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STRUCTURE OF THE PULASKI-SALEM THRUST SHEET AND THE EASTERN END OF THE CHRISTIANSBURG WINDOW, SOUTHWESTERN VIRGINIA

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ABSTRACT

The map area comprises about 35 square miles in Montgomery County, Virginia. The bedrock ranges in age from Cambrian to Devonian and belongs to five fault blocks. From north to south they are Catawba, Salem, Pulaski, Saltville and Max Meadows.

The northwesternmost and structurally lowest block is the parautochthonous Saltville block, which is exposed in the Christiansburg window of the Pulaski overthrust sheet. Exposed are the upper Elbrook Formation, the Saltville block Cambro-Ordovician Knox Group of carbonates and Middle Ordovician limestones. Strata within the window represent the essentially recumbent north limb of the Christiansburg anticlinorium. The Pulaski-Salem blocks contain folded, highly-fractured, and brecciated Elbrook and Knox Formations. Structural complications of the Salem block include the high-angle Willow Springs, Cambria, North Cambria and Kettle Ridge faults. The Salem thrust is interpreted as a relatively minor, yet important, splay off the Mississippian-age Pulaski thrust. The Pulaski overthrust has a minimum of nine miles of northwestward essentially horizontal displacement. Rocks of the Catawba block along the northern margin of the area range from Middle Cambrian to Mississippian in age. The Max Meadows block is the southeasternmost and highest structural block in the area. Cambrian Rome Formation and possibly some lower Knox Copper Ridge Formation comprise this thrust sheet, as well as allochthonous strata to the northwest. Klippen of the Max Meadows thrust sheet indicate a minimum northwestward horizontal movement of 15,000 feet for this thrust sheet.

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INTRODUCTION

The area mapped comprises approximately 35 square miles of the Appalachian miogeosyncline, which extends from the Blue Ridge to the Appalachian Plateau. Within the 35-mile wide zone are numerous folds and associated major longitudinal faults, which trend northeast. From the southeast to northwest the major folds are: the Blue Ridge anticlinorium, the Riner synclinorium, the Christiansburg anticlinorium, the Salem synclinorium, the Price Mountain anticline, the Blacksburg syncline of Campbell and Holden (1925), the Poplar Hill anticline, the Spruce Run syncline, the Clover Hollow anticline, the Johns Creek syncline, the Bane Dome, the Butt Mountain syncline and the Hurricane Ridge syncline. In the area mapped the Christiansburg anticlinorium separates the Riner synclinorium to the southeast from the Blacksburg synclinorium (Cooper, 1961) to the northwest. The structure of the mapped and adjacent areas is imbricate, with parautochthonous Saltville block rocks forming the structurally lowest block.

From the Blue Ridge Upland on the southeast to the Appalachian Plateau on the northwest, the faults are: the Friest thrust, the Blue Ridge overthrust, the Max Meadows overthurst, the Salem thrust, the Pulaski overthrust, the high-angle Saltville fault, the Narrows fault and the Saint Clair fault (Figure 1). The mapped area, which is in the overthrust belt, extends from the Max Meadows fault northwest to the Salem fault, which is an imbrication of the Pulaski thrust sheet. The Pulaski thrust sheet includes seven major fensters: Kent, Christiansburg, Ingles-Barringer Mountain, East Radford, Price Mountain, Blacksburg and Read-Coyner Mountain (Figure 1).

Within the mapped area (Figure 2) is the eastern extremity of the Christiansburg window in the Pulaski thrust sheet, with Saltville block carbonates exposed. The Max Meadows thrust sheet overlaps the Pulaski block along the southern margin of the mapped area, with several Max Meadows klippen occuring in the eastern half of the area (Figure 2). The southwestern margin of the Salem synclinorium, delineated by the Salem fault trace, forms the northern boundary of the map area. This Salem fault is a relatively minor, yet important, splay off the Pulaski thrust. About two miles south of the southeastern corner of the area is the Blue Ridge thrust sheet (Figure 1), which overlaps the Max Meadows sheet.

Acknowledgments

The writer wishes to thank W. D. Lowry of the Department of Geology, Virginia Polytechnic Institute and State University for his critical review and constructive criticism of the manuscript. The writer also wishes to thank the Department of Geological Sciences for the use of their research facilities.

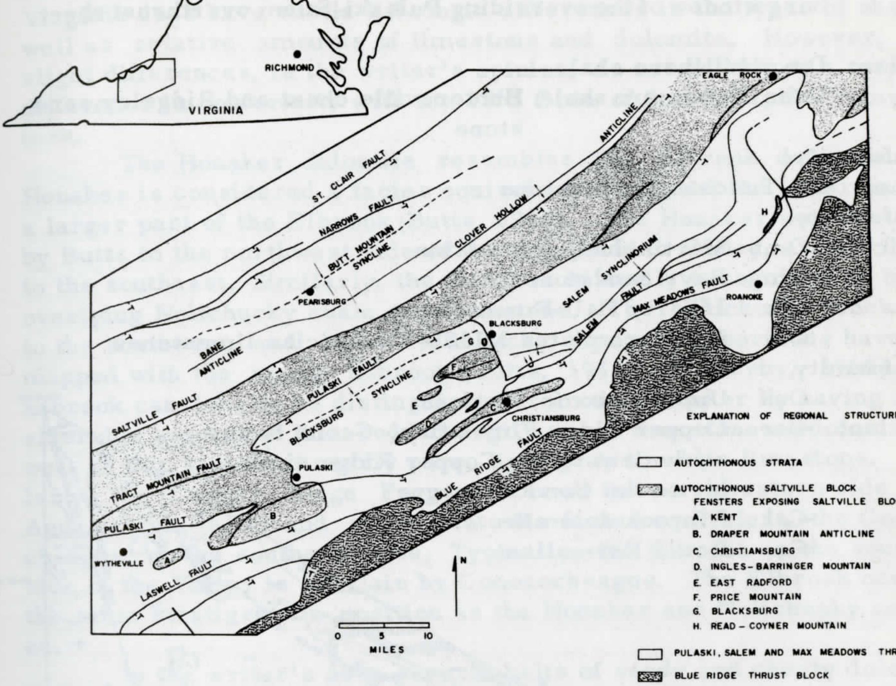


Figure 1. The major regional structures of the Appalachian Valley in southwestern Virginia.

STRATIGRAPHIC FACIES CONSIDERATIONS

Stose (1906) first designated the Elbrook Formation as a middle member of his Shenandoah Group in the western portion of South Mountain, Pennsylvania, and the adjacent part of the Cumberland Valley. He noted the Elbrook Formation was above the Waynesboro Formation and below the Knox Group, and assigned the formation a Middle-Upper Cambrian age. He described the unit as a massive, bluish-gray, magnesian limestone, with numerous thin layers and nodules of chert and beds of shale. He noted that red and green shales are present in the middle of the formation and that beds of sandy limestone occur higher in the section. A thickness of 2, 000 feet is cited. The name Waynesboro is still used north of Roanoke, Virginia, whereas Rome terminology is preferred to the south.

The Elbrook Formation in southwestern Virginia and northeastern Tennessee is somewhat different lithologically from the type Elbrook. The use of the name, therefore, may be inappropriate in this area. Historically (Butts, 1940) applied the name Elbrook to the

Figure 2. Deatiled geology of the study area showing the southern margin of the Catawba block, the leading edge of the Max Meadows thrust sheet and the exposed Saltville block in the Christiansburg window of the overriding Pulaski-Salem overthrust sheet.

Devonian: Dm Millboro shale
Drhn Needmore shale, Huntersville chert and Ridgeley sandstone

disconformity

Silurian: St. Tuscarora sandstone

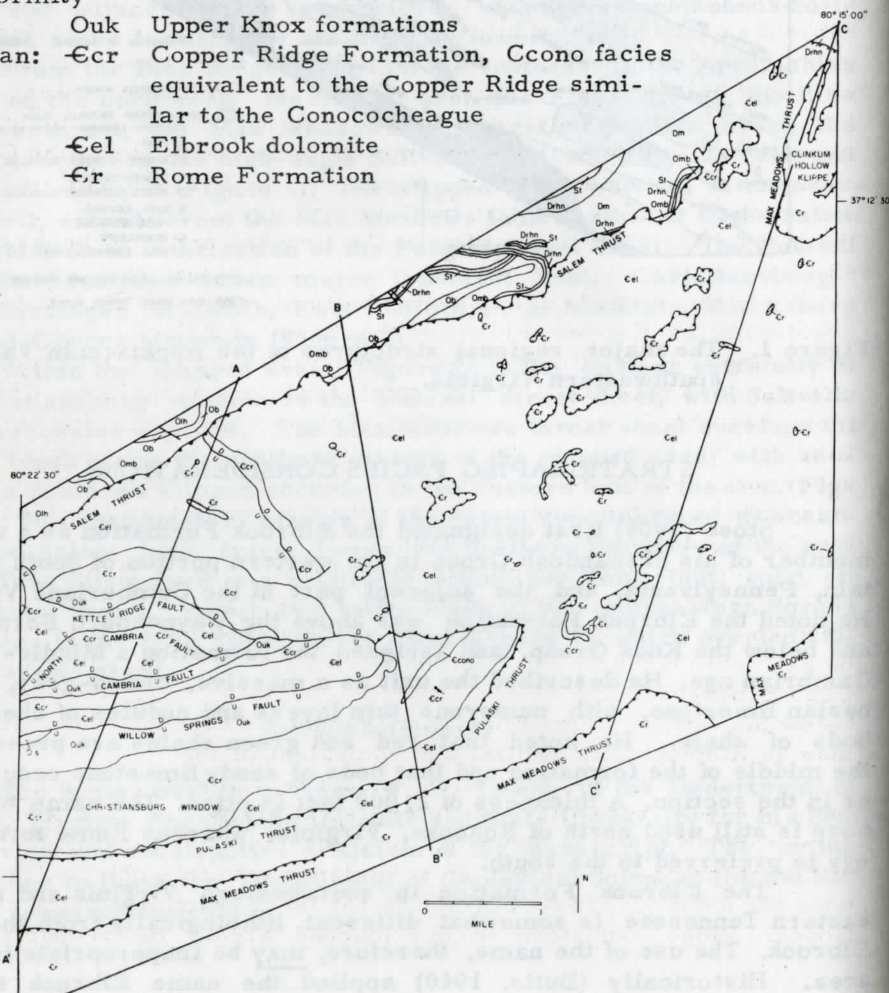
disconformity

Ordovician: Omb Martinsburg Formation
Ob Bays sandstone
Olh Liberty Hall Formation
Ols undifferentiated Middle Ordovician limestones

disconformity

Ouk Upper Knox formations

Cambrian: Ꞓcr Copper Ridge Formation, Ccono facies equivalent to the Copper Ridge similar to the Conococheague
Ꞓel Elbrook dolomite
Ꞓr Rome Formation



dolomite sequence southeast of the Pulaski thrust fault that occupies the same stratigraphic position as the Honaker and Nolichucky Formations to the northwest. The writer admits that the Elbrook of southwestern Virginia does have subtle lithologic differences in the types of shale, as well as relative amounts of limestone and dolomite. However, these slight differences, in the writer's opinion, do not constitute a mappable difference in conformity with the 1961 Code of Stratigraphic Nomenclature.

