hill fold. No evidence for an angle of plunge was found, but, if the quartzite in Cannon Hill (A in section Y-Y') has an axial inclination of only 10 degrees, it would have passed upward to position A' in section X-X', with a vertical interval of 400 feet above the position of the present stream in the gap. This would have been more than adequate for the necessary erosion, and would result in a pattern of folds overturned to the southeast as are others in this region.

COMPARATIVE FEATURES OF THE KYANITIC QUARTZITE

The distribution of the kyanitic quartzite is limited to that shown in Figure 1. To the writer's knowledge, no other large deposits of kyanitic quartz rock have been reported between York County, South Carolina and Lincoln County, Georgia, a distance of 100 miles.

Although bedding is clearly preserved, the Little Mountain rock is crystalline and properly a metamorphic quartzite in the sense that it originally consisted mainly of sedimentary silica. This implies a genetic similarity to the kyanitic quartzites of the Farmville area in Virginia and of the Kings Mountain area in the Carolinas. These rocks have been well described and mapped by Espenshade and Potter (1960), and their sedimentary origin is not in dispute. Although in those areas quartzites occur in bodies isolated by the erosion of complicated structures, the wide distribution of these bodies is believed to reflect deposition of the original sediment over broad areas. For instance, the quartzites near Kings Mountain occur in an area 40 miles long and 5 miles wide, and those near Farmville in an area 30 miles long and 18 miles wide.

In contrast, the Little Mountain quartzite occurs in an area less than 4 miles long and 0.25 mile wide. Even with allowance for its eroded parts as inferred from the cross sections, the width of the deposit before folding could hardly have been more than 1.5 miles. Any assumption that its original extent was vast, and that erosion has left only a small part of it would imply that this part marks the lowest structural level in the entire region. Such an implication cannot be supported by any of the adjacent regional studies, and hence the scale of the original deposit in relation to the present outcrops, as reconstructed in Figure 1, is believed to be reasonably correct. Even if it were twice as great, it would still be incompatible with the usual great extent of rocks formed primarily from sediment as common as quartz sand.

ORIGIN

Table 1 shows on line 1 the main components of the kyanitic parts of the Little Mountain quartzite, as computed from the estimated averages of 62 percent quartz, 19 percent kyanite, and 16 percent mica given by McKenzie and McCauley (1968). They list the kyanite as ranging from 2.50 to 34.75 percent. The same authors estimated 47 percent each of quartz and mica in the schist associated with the quartzite, and these averages have been similarly computed for line 2. As the mica is reported to include muscovite and paragonite, K_2O and Na_2O are not distinguished.

It has been assumed that these rocks are metamorphosed slate-belt sediments, mostly of pyroclastic origin. Secor and Wagener (1968) have shown that such materials were originally available from an adjacent volcanic environment containing felsic tuff and pumice. Their composition probably resembled that of 11 New Mexico rhyolitic volcanic rocks, which have been averaged on line 3 of Table 1. It is evident that rhyolitic materials are too high in alkalies and rather low in alumina to have formed either of the Little Mountain rocks.

In considering origin, it is important to remember that the kyanitic quartzite is of limited regional occurrence whereas the mica schist is not similarly limited. The schist could have had its origin in felsic material such as that of line 3, with admixture of a high-alumina clay. For simplification, the clay is assumed to have been pure kaolinite, and its composition on a water free basis is given on line 4. A mixture of

Table 1. Principal Components of Little Mountain Rocks and Their Possible Source Materials.

	SiO_2	Al_2O_3	K_2O+ Na_2O	
1. Kyanitic quartzite	76	18	2	Computed from average mineral compositions given by McKenzie and McCauley (1968).
2. Quartzose sericite schist	68	18	6	
3. Felsic volcanics	74	13	9	Average of 11 analyses of Tertiary rocks from New Mexico (Wells, 1937)
4. Kaolinite (water free)	46	54	-	Theoretical
5. 80% of line 3 and 20% of line 4	68	21	7	Possible components for quartzose sericite schist, line 2
6. Siliceous sinter (water free)	95	3	1	Average of 15 analyses of sinters from Yellowstone (Clarke, 1915)
7. 65% of line 6 and 35% of line 4	78	21	1	Possible components for the kyanitic quartzite, line 1

eighty percent of the felsic material and twenty percent of the water-free kaolinite would furnish the components shown on line 5, which approximate those of the schist, line 2. With adjustments for water and the normal impurity of kaolinitic sediment, the similarity might be even closer.

Possible source materials for the Little Mountain quartzite have less latitude because the distribution of the quartzite apparently reflects a local condition. Even if special conditions of alteration could bring the felsic material of line 3 to the composition of line 1, it would account only for the kyanitic type of the quartzite and not for the type essentially free of kyanite and mica. The latter type is the key to the origin of this entire unit.

A feature of volcanism that is characteristically local even in widespread volcanic provinces is hot-spring activity, which is known to form, in quantity, two materials with chemical components essential to those of the Little Mountain rocks. One of these is siliceous sinter, whose water-free composition is similar to that of quartz sand. The average of 15 water-free analyses of Yellowstone sinters (Clarke, 1915) is given on line 6 of Table 1. The other material is kaolin, long known as a hot-spring alteration product, and recently described in clear genetic relations to felsic volcanics in Mexico by Keller and Hanson (1968) and Kesler (1970).

Erosion of the siliceous sinter alone (line 6) could have yielded the silica sediment of those parts of the Little Mountain quartzite having little kyanite or mica. Combined with 35 percent of water-free kaolinite, as on line 7, it could have provided the main constituents for the average kyanitic quartzite (line 1). The average of 16 percent mica estimated for the kyanitic quartzite is almost certainly too high, and consequently the alkalies on lines 1 and 7 should be closer than shown. Thus, locally occurring hot springs in a known environment of felsic volcanic rocks could have provided not only the source materials of the Little Mountain quartzite, but also its element of distinction among all of the rocks of its region.

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