The Honaker dolomite resembles the Elbrook dolomite. The Honaker is considered a facies equivalent of at least the lower and also a larger part of the Elbrook (Butts, 1940). The Honaker was restricted by Butts to the northwest side of the Appalachian Valley, and the Elbrook to the southeast. Similarly, the Honaker is easily recognized by having overlying Nolichucky shale and limestone. The Nolichucky pinches out to the southeast, and some Nolichucky equivalent horizons have been mapped with the upper Elbrook (Butts, 1940 and Derby, 1966). The Elbrook can locally be distinguished from the Honaker by having a considerable amount of thin-bedded argillaceous limestone or dolomite, as well as thin layers of nearly pure, light gray to white limestone. Similarly, the Copper Ridge Formation is on the northwestern side of the Appalachian Valley and grades into its facies equivalent, the Conococheague, on the southeast side. Typically, the Elbrook on the southeast side of the Valley is overlain by Conococheague. The Elbrook occupies the same stratigraphic position as the Honaker and Nolichucky combined.

In the writer's area several belts of sandy and cherty dolomites are defined as Copper Ridge on the basis of a thick basal sandstone. According to B. N. Cooper (oral communication, 1971) the Conococheague does not have a similar thick sandstone at the base. The writer feels that the Elbrook Formation of the area, considering even slight lithologic variations, is the same unit as those dolomites conformably under the Copper Ridge Formation. The writer cannot distinguish on a mappable basis the dolomites under the Copper Ridge from those under the Conococheague of Butts (1940).

Cooper (oral communication, 1971), Ritter (1969) and Glass (1970) consider the Elbrook dolomite, as well as the Honaker, to be defined on the basis of what units are above and below. Therefore, based on <u>external</u> evidence, the Elbrook Formation can occur only below the Conococheague Formation. According to this view, those dolomites below the Copper Ridge Formation cannot be Elbrook even though they may be lithologically similar and equivalent in age.

Possibly, the formation should be delineated on <u>internal</u> lithologic grounds, regardless of future stratigraphic considerations. The same formation (i. e., Elbrook) can therefore be conformably below different facies equivalents. In accordance with this opinion, the writer uses the name Elbrook Formation for those dolomites below the basal Copper Ridge sandstone in his mapped area. He recognizes the possibility that they may not be Elbrook in the strictest sense, but they cannot be separated from such on a mappable basis.

CHRISTIANSBURG ANTICLINORIUM OF THE PARAUTHOCHTHONOUS SALTVILLE BLOCK

The Saltville block (Figure 1) forms the lowest structural unit within the area. Essentially autochthonous rock of this block is exposed largely in the Christiansburg window (Figure 2). The window is about 10 miles long and extends westward from the map area to west of Christiansburg, Virginia. The overall length along strike of the window complex (Figure 1) is 17 miles between Christiansburg and Radford, Virginia. The Christiansburg window is separated on the south from the Max Meadows thrust sheet by a narrow strip of Elbrook dolomite of the Pulaski block. The Christiansburg window extends 4.3 miles eastward from the western boundary of the map area to its eastern limit near Montgomery Station. The width of the window within the area is approximately one mile.

Prior to Pulaski thrusting, a major northeast-trending synclinorium, referred to as the Montgomery basin by C. E. Sears (oral communication, 1971), extended from the Sinking Creek anticlinorium on the northwest to the upwelling Blue Ridge anticlinorium on the southeast. The Christiansburg anticlinorium divided the Montgomery basin into the Blacksburg synclinorium (Cooper, 1961) on the northwest and the Riner synclinorium on the southeast. Whether this Christiansburg anticlinorium existed prior to Pulaski thrusting is not clear, but it certainly appears that a substantial part of its development occurred after thrusting. Minor faulting, such as the Willow Springs Fault, which cuts the Saltville and Pulaski blocks, is attributed to the continued rise of this anticlinorium.

The Saltville block exposed in the map area belongs to the nappe-like overturned northwestern limb of the Christiansburg anticlinorium. West of the area the anticline is less asymmetric and both limbs are exposed (Glass, 1970). Along the western margin of the map area, the overturned northwest limb dips less than 40 degrees south, and on Den Hill, four miles eastward, the limb is recumbent with dips as low as 0 to 10 degrees south.

The Christiansburg anticlinorium is exposed in the Christiansburg, Ingles-Barringer and East Radford windows. Of these, only the Christiansburg window is present in the writer's area. These windows are due largely to the Christiansburg anticlinorium, which extends from east of Christiansburg, with easterly plung to more than 17 miles to the west, where it plunges to the west near Radford, Virginia.

Strata of the Saltville block exposed within the map area are, from south to north, a thin margin of Elbrook Formation, the Knox carbonates of Late Cambrian and Early Ordovician age, and undifferentiated Middle Ordovician limestones. The distribution of sandstones of the Copper Ridge Formation largely defines the overall structure of this block.

The sandstone at the base of the Copper Ridge Formation near the southern boundary of the window is overturned and dolomite of the underlying Elbrook is cut off by the Pulaski thrust.

The northern margin of the writer's study area lies within the Catawba syncline, which contains formations ranging in age from Cambrian to Mississippian. Ritter (1969) and Cooper (1961) think that this syncline is a parautochthonous part of the Saltville block. This matter is discussed later.

PULASKI THRUST BLOCK

Pulaski Thrust

The Pulaski thrust fault was named by Campbell and Holden (1925) after the town of Pulaski, Virginia. It was thought by them to extend from Green County, Tennessee, to the vicinity of Eagle Rock on the James River, Botetourt County, Virginia. Butts (1933, 1940) traced the Pulaski fault northeast from Eagle Rock to Greenville, Augusta County, Virginia. As outlined above, the Pulaski fault is about 200 miles long. Cooper (1946) recognized the Staunton fault between Greenville and Staunton, Virginia, as a northern extension of this major structure. Cooper (1960) noted that the Pulaski fault trace can be recognized in the vicinity of Newport, Tennessee, where it passes beneath the Blue Ridge fault. Consequently, the length of the fault may be more than 350 miles. The Pulaski thrust sheet is one of the major thrust blocks of the Southern Appalachians.

The leading edge of the Pulaski thrust sheet lies along the southeast base of Brush Mountain, seven miles northwest of the study area. Classically, the Pulaski fault is believed to mark the northwest margin of the exposed Catawba syncline. The Catawba syncline would thus be considered to be part of the Pulaski block (Campbell and Holden, 1925). Cooper (1961) believed that the Catawba syncline is part of the Saltville block, and thus the trace of the Pulaski fault would curve southeast through Blacksburg along Ritter's (1969) Yellow Sulphur fault trace (Figure 3). Cooper thought the trace of the Salem fault of Eubank (1967) along the northern margin of the writer's map area might be the Pulaski fault. According to Cooper's interpretation the fault trace curves back to the north, northwest of Roanoke (Figure 4). The Salem synclinorium would thus be considered a major reentrant of the Pulaski thrust. Cooper first postulated that the fault along the northwest margin of the Catawba syncline was an extension of the Tract Mountain fault, though later he (Ritter, 1969) considered it to be the trace of a fault which Ritter named the Catawba fault.

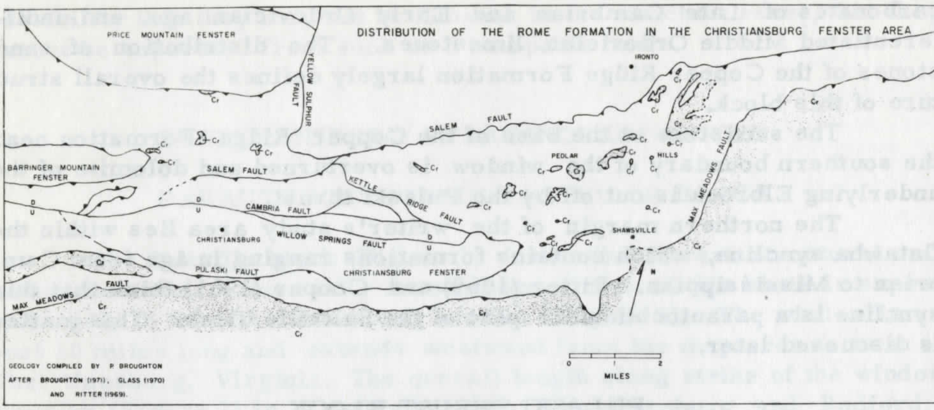


Figure 3. The distribution of the Rome Formation in the Christiansburg window area.

Structure of the Pulaski Thrust Sheet

The essentially horizontal nature of the Pulaski fault is emphasized by the existence of several windows in this part of southwestern Virginia. The Pulaski fault is a low-angle thrust which is not far under the surface within much of the writer's area (Figure 5). The fault comes to the surface in the western half of the area and forms the southern margin of the Christiansburg window. North of the Christiansburg window is the Price Mountain window, and to the west is the Ingles-Barringer Mountain window. The overridden Saltville block is exposed in these windows.

North of the Christiansburg window are several east-trending, high- to low-angle faults. These, the Cambria, North Cambria and Kettle Ridge faults, are discussed later with the Salem block. The Willow Springs fault, which cuts both Saltville block and Pulaski block strata, forms the northern margin of the Christiansburg extends eastward into the map area from a point about 1, 000 feet south of the Norfolk and Western railroad. This fault throws a reefy facies of the Middle Ordovician limestone of the Saltville block against Elbrook strata of the displaced Pulaski block on the northwest. The fault cuts across the Middle Ordovician limestone and terminates that belt about 1, 000 feet east of the western margin of the map area. Farther east, Knox carbonates of the Saltville block abut the Elbrook of the Pulaski block. The Willow Springs fault joins the low-angle Willow Springs fault at the east end of the Christiansburg window at a point about 3, 000 feet south of Montgomery station. The high-angle Willow Springs fault probably resulted from the release of stresses from the rising Christiansburg anticlinorium, under the Pulaski thrust sheet. The rising anticline, as

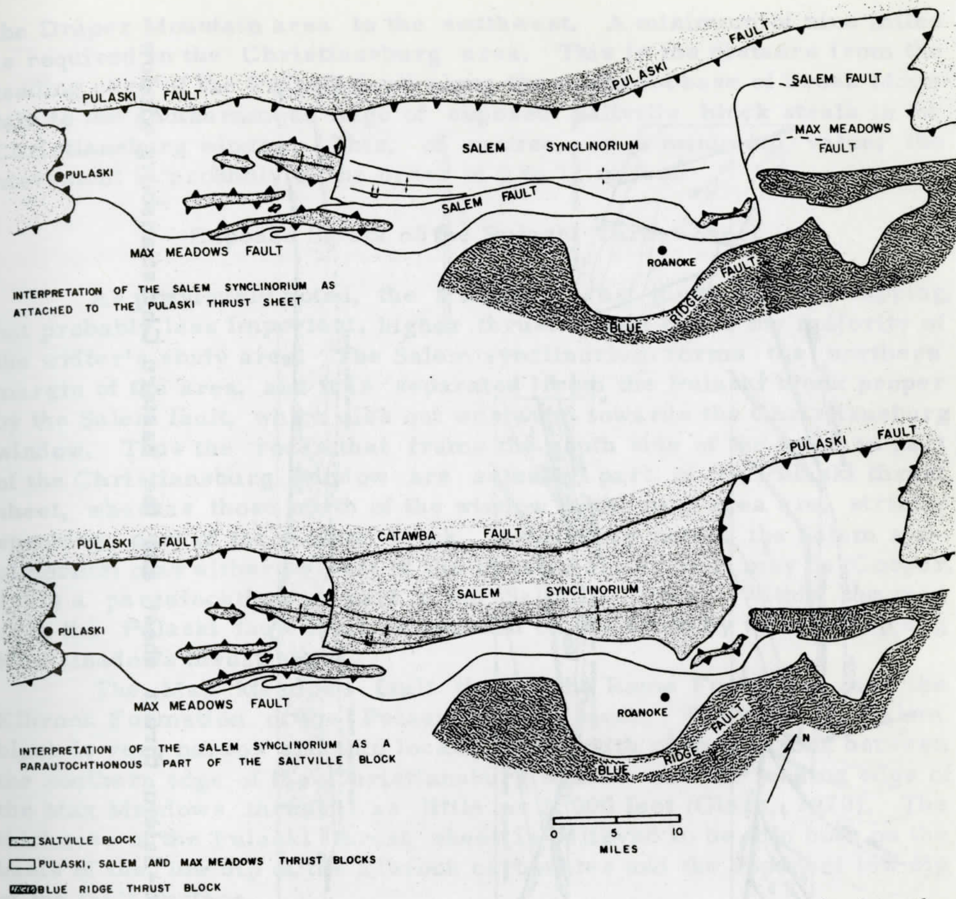


Figure 4. Interpretation of the Salem synclinorium as attached to the Pulaski thrust sheet (top), and the interpretation of the Salem synclinorium as parautochthonous part of the Saltville block (bottom).

well as displaced strata from Willow Springs faulting, promoted window erosion down to the present surface level.

Movement of the Pulaski Thrust Sheet

Estimates of the horizontal displacement of the Pulaski thrust sheet between Pulaski and Buchanan, Virginia, range from 9 to 20 miles. Campbell and Holden (1925) note that the horizontal movement in the vicinity of Pulaski, Virginia, is not less than 9 miles, and probably much more. Butts (1940) believes there may have been as much as 20 miles of northwest movement, whereas Woodward (1932) postulates at least 11 miles. Cooper (1939) notes at least 11 miles of movement in

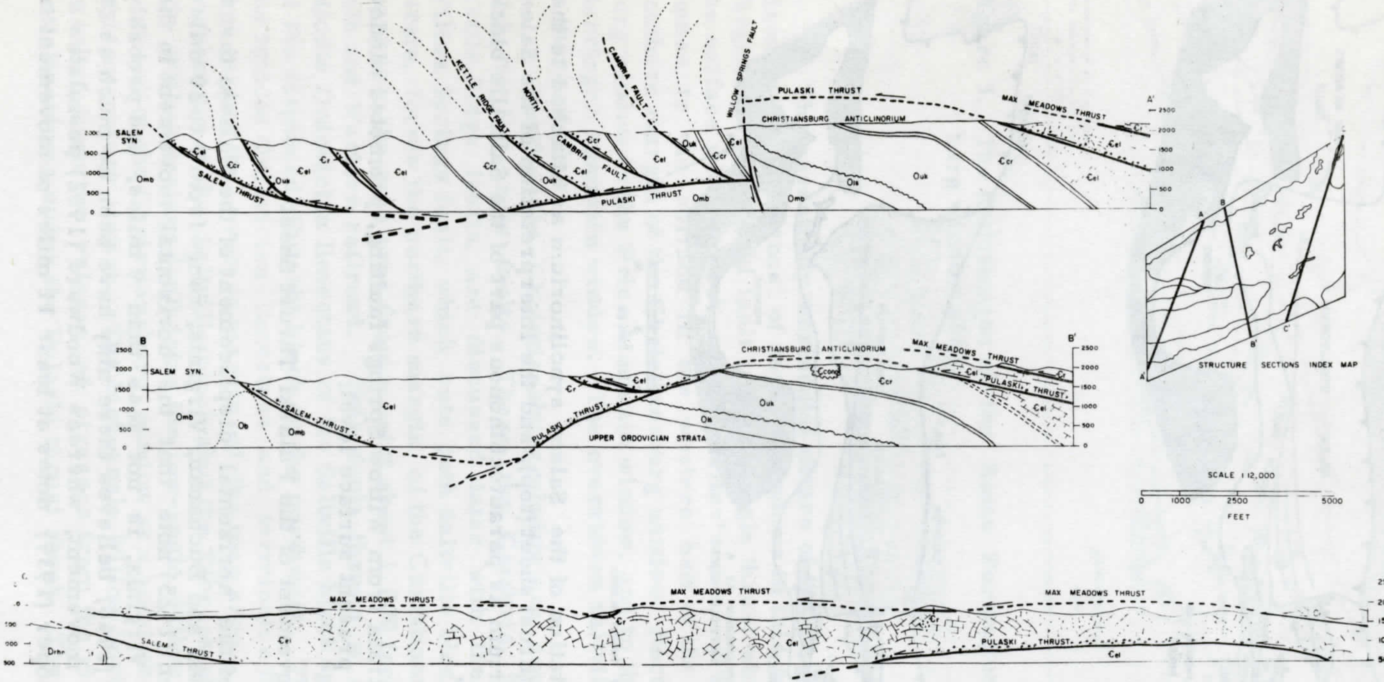


Figure 5. Structure sections of the Pulaski-Salem thrust sheet and the eastern Christiansburg window area of southwestern Virginia.

the Draper Mountain area to the southwest. A minimum of nine miles is required in the Christiansburg area. This is the distance from the leading edge of the Pulaski fault along the southeast base of Brush Mountain to the southernmost edge of exposed Saltville block strata in the Christiansburg window. This, of course, is a minimum value; the movement is probably of the order of 9 to 12 miles.

Exposed Strata of the Pulaski Thrust Sheet

As previously noted, the Pulaski thrust sheet and overlapping, but probably less important, higher thrust sheets cover the majority of the writer's study area. The Salem synclinorium forms the northern margin of the area, and it is separated from the Pulaski block proper by the Salem fault, which dies out westward towards the Christiansburg window. Thus the rocks that frame the south side of the western part of the Christiansburg window are actually part of the Pulaski thrust sheet, whereas those north of the window in the map area are, strictly speaking, part of the Salem block. As already noted, the Salem synclinorium may either be part of the Pulaski block, or it may be (Cooper, 1961) a parautochthonous part of the Saltville block. Within the map area the Pulaski fault block is bounded on the south by the overlapping Max Meadows thrust sheet.

The Max Meadows fault throws the Rome Formation over the Elbrook Formation of the Pulaski-Salem block. The Pulaski-Salem block is very narrow and thin locally. The width of the Elbrook between the southern edge of the Christiansburg window and the leading edge of the Max Meadows thrust is as little as 2,000 feet (Glass, 1970). The thickness of the Pulaski thrust sheet is believed to be thin both on the basis of the low dip of the Elbrook carbonates and the apparent low dip of the fault surface.

The Pulaski block, exclusive of the Salem synclinorium, is dominantly Elbrook dolomite. However, in the map area north of the Christiansburg window, several belts of Knox strata of the Pulaski-Salem block are repeated by faulting. Copper Ridge, Chepultepec, Longview formations have been mapped (Figure 2). Extensive areas of Rome Formation in the eastern half of the area, and a few smaller areas north of the window are interpreted by the writer as klippen of the Max Meadows thrust sheet. However, some of the smaller, lower-elevation Rome outcrop areas could have been clipped off the underlying Rome Formation and carried forward by the advancing Pulaski thrust sheet.

SALEM THRUST SHEET

W. B. Rogers (1884) first recognized the existence of the fault later named the Salem fault by Campbell and Holden (1925). They traced the fault from Salem, Virginia, southwest along the southern margin of the Salem synclinorium to the area of the Ingles-Barringer Mountain window. Butts (1933) extended the Salem fault northeast of Roanoke to near Cloverdale, Botetourt County, Virginia. West of the writer's area, Glass (1970) believes that the Salem fault curves southward away from the southern margin of the Salem synclinorium and dies out at the edge of the Christiansburg window.

The western edge of the southern margin of the Salem synclinorium was named the Ellett Road fault by Glass. Northeast of Roanoke the Salem fault trace was extended to Read Mountain by Woodward (1932) and Hazlett (1968) shows it bounding the Read-Coyner Mountain window on the southeast from which point it swings to the northwest.

In the writer's area, the Salem fault has a dip of about 15 degrees in the northeast corner of the map at Clinkum Hollow. To the west it steepens to 30-40 degrees.

Salem Fault Block

Salem Fault: The Salem block constitutes the hanging wall of the Salem fault; it is a relatively major structure superimposed upon the Pulaski block. The writer's area is essentially Pulaski block, but more correctly it is Pulaski-Salem block. It is postulated that the Salem fault surface merges with the essentially horizontal Pulaski fault surface at a relatively shallow depth (Figures 2 and 5). Thus, the Salem fault is considered to be a low-angle splay of the Pulaski fault. The wide zone of fracturing and brecciation exposed along Den Creek east of the Salem fault appears too great for the displacement along that fault alone and may be better explained as phenomena associated with Pulaski overthrusting which the Salem Fault brought surfaceward.

The southern margin of the Salem block is difficult to determine. Even though Cambrian has been thrust upon Devonian strata along the northern margin of the map area, the throw is relatively insignificant near the latitude of the Christiansburg window. Movement along the Salem fault probably overlaps Pulaski faulting, and the southern margin of the Christiansburg window probably should be termed the Pulaski-Salem fault.

The Salem fault offers a mechanism for removing the post-Knox section south of its leading edge. Strata of the upper Knox to Mississippian are missing from the Salem block. It is postulated by the writer that the Salem fault is relatively near the surface in the eastern half of the area where it has significant stratigraphic throw. However, at the western margin of the block, the fault may lie at greater depth and have less stratigraphic throw. This is supported by the wide shear zones in the eastern half of the area. It provides a reasonable explanation for the presence of both Cambrian and Lower Ordovician (Knox Group) section in the western half and their absence in the eastern half. The eastern half of the area contains only Elbrook because the fault raised the post-Elbrook section above the present ground surface.

Structural Elements of the Salem Block

The Salem fault appears to be relatively steeply dipping but to lose displacement southwestward and thus to die out near the Christiansburg window. Movement along the Salem fault may actually die out just north of the present northern margin of the window (Willow Springs fault). A weak, counter-clockwise rotational movement would produce greater stratigraphic throw and horizontal displacement at a lower angle near the surface at the northeastern corner of the map (W. D. Lowry, personal communication, 1971). The rotational character of the fault may provide an explanation for the relatively short, east-trending Ellett Road fault, and also the Kettle Ridge, Cambria and North Cambria faults, and the shear and breccia zones in the eastern map area.

The Kettle Ridge fault extends from the Salem fault at the northwest corner of the map area to the eastern margin of the Christiansburg window. The fault throws Elbrook dolomite over upper Knox strata. In a few localities the Elbrook forms both the hanging and footwalls of the fault, and its position is interpretive. The fault may die out just west of the northeastern margin of the window but was extended by the writer to terminate against the northern boundary of the window.

Glass (1970) named the Cambria fault after large quantities of tectonic breccia and crush conglomerate that trend northeast from its abutment with the Willow Springs fault west of the map area. The Cambria fault generally follows the Norfolk and Western railroad west of the map area; it enters the writer's map area 1, 000 feet south of the railroad right-of-way (Figure 2). The trace of the Cambria fault in general parallels both the railroad and the Willow Springs fault. Possibly the Cambria fault merges with the Kettle Ridge fault and continues until termination against the northeastern margin of the Christiansburg window, or it may die out.

The breccia locally contains limestone and dolomite boulders as much as two feet in diameter. In places competent blocks of strata are interbedded with the tectonic breccia and crust conglomerate. The fault is generally marked by a breccia zone up to 200 feet wide, rather than a distinct contact. In the map area, the Cambria fault places overturned upper Knox carbonates of the hanging wall on the south against overturned Elbrook dolomite and limestone of the footwall. The fault is a reverse fault that dips southeast. Younger rocks are thrust over older because the Knox carbonates constitute the axial part of a syncline (Figure 5).

Heavy concentrations of tectonic breccia that characterize the Cambria fault west of the map area, extend only 1, 500 feet eastward into the map area. East of this, the fault is extended on stratigraphic considerations. Along the western boundary of the map area, the

Cambria fault splits into two branches. The southern branch is the trace described above. The northern branch swings northeast for about 3, 000 feet, turns due east and then southeast again to merge with the Cambria fault. The trace of this splay, herein called the North Cambria fault, results in a small structural block about 9, 000 feet long and 2, 000 feet wide. Overturned Copper Ridge carbonates and Elbrook dolomite constitute the slice. The abrupt termination of the lower Copper Ridge sandstone on both ends sharply delineate the trace of this fault.

Pulaski Breccia. In most places in or adjacent to the map area, the Elbrook dolomite of the Pulaski block is both tightly folded and characteristically cut by low-displacement faults. Associated with these structures are tectonic breccias and conglomerates similar to those described by Cooper and Haff (1940) and Cooper (1946, 1961). Dolomite and limestone clasts in the crush conglomerate, which predominates, range from half an inch to six inches or more in diameter. Typically the matrix is finely ground dolomite of rock-flour consistency. The breccia is irregular in thickness, as exposed along the Norfolk and Western Railway cuts in the map area. In places, this crush conglomerate marks the trace of Cambria fault. Typically, the position of the breccia between competent beds of dolomite or limestone indicates the breccia was formed by slippage along bedding planes. The breccia weathers to a characteristic limonitic buff-yellow, and is marked with holes resulting from the differential leaching of clasts.

New cuts along Interstate 81 in the Pedlar Hills expose numerous zones of tectonic breccia in the Elbrook dolomite. Breccia zones grade into shear zones. The chaotic nature of the Elbrook dolomite in many places in the area supports the idea that the Pulaski-Salem fault is not very far below the present surface. Deformation might also be expected from the emplacement of the overriding Max Meadows thrust sheet.

Origin of Pulaski Thrusting

The origin of the Pulaski thrust sheet has been attributed by B. N. Cooper (personal communication, 1971) to the gravity sliding of the southeast limb of the Appalachian miogeosyncline. Subsequent deformation by vertical displacement involved both overriden and overriding blocks.

Lowry (1965) emplasized the genetic relationships between Pulaski thrusting and the Blue Ridge anticlinorium, and proposed that the Pulaski, Salem, Max Meadows and Blue Ridge faults are all genetically related. He referred to these faults collectively as the Pusabre imbricated mass. Hazlett (1968) postulated a master fault (proto-Pulaski) that developed in the basement of the rising Blue Ridge anticlinorium, and extended upward along the base of the incompetent Rome shale beneath the then still-rooted Salem synclinorium. He showed that as the syncline was detached and moved northwestward, the structurally incompetent Rome became detached from the thrust mass. The Elbrook, Knox and younger beds continued northwestward, permitting Precambrian crystalline and Lower Cambrian Chilhowee rocks to be thrust directly onto the Rome. Then as the "Pusabre" mass continued forward, relatively incompetent lower Elbrook carbonates dropped off. The thrust mass, now with a sole of competent upper Elbrook moved northwestward to its present position. Lowry (oral communication, 1971) thinks the imbricate mass overrode the rising Christiansburg anticlinorium on its way to the trough of the Blacksburg syncline. The mass is postulated to have ridden up the northwest limb of the developing Blacksburg synclinorium, which is the southeast flank of the Poplar Hill-Sinking Creek anticline. This postulated impediment, plus the increased friction from the increase in size of the surface area being overridden, caused the frontal portion of the Pulaski thrust sheet to advance at a slower rate than the rear portion. The build-up of stresses eventually caused an extensive low-angle splay, the Salem fault, to develop upward from the Pulaski fault surface. However, it failed to cut the Pulaski block into two distinct segments. Presumably the splay brought the sole of the Pulaski block to its present elevation in the area southeast of the Salem fault trace.

Lowry (1971) and the writer believe that the Pulaski and Blue Ridge thrusts represent the same master fault. It is postulated that the original detachment occurred above most of the Rome shale. The Max Meadows thrust picked up the previously overridden and still-rooted but deformed Rome and thrust it over the Elbrook of the Pulaski sheet. Thus, the Blue Ridge thrust on the southeast was elevated by the Max Meadows thrusting in respect to its former northwest extension, the Pulaski thrust.

SALEM SYNCLINORIUM

The northern boundary of the map area includes the southern margin of the Salem synclinorium, the north-easternmost large syncline of the Southern Appalachians. Recently when the Salem synclinorium has been discussed in a tectonic framework, it has been referred to as the Catawba block (Ritter, 1969, and Glass, 1970). The largest fold of the synclinorium is the Catawba syncline, whose axis lies about six miles northwest of the Precambrian crystalline rocks of the Blue Ridge anticlinorium. The Catawba syncline is asymmetric and doubly plunging with its depression and recess north of the map area. The synclinorium includes the Slate Lick Run anticline and the Fagg syncline. Lowry (personal communication, 1971) thinks that in the southwestern part of the Salem synclinorium the major axis is the Fagg syncline rather than the Catawba syncline as mapped by Eubank (1967). According to Eubank, the Slate Lick Run anticline and Fagg syncline are complementary structures south of the Catawba syncline. He notes that the Catawba and Fagg synclines were a major structure with the intervening Slate Lick Run anticline developing later. These structures are overridden by Elbrook dolomites of the Salem block.

The Catawba syncline was named by Campbell and Holden (1925) for the major synclinal structure that comprises a major part of the northwest or footwall side of the northeast-trending Salem fault (Figure 2). The axial traces of the Catawba, as mapped by Eubank, and Fagg synclines are parallel to the northeastward regional trend. The southeast flank of the Salem synclinorium steepens and then overturns immediately east of Ironto.

Eubank (1967) proposed the name Mill Knob anticline for a tight anticline complementary to the Fagg syncline on the northwest. The anticline is overturned and overridden by the Salem fault at the north end of Coffee Valley. It lies due east of Fagg along the northern margin of the map area. The overturned northwest flank of the Mill Knob anticline is repeated by a fault which loses displacement to the northeast and dies out near Coffee Valley (Seneca Hollow). The overturning of the southeast flank of the Fagg syncline and the folding of the Mill Knob anticline probably were contemporaneous and prior to the thrusting along the Salem fault. The Salem fault trace closely parallels these structures.

Catawba Block

Strata of the Salem synclinorium range from the Cambrian Copper Ridge Formation up to the coal-bearing Mississippian Price Formation (Campbell and Holden, 1925). Thick Ordovician, Silurian and Devonian sequences are present. Yet unconformities exist between the upper part of the Knox group (Lower Ordovician) and the Middle Ordovician Ellett red beds (Eubank, 1967), and between the Silurian and Devonian systems.

The structural interpretation of the Catawba block follows two schools of thought (Figure 4). The writer distinguishes these as the <u>Catawba-Saltville block</u> interpretation as advocated by Cooper (1961), Ritter (1969) and Glass (1970) and the <u>Catawba-Pulaski block interpretation</u> as advocated by Campbell and Holden (1925), Butts (1940), Lowry (1965), Eubank (1967), Tillman and Lowry (1968), Amato (1968) and Hazlett (1968). Cooper (1961) believes the Catawba block is parautochthonous with the Saltville block which, according to all workers, is exposed in the various windows of the Pulaski thrust sheet. The followers of Campbell and Holden (1925) believe that the Catawba block is allochthonous, the structurally lowest part of the Pulaski thrust sheet.

Catawba-Pulaski Block Interpretation

Under this interpretation the Salem synclinorium is part of the leading edge of the northwest-displaced Pulaski structural mass. At least 10 miles of horizontal movement is postulated. The root zone of the Catawba block would be the Riner syncline northwest of the adjacent Blue Ridge anticlinorium. The time of this displacement would be Mississippian or later, as Mississippian beds are the youngest strata of the block.

Critical to this interpretation is the location of the Pulaski fault trace (Figure 4, top). The writer and other previous proponents of this original interpretation agree with Campbell and Holden who mapped the Pulaski fault trace on the northwest side of the Catawba syncline at the southeast foot of Brush Mountain. This interpretation requires that the Pulaski fault in the writer's area continue essentially horizontally under the present land surface north of the Christiansburg window. Its trace would not come to the surface on the south side of the Catawba block but would continue under the block. It is the writer's contention that the Salem block of his mapped area represents a structurally raised southeastern flank of the Salem synclinorium.

The stratigraphic considerations of the Salem synclinorium support the writer's contention that it is the leading edge of the Pulaski thrust sheet. There are several distinct differences between the lithologies of the Salem synclinorium and that of the adjacent Saltville block strata exposed in the Christiansburg window. The most striking difference is the apparent absence of the Bays sandstone in the Christiansburg window, and its thick and coarse accumulations to the northwest in the Salem synclinorium. Source direction of the Bays is from the southeast, thus the interpretation that the original deposition of the Salem synclinorium clasts was to the southeast of the Christiansburg window rocks. The writer does note, however, that it is possible that the Christiansburg anticlinorium was positive in Bays time, and was bypassed by sediment transport mechanisms. The second most striking difference is that the Middle Ordovician limestones are a reefy facies in the Christiansburg window, unlike those of the Pulaski block and Salem synclinorium. The lower part of the Liberty Hall Formation can also be considered. Its brownish green shales in the Salem synclinorium are not found in the Christiansburg window, where the Liberty Hall is a dark gray banded limestone. Many less striking differences are also noted. Among these are trilobite-bearing limestones of the lower Knox in the Saltville block window rocks, but not exposed in the Salem synclinorium (Broughton, 1971). There is possibly more chert in the upper Knox of the window rocks than in the synclinorium. Silurian and Devonian units are not markedly different. The Fagg-type conglomerates and the fossiliferous Needmore shale of the synclinorium are not recognized in the window. The Oriskany age sandstones are distinctly finer grain in the window rocks. Thus, the writer takes issue with the statements of Cooper and Cashion (1970), who contend that the Salem synclinorium is the same structural unit as the Saltville block strata. The writer's interpretation requires that Mississippian age rocks be deposited and preserved further southeast (by at least 10 miles) than any other known locales. Cooper and Cashion (1970) state that the Salem synclinorium succession does not resemble that found on the Pulaski block anywhere to the southwest in Virginia or in Tennessee, but do not cite specific evidence leading to this speculative conclusion. The writer feels as though he has demonstrated sufficient differences between the Saltville block and Salem synclinorium to warrant consideration of the Salem synclinorium as part of the Pulaski block.

Catawba-Saltville Block Interpretation

As noted, Cooper (1961) Ritter (1969) and Glass (1970) are the main proponents of Catawba-Saltville block interpretation (Figure 4, bottom). They propose that the Catawba block is parautochthonous with window rocks to the southwest. Cooper postulated that the Pulaski fault instead of continuing northwestward along the base of Brush Mountain swings southward through Blacksburg and then turns eastward to join or underlie the Salem fault along the southern margin of the exposed Salem synclinorium.

Ritter (1969) proposes that his Catawba fault (Campbell and Holden's Pulaski fault) along the northwest edge displaced the Catawba block west-northwestward, riding over what is now the Saltville block of the Price Mountain window. Subsequently, the Pulaski fault block overrode both the strata now exposed in the Price Mountain window and the postulated overlapping Catawba block. Ritter (1969) gives an estimate of 10,000 feet of stratigraphic displacement to the eastward-dipping Catawba fault on the basis that Cambrian rocks of the Catawba block overlie Mississippian strata of the Price Mountain anticline. Ritter further postulated that the high-angle reverse Yellow Sulphur fault finally cut the older Catawba and Pulaski faults. The surface expression of the Catawba fault, as proposed by Ritter (1969), is the fault trace along the base of Brush Mountain, which Campbell and Holden (1925) termed the Pulaski fault. After all the above thrusting, Ritter (1969) postulated that the entire Pulaski thrust sheet which had overridden the Salem synclinorium east of the Yellow Sulphur fault trace was eroded.

MAX MEADOWS THRUST SHEET

The Max Meadows block constitutes the southern margin of the writer's map area. Like the Elbrook of the Pulaski block to the north, the Rome Formation, which composes essentially the entire block, is complexly folded and faulted on a small scale. Strikes of the Rome Formation are generally east-northeast and parallel the leading edge of the fault.

The Max Meadows thrust fault was named by Cooper (1939) after the town of Max Meadows in the Draper Mountain area of southwestern Virginia. The Max Meadows fault has been traced from near Pulaski northeast to the Roanoke area. In and adjacent to the map area the fault approximately parallels the traces of the Salem fault to the northwest and the Blue Ridge fault to the southeast. In the Christiansburg area, Dietrich (1954) originally called the Max Meadows fault the Christiansburg thrust. Hergenroder (1957) demonstrated near Radford that the Christiansburg thrust is equivalent to the Max Meadows thrust. The trace of the Max Meadows fault in the Christiansburg region closely parallels the southern margin of the Christiansburg window to nearly the eastern edge of the writer's area. Along this eastern edge (Figures 2 and 3) the fault trace turns north and approximately follows the coarse of the South Fork of the Roanoke River. The Max Meadows fault trace continues east of the writer's map area, northward to the Pedlar Hills.

The writer mapped about 30 discrete or isolated areas of Rome Formation spread over the eastern half of the area. These include three small patches, less than 20 feet across, of breccia composed of red Rome shale and Elbrook carbonate. The first of these breccia zones is along the railroad tracks at Montgomery Station, southwest of the railroad tunnel. It is a coarse facies of tectonic conglomerate (Cooper, 1944, 1946). Blocks as much as 20 feet in diameter are enclosed in a chaotic mixture of red Rome shales and Elbrook Formation dolomite. Another tectonic mixture of Rome and Elbrook also occurs along the Norfolk and Western Railway tracks at the north end of Poplar Hollow. A third is 3, 700 feet east of the latter, exposed in a roadcut along U. S. Route 11, 1.7 miles west of Shawsville (Cooper, 1968). As the Pulaski-Salem fault is not far below the surface, these three examples may represent breccia generated during emplacement, and brought to the present surface along fractures and faults in the lower part of the Pulaski block.

As Cooper and Haff (1940) pointed out, the base of the Max Meadows overthrust sheet has been dynamically metamorphosed to a phyllite. They denote three zones of Max Meadows brecciation. The zone of crush conglomerate grades into autoclastic breccia zones above and below, and is formed by "the mingling, crushing and rolling out" of the parts of the autoclastic breccia. The crush conglomerates are well exposed along the Max Meadows fault trace in the southwestern portion of the map area. Typically, individual clasts of dolomite average between one and two inches in diameter in a matrix of dark-gray to leached-limonitic yellow, finely granulated dolomite, and soft crumbly macerated phyllite.

The Max Meadows fault is younger than Salem faulting and probably represents later movement at or near the trailing edge of the Pulaski block, as the leading edge become impeded (Tillman and Lowry, 1968; Lowry, 1965). The Max Meadows thrust probably extends downward and intersects the Pulaski fault. According to Lowry (1971) the fault cuts across and extends beneath the Pulaski thrust. Regardless of the interpretation, Max Meadows thrusting throws Rome beds of the Riner synclinorium over post-Rome carbonates of the Pulaski block proper.

Several interpretations of the origin of the 30 bodies of Rome shale on the Pulaski-Salem block are possible. As pointed out, the writer attributes the small, low-elevation, macerated Rome phyllitic breccias to the squeezing upward from a lower level of the Pulaski block during its emplacement. Larger, more coherent bodies may be attributed to irregularities in the lower surface of the Pulaski block. With this interpretation, the upper portions of the Rome Formation would have been trapped and held in depressions of the sole of the Pulaski block and carried along as the whole block advanced. Salem and associated faulting would bring up these masses. The writer would attribute the presence of Rome beds north of Den Creek, and perhaps those in Wells Hollow and Falling Springs Hollow to this mechanism. These outcrop areas characteristically occur in the lowest valleys of the map area. Presumbabaly, the Pulaski-Salem fault is near the surface at these points.

Remaining outcrops of Rome are interpreted by the writer as klippen of the Max Meadows thrust sheet. Most of these Rome exposures cap ridge crests but some are also found at intermediate elevations of lower slopes. These klippen are recognized as part of the Rome Formation by the presence of red shales. Significant areas of Rome Formation klippen may have gone unrecognized because of poor exposure of rocks of this formation. Several areas of the Rome Formation mapped as distinct klippen may actually interconnect and comprise larger and more extensive masses. The largest Max Meadows klippe recognized in the writer's area is that at Clinkum Hollow. The klippe is 8,000 feet long and approximately 3,500 feet wide (Figures 2 and 3).

Red shales in the map area were considered Rome Formation along fault contacts in view that the Elbrook Formation in southwestern Virginia lacks red shales. Locally in the area, however, it is possible that some of the smaller outcrops of red shale are in reality basal Elbrook.

The writer believes that the Max Meadows thrust sheet originally extended at least as far northwest as the north side of Interstate 81 in the map area. If so, the fault has a minimum of 12,000-15,000 feet of horizontal displacement. Cooper (1939) attributes 7 or 8 miles of horizontal displacement to the Max Meadows fault. Ritter (1969) considers the Rome shale exposed in the Blacksburg area to be erosional remnants of the Max Meadows block; this implies a horizontal movement of about 10 or 11 miles.

Hergenroder (1957) notes that the Max Meadows fault originally extended to the southern margin of the Ingles-Barringer Mountain window, but was later eroded back to its present position south of Saltville block rocks now termed the western end of the Christiansburg window complex. The present writer contends that the Max Meadows fault originally extended as far north as a line from the south side of the Ingles-Barringer Mountain window to the trace of the Salem fault along the southern margin of the Salem synclinorium at Clinkum Hollow. This implies that more than two-thirds of the writer's area (Figure 2) was once covered by the Max Meadows block.

Max Meadows Block Erratics

Copper Ridge sandstone blocks east of the Christiansburg window are believed to have been carried northwestward by the Max Meadows thrust sheet. This concentration of sandstone blocks would constitute a section of Knox clipped off by the fault from the south limb of the Christiansburg anticlinorium (Glass and Lowry, 1970). Lowry (1971) and the writer envision the Max Meadow fault as originating below the Pulaski fault surface and cutting upward through Rome to intersect locally, such as south of the Christiansburg window, the older Pulaski thrust before continuing surfaceward.

Numerous siliceous erratics were observed by the writer along the Max Meadows fault trace at the southern margin of the map area. Most are white to grayish-white chert and granular chert cobbles 6 to 8 inches in diameter.

SEQUENCE OF THRUSTING

The youngest beds exposed in the map area are Devonian black shales (Millboro) of the Salem synclinorium. Younger Mississippian beds of the Price Formation are exposed in the trough of this syncline northeast of the area. Northwest of the map area and southwest of Blacksburg, the Price Mountain window exposes the Mississippian Stroubles Formation, which overlies the Price Formation.

The Price Mountain anticline of the parautochthonous Saltville block is partially covered by the Pulaski block. It is assumed that Pulaski thrusting occurred soon after the deposition of the Stroubles Formation and thus is late Mississippian in age.

The Pulaski fault is believed to be the oldest fault of the map area if the Salem synclinorium is considered part of the Pulaski block. If the synclinorium is considered parautochthonous Saltville block, then the Catawba fault predates Pulaski faulting. As the Salem fault is considered by the writer to be a complication of Pulaski thrusting, it would be younger than the Pulaski, but older than the Max Meadows fault to the south. The Cambria, North Cambria and Kettle Ridge faults, it is proposed, were developed in connection with Salem thrusting. The Willow Springs fault is believed to be associated with the continued rise of the Christiansburg anticlinorium and thus postdates the emplacement of the Pulaski-Salem thrust sheet.

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UPPER CRETACEOUS MARINE TRANSGRESSION IN NORTHERN DELAWARE

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ABSTRACT

The first marine transgression in the Coastal Plain of northern Delaware occurred in Upper Cretaceous time and is shown by the Magothy and Merchantville Formations.

The Magothy Formation is composed of sands, which were mostly deposited in a small delta and an offshore channel, and clayey silts deposited in estuaries, back-bays and marshes. The source of most of these sediments was the Potomac Formation (staurolite-zircon-tourmaline-kyanite heavy mineral zone, Groot, 1955) that was exposed landward (northward) of the transgressive sea.

The Merchantville Formation is composed mostly of silty clays and fine, silty sands that contain glauconite, marine fossils and a considerable amount of organic matter. These sediments were probably deposited in estuaries and marshes in the northern portion of the study area, and in shallow littoral and neritic sea in the rest of the area.

The contact between the Magothy and overlying Merchantville Formations is gradational at least in the Magothy wedge-out zone. However, down-dip the contact is sharp; this is believed to indicate a change in the kind of sediments supplied to the area rather than a break in deposition.

Sometime after the deposition of the Magothy Formation the area of study was affected by tilting. The exact time and cause of this movement is unknown.

INTRODUCTION

The first transgressive unit in the Coastal Plain of northern Delaware is the Magothy Formation. It is composed of white, light gray, and gray sands and layers of dark gray to black silts. The sands are frequently cross-bedded and not uncommonly contain lignite. The silts, in addition to lignite, also contain considerable amounts of other disseminated organic matter.

The Magothy Formation was first described by Darton (1893) along the Magothy River in Maryland; it was later recognized in Delaware and New Jersey by Clark (1904). In Delaware these sediments crop-out along the western part of the Chesapeake and Delaware Canal. The Magothy was first mapped in Delaware by Bascom and Miller (1920). This formation is considered to be transitional between the underlying continental sediments of the Potomac Formation and overlying marine deposits of the Merchantville Formation.

There is some disagreement among the geologists on the age of the Magothy Formation. Berry (1911) and Richard (1967) cite it as late Cenomanian-early Turonian. Doyle (1969) considers it to be Santonian. Dorf (1952) gives a Coniacian age and Groot, Penny and Groot (1961) late Turonian to early Coniacian. However, it is possible that the age of this formation varies with its location (Jordan, 1970).

The Merchantville Formation is composed of dark blue to black, micaceous, glauconitic silty clays, and dark, greenish-brown, micaceous, glauconitic fine silty and clayey sands. It is the oldest known marine unit in the Coastal Plain in Delaware, and it was recognized as a formation here by Groot, Organist, and Richards (1954).

The exact age of the Merchantville Formation is unknown. Weller (1907) described a fauna of 102 species of Senonian age. The microfauna studied by Gill (1957) suggests a correlation with the Taylor of Texas. Mumby (1961) proposed an upper Campanian age for the Mt. Laurel Formation in Delaware; this formation is stratigraphically higher than the Merchantville. Sohl and Mello (1970) consider it to be of early Campanian.

The purpose of the present study is to investigate the origin and depositional environments of both the Magothy and Merchantville Formations in a small area in northern Delaware.

Acknowledgments

Robert R. Jordan and Thomas E. Pickett, Delaware Geological Survey, University of Delaware, critically read the manuscript.

Method of Study

The area of present study is located west of Delaware City (Figure 1). Twenty-four holes drilled in this area reached the Potomac Formation and they were all electrically logged. Data for lithofacies analyses were obtained from the electric logs. The areas under the SP curves were measured planimetrically and the sedimentary parameters necessary for the construction of lithofacies maps were computed in the following manner:

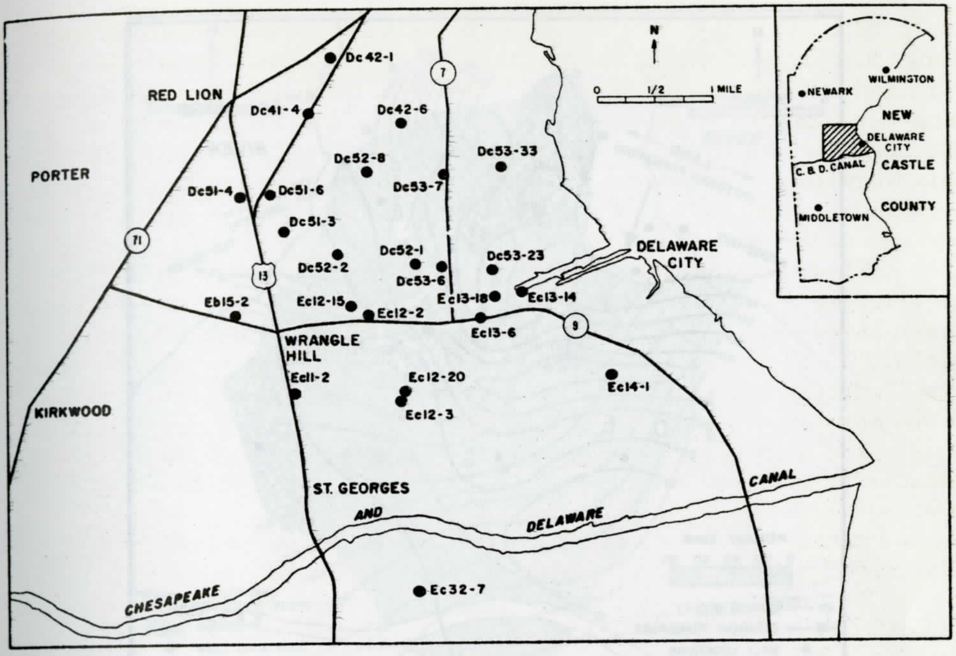

Figure 1. Area of investigation showing well locations. The well numbering system is that used by the Delaware Geological Survey.

$$\frac{(SP \cdot H) \cdot f}{H} = SP$$

$$\frac{SP}{SSP} = \Phi$$

sand thickness = ss = $H \cdot \Phi$

clay thickness = sh = $H - (H \cdot \Phi)$

sand percent = $\frac{ss}{H} \cdot 100$

clay percent = $\frac{sh}{H} \cdot 100$

Explanation of the symbols

H = thickness of formation

SP· H = planimeter reading

f = planimeter correction factor

SSP = static self-potential

Heavy mineral analyses utilized in this study were taken from Groot (1955).

RESULTS

Lithofacies Analysis of the Magothy Formation

The lithofacies map of the Magothy Formation (Figure 2) reveals a small delta in the western part of the study area. Sand distribution suggests that the delta was fed from the north, probably by a small

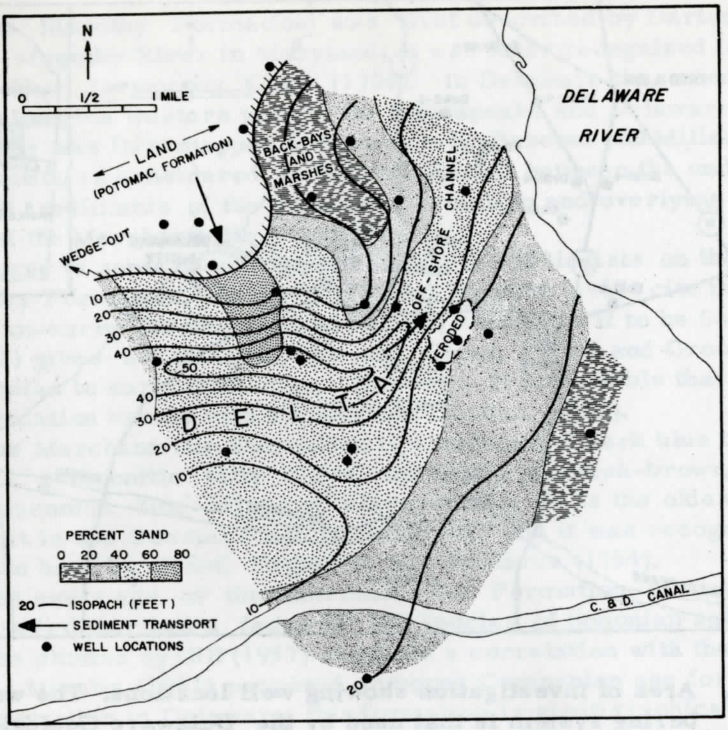

Figure 2. Lithofacies map of the Magothy Formation.

river. It seems that the sediments brought into the delta were partly deposited there, and partly carried farther toward the northeast and laid down in an offshore channel (Figures 2 and 3). This offshore channel was either a valley in the underlying Potomac surface or was formed by the longshore currents, or both. The sediment transport in the channel seems to have been toward the northeast, as suggested by the sand distribution, following the configuration of the Magothy wedge-out.

On the seaward side (toward the southeast) the study area appears to have been bounded by islands (Figure 3). These islands were topographic highs in the underlying Potomac surface and were submerged in the later phases of the transgression, thus forming shoals. Both the islands and later shoals acted as protective barriers to the direct influence of open sea waves.

Portions of the study area that were devoid of the direct contributions of sand size materials through the delta were the depositional sites for fine clayey silts containing abundant organic matter. Such areas were located northeast of the delta and are interpreted as marshes and back-bays (Figure 2).

Marine fossils are absent from the Magothy Formation in the area of study. This probably signifies a considerable influence of fresh

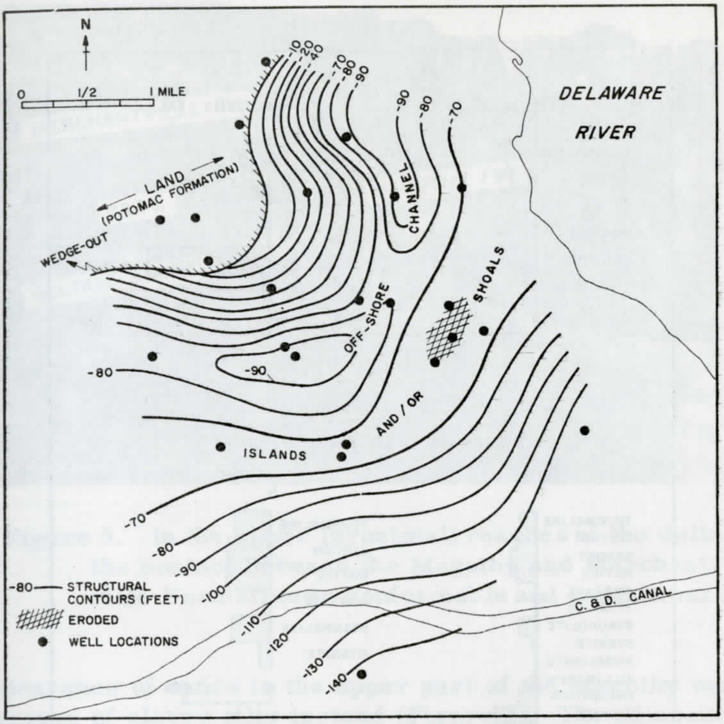

Figure 3. Structural map of the base of the Magothy Formation (sea level datum).

water brought into this area through the delta. The presence of islands (in early phases) and shoals (in later phases of the transgression) apparently prevented effective mixing of this fresh water with the saline water of the sea.

Magothy Source Area and Its Modifications

Groot (1955) proposed that the crystalline rocks (schists and gneisses) of the Appalachian Piedmont, similar to the ones that supplied materials for the build-up of the staurolite-zircon-tourmaline-kyanite heavy mineral zone of the Potomac Formation (exposed northward of the Magothy wedge-out) in Barremian - Aptian(?) time (Doyle, 1969), were again reactivated in Upper Cretaceous to produce the Magothy Formation. His conclusion is based primarily on the similar heavy mineral compositions of the two formations. Although there is no proof that Groot's suggestion is invalid, a different hypothesis about the source of the Magothy sediments is proposed here.

The remarkable similarity between the heavy mineral compositions of the Potomac and Magothy Formations (Figure 4) is believed to

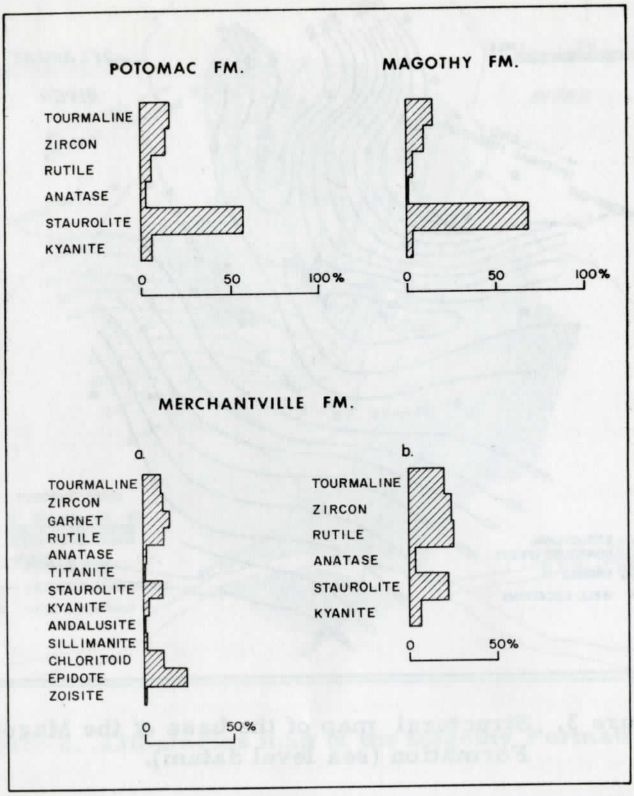

Figure 4. Heavy mineral assemblages of the Potomac (staurolite-zircon-tourmaline-kyanite zone; Groot, 1955), Magothy, and Merchantville Formations. The heavy mineral composition of the Merchantville Formation is distinctly different from those of the Magothy and Potomac Formations. Column "b" shows only those minerals found in the Magothy and Potomac and is included for comparison purposes only.

indicate that the Potomac Formation was the source for the Magothy sediments. If this is true then the streams that brought these materials from their sources into the depositional sites were short and their drainage basins were located almost entirely in the Coastal Plain area. Denudation of the Potomac source area eventually lowered the stream gradients to a point where these streams became unable to continue to supply sand to the delta. At this stage the delta sedimentation ceased and the deltaic environment was replaced by estuaries; this is shown by

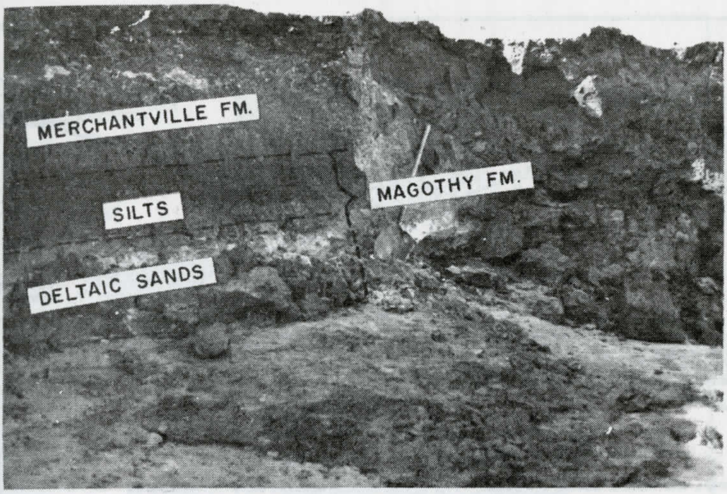

Figure 5. In the upper (proximal) reaches of the delta the contact between the Magothy and Merchantville Formation is conformable and transitional.

the disappearance of sands in the upper part of the Magothy section and the appearance of clayey silts instead (Figure 5). The adjacent Potomac land (source area) probably became flat and almost featureless and along the coast extensive marshes and tidal flats developed. This is suggested by the presence of disseminated organic matter and lignite in these deposits. Decreasing fresh water influence is indicated by the first appearance of marine fossils; this marks the beginning of the Merchantville deposition.

Thus it seems that the transition from the Magothy sedimentation into true marine deposition (Merchantville Formation) was a gradational process without breaks in deposition (Figure 6). Farther south (down-dip), however, the contact between the deltaic Magothy sands and overlying Merchantsville sediments is sharp (Figure 7). This is believed to signify a change in the kind of sediments supplied to this area and also to reflect the processes that modified the adjacent source area. Specifically, the denudation of the source area that produced a change from deltaic to estuarine depositional environments had an effect on the upper (proximal) reaches of the delta indicated by the deposition of the Magothy silts. The effect on the lower (distal) reaches of the delta was slight or none and here the Merchantville was laid down directly over the Magothy deltaic sands (Figure 5).

Lithofacies Analysis of the Merchantville Formation

The Merchantville Formation is characterized by a relatively uniform lithology in the area of study. The lithofacies map (Figure 8)

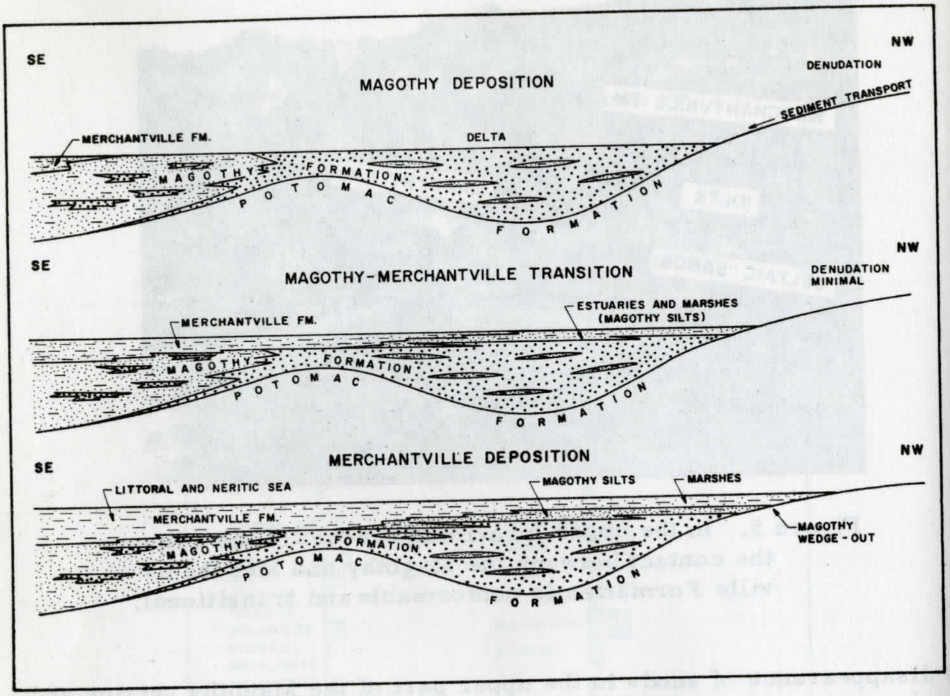

Figure 6. Depositional processes and environments which are believed to have accompanied Upper Cretaceous marine transgression in the area of present study.

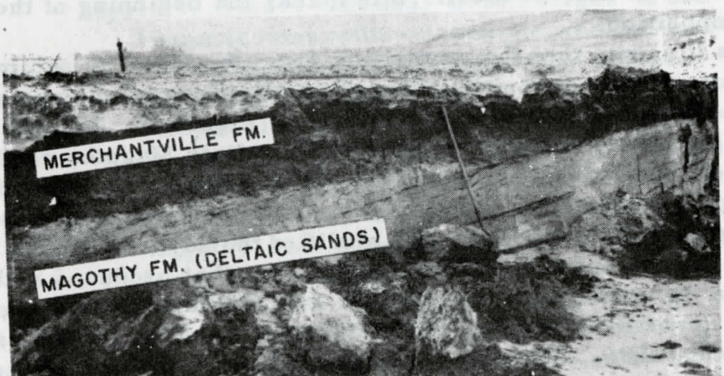

Figure 7. In the lower (distal) portions of the delta the contact between the Magothy and Merchantville Formations is sharp; Magothy deltaic sands are directly overlain by the Merchantville sediments.

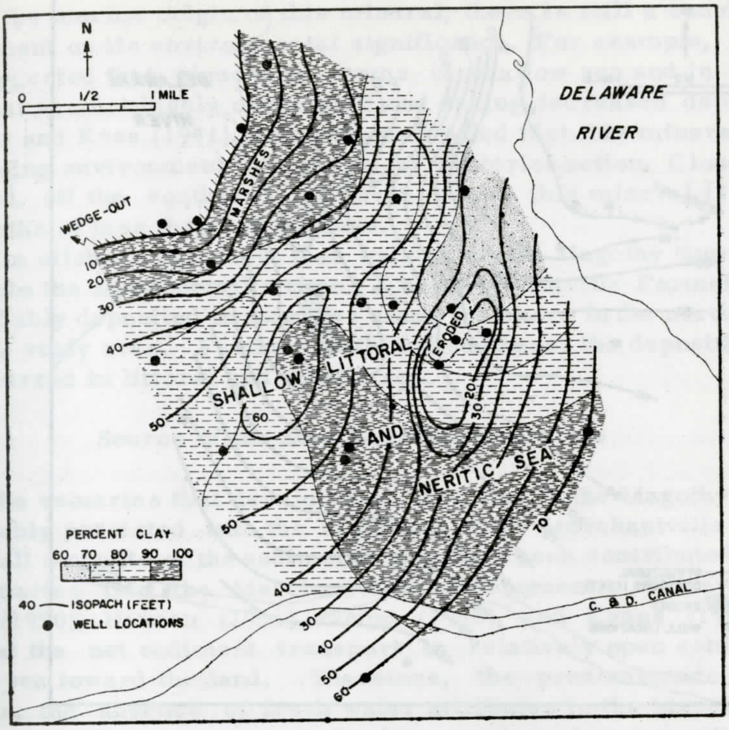

Figure 8. Lithofacies map of the Merchantville Formation.

shows that as a whole, this formation is composed of fine sediments, silts and clays. This is in slight disagreement with the observed lithologies of this formation in the outcrops where it is composed of fine clayey sands and silts. The reason that the electric logs have recorded the Merchantville Formation in the subsurface as very clayey is the fact that the amount of materials smaller than two microns, in both sands and silts, often exceeds 40 percent.

Structural contour map of the base of the Merchantville Formation (Figure 9) indicates that the Magothy offshore channel was not completely filled with the Magothy sediments at the onset of the Merchantville deposition. The remnant of this channel seems to have acted as a trap for the thickest accumulation of the Merchantville sediments.

Sedimentary structures in the Merchantville Formation in the study area are usually obscured or destroyed by post-depositional changes (shown by the presence of authigenic pyrite and siderite) and by fossils. When present, however, the sedimentary structures appear to be represented by thin bedding (Figure 10).

In addition to marine fossils the presence of glauconite also indicates true marine origin for these sediments. Although the geologists

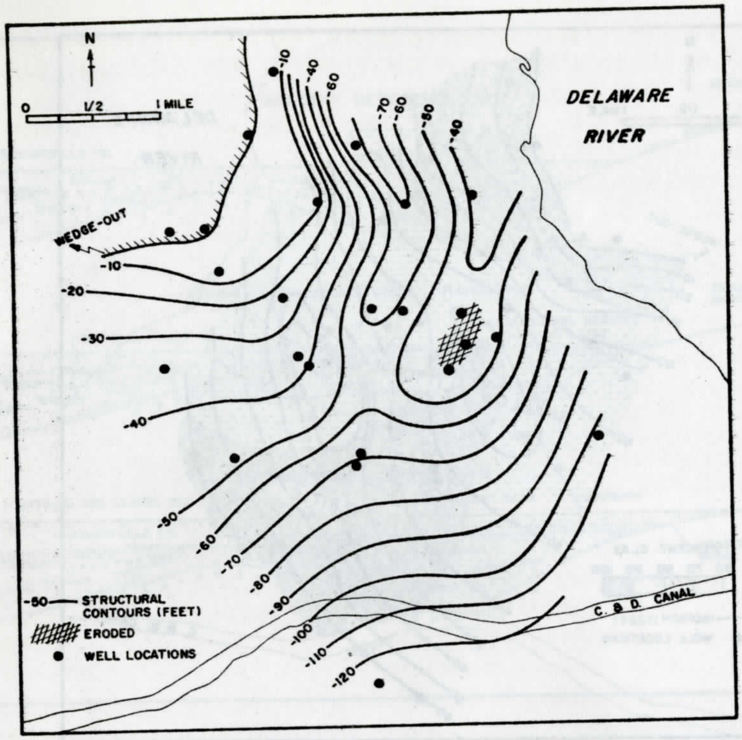

Figure 9. Structural map of the base of the Merchantville Formation; sea level datum.

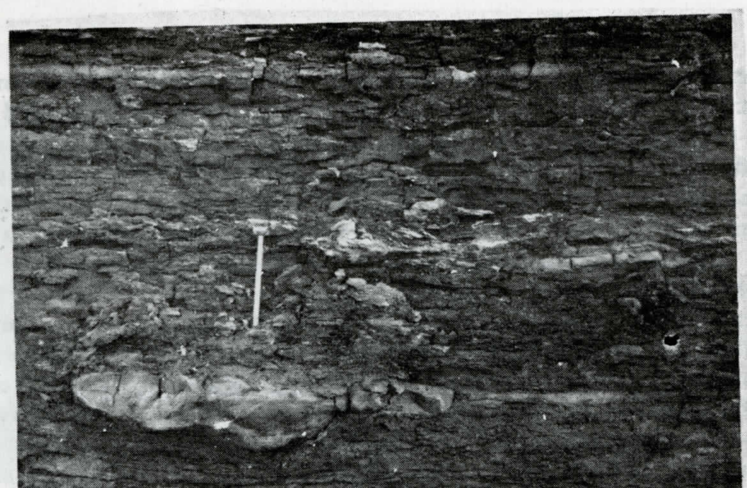

Figure 10. Sedimentary structures in the Merchantville Formation, when preserved, are represented by thin bedding. Siderite and pyrite nodules are common in these sediments and are elongated parallel to the bedding planes.

agree on the marine origin of this mineral, there is still a considerable disagreement on its environmental significance. For example, Hadding (1932) suggested that glauconite forms in shallow sea and in agitated waters that are not highly oxygenated and during decreased deposition. Hendricks and Ross (1941), however, concluded that this mineral forms in a reducing environment maintained by bacterial action. Cloud (1955) found that, off the southern California coast, this mineral is rare in water depths of less than 100 fathoms.

The offshore barriers that existed in the Magothy time did not persist into the Merchantville time. The Merchantville Formation was most probably deposited in marshes along the shore in the northern portion of the study area. Farther south and southeast the deposition probably occurred in littoral and neritic sea.

Source of the Merchantville Sediments

The estuaries that developed at the close of the Magothy deposition probably persisted into the beginning of the Merchantville time as well. Small amounts of the sediments may have been contributed through such estuaries into the Merchantville environments. However, Van Straaten (1950), Hansen (1951), Gripp (1956), and Evans (1965) have shown that the net sediment transport in relatively open estuaries is from the sea toward the land. Therefore, the predominance of fine sediments, the absence of fresh water attributes in the Merchantville sediments and their characteristic heavy mineral composition, are thought to indicate that these materials were transported by the action of the sea into the depositional environments. This is in general agreement with Groot's (1955) conclusion although there is no evidence that the source area was located immediately south of Delaware, as he suggested.

TILTING OF THE AREA

In the present study area the Magothy sediments are completely covered by the Merchantville Formation. Therefore, it is believed that the Magothy wedge-out represents the original northward extent of these sediments. For this reason the structural contour map of the base of the Magothy (Figure 3) has been investigated to determine any possible tectonic disturbance that may have occurred in the area.

The intersection between the structural contours and the wedge-out in the northwestern part of the area shows that the area indeed tilted toward the east.

In the northern part of the area the Merchantville Formation (including its wedge-out) is directly overlain by the Pleistocene fluvial sediments. Deposition of the Pleistocene sediments was accompanied by deep erosion of the underlying older sediments, as shown by

Spoljaric (1971). Therefore, there is little doubt that a considerable amount of the Merchantville Formation was eroded and removed from this area in the Pleistocene time, and the Merchantville wedge-out, as seen in Figures 8 and 9, most probably, does not coincide with the original northward extent of the Merchantville transgression. Because of this the age of the tilt is unknown, except that it occurred in the post-Magothy time. This is not in agreement with Groot's (1955) conclusion that the movement took place contemporaneously with the deposition of the Magothy. If, in fact, the tilt was contemporaneous with the deposition, the structural contour map would not reveal such a movement.

CONCLUSIONS

The results of the present study suggest that the deposition of the Magothy and Merchantville Formations took place simultaneously; while the Magothy sediments were laid down along the shore, the Merchantville was deposited farther south, perhaps even beyond the boundary of the present study area. The study has also revealed the existence of considerable down-dip and along-the-strike lithological variations within the Magothy Formation. Because of this, and because of the small size of the area of study, no attempt was made to compare the results of this study with the results and findings made by other workers on both the Magothy and Merchantville Formations (or their equivalents) in other parts of the Atlantic Coastal Plain. Such a comparison would most probably be an oversimplification and conclusions based on it misleading.

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ASPECTS OF THE TEXTURE AND MINERALOGY OF SURFICIAL SEDIMENTS, HORRY AND MARION COUNTIES, SOUTH CAROLINA

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ABSTRACT

Horry and Marion Counties, South Carolina, have a range of sedimentary deposits that offers an ideal situation for the study of sediment mixing by both marine and fluvial processes. Size properties, mineralogy of the sand and clay fractions, and aspects of the weathering profiles were used to infer the contribution from Piedmont and Coastal Plain sources.

When abundance of heavy minerals is compared with mean size of sampled sediment there appears to be little relationship in the case of coastal barrier samples, but a strong association in the case of samples collected in the river valleys, particularly for epidote and hornblende. It is possible that this reflects mixing above the study area for fluvial sediments, and mixing within the area for the deposition of barrier sediments. The older the coastal barrier the more stable heavy minerals such as zircon and staurolite tend to dominate. Likewise, the Waccamaw and Little Pee Dee valleys possess a more stable assemblage

of minerals compared with the Great Pee Dee which receives sediment from the Piedmont.

Clay mineralogy was not as useful a tool for differentiation of geomorphic units because the effects of soil development has tended to produce the same minerals in sediments of diverse origin. However, characteristic suites for each of the deposits are recognized. Clay mica is most abundant in back barrier flat sediments. Weathering profiles are important in the consideration of size and mineral assemblage relations. In addition to the variation of total iron, the degree of weathering also influences the grain size configuration of the A_2 horizon and the stability of the mineral assemblage.

INTRODUCTION

The presence of a range of sedimentary deposits of different age on the Coastal Plain in Horry and Marion Counties, South Carolina, has provided the writers with the opportunity to examine textural and mineralogical variations within a regional framework. Samples were collected near the surface (upper 6 feet) from five coastal barrier and backbarrier flats, as well as fluvial flood plains of the Waccamaw, Great Pee Dee and Little Pee Dee Rivers. Analyses of textural properties, together with the determination of heavy and clay mineral suites and total iron, were conducted on the samples. Accordingly, the paper has two objectives: one, to examine variations in mineralogic composition with texture (expressed as mean grain size) and with the age of surficial deposit; and two, to examine the relative contribution of the two source areas, the Piedmont and the Coastal Plain.

Data considered in this paper were collected during the course of a general study of the geomorphology, stratigraphy, and sedimentology of Horry and Marion Counties, South Carolina. Aspects of this study have been reported elsewhere (Thom, 1967, 1970; Adams and Thom, 1968; DuBar, 1969 and in preparation). The paper by Adams and Thom (1968) reviews the problem of areal variation in size distribution properties from samples collected near the surface, and the present paper elaborates on the general theme of sediment distribution.

Geologic mapping has shown the occurrence of five depositional surfaces of coastal origin within the study area (Figure 1), each surface being composed of two outcropping facies (barrier and backbarrier flat). For the purpose of this study each surface is designated by informal names: Recent, Myrtle, Jaluco, Conway and Horry. A fluvial facies sometimes extends inland along river valleys (Thom, 1967, Figure 4). The study of sediment distribution revealed statistically significant variation in mean and standard deviation properties between barriers of different age (Adams and Thom, 1968). Backbarrier flat sediments, however, showed more variation within a given surface than between surfaces.

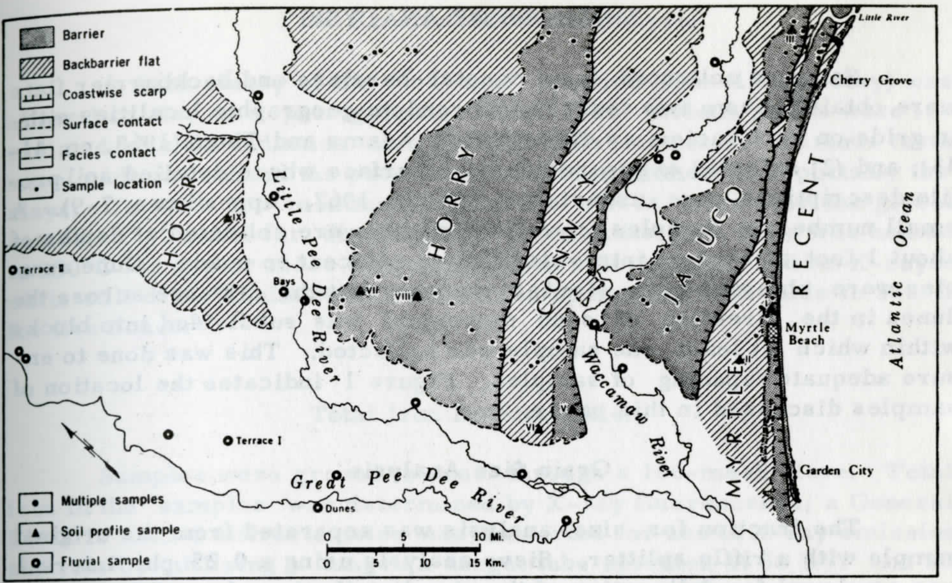

Figure 1. Coastal depositional surfaces, Horry and Marion Counties, S. C., showing location of samples used for grain size, heavy mineral, clay mineral, and total iron determination.

The observed differences in sedimentary size properties between barriers of different age can be the result of several factors, some of which may be interacting. These include variation in primary size distributions, different primary mineral composition, and the effects of weathering. Thom (1967) has shown progressive degrees of soil profile development with age of barriers. This suggests that destruction of the less resistant minerals of the primary deposit is a factor affecting grain size variation in surficial sediments. The addition of weathered fines by translocation from above is another possible source of weathered products.

Acknowledgments

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METHODS OF ANALYSIS

Samples collected from coastal barriers and backbarrier flats were obtained from two sources: (1) random geographic localities within grids on each facies, as described by Adams and Thom (1968, pp. 41-43); and (2) selected exposures on each surface where detailed soil profile descriptions were undertaken (Thom, 1967, Appendices 7-9). A small number of samples from flood plains were collected at depths of about 1 foot at random intervals at sites adjacent to roads. Dune samples were obtained in a similar manner, but the distance across the dunes in the direction of sand movement was subdivided into blocks within which at least one sample was collected. This was done to ensure adequate spacing of samples. Figure 1 indicates the location of samples discussed in this paper.

Grain Size Analysis

The fraction for size analysis was separated from the original sample with a riffle splitter. Sieve analysis using a 0.25 phi interval was employed for delineation of the size distribution of particles coarser than 4 phi (0.0625 mm). Hydrometer analysis described by ASTM Committee D-18 (ASTM, 1964) was employed for sediments with an appreciable fraction finer than 4 phi. The calculations were made by a computer program prepared by Adams (1967). Statistics of the textural characteristics were calculated by a modified version of the computer program described by Kane and Hubert (1963). Inman's statistics, the median, and the percentage of sand, silt, clay were found to be the most useful measures for these sediments (Krumbein and Pettijohn, 1938; Inman, 1952).

Heavy Mineral Analysis

Samples of 40 grams were acidified in 10 per cent dilute HCl, washed through a 0.062 mm screen to remove clay and silt, oven dried, and weighed for net sand. Heavy minerals were separated from the sand by bromoform, and each heavy mineral crop was weighed to 0.001 grams. Knowing the weight of the heavy mineral crop per unit of sediment permits the calculation of the approximate weight in milligrams of each heavy mineral. The heavy minerals were then split in a microsplitter and mounted in Caedex. The Doeglas (1940) method was used in counting heavy minerals. Doeglas counts exactly 100 grains, notes the percentage of opaques, then continues counting until 100 transparent minerals have been identified and recorded. This eliminates the masking effect of the more abundant opaques and focuses attention on the more diagnostic transparents. No distinction was made among ilmenite, magnetite, and opaque rutile.

Fine Fraction Mineral Analysis

A thick slurry of the fine mud fraction (fine silt and clay) was allowed to dry on a petrographic slide. The various slides were irradiated with Cu radiation on a Phillips diffractometer. Various tests were made to confirm the presence of different minerals especially the clays. To test for montmorillonite the slides were ethylene glycol saturated. To test for dioctahedral vermiculite each slide was heated to 350°C for two or more hours, cooled in a desiccator, and re-X-rayed while still dry. Kaolinite was confirmed by heating each slide to 550°C for five hours.

Total Iron Determination

Samples were ground to pass through a 100-mesh sieve. Total iron in the samples was determined by X-ray fluorescence; a General Electric XRD-5 diffractometer was converted for use in X-ray emission analysis by utilizing a tungsten target tube operated at 50 KVP and 10 ma, 2 seconds time constant, and lithium fluoride as the analyzing crystal. Specimens were prepared by mixing the sample powder in a small Bakelite holder with drops of amyl acetate to make a paste, which was then smoothed with a glass slide. The FeK (alpha) radiation excited at 57.52° theta angle was counted for 10 seconds with a gas flow proportional detector. Six countings were taken for each sample, and an average was used for calculation.

A standard calibration curve was also prepared as follows. One to 30 per cent of pure iron powder in finely ground silicon dioxide (325 mesh) was weighed into polyethylene vials. Each sample was thoroughly mixed on a Spex vibrating mixer for five minutes. Specimens were prepared and the FeK (alpha) radiation for each standard was counted in exactly the same manner as described above. A straight line was obtained by plotting the counts per 10 seconds versus percentage of Fe. Thus, the percentage of Fe in the samples could be obtained directly from the standard curve. The SiO_2 powder was chosen in the standards in order that the matrix effect could be minimized, since the samples contain largely quartz sand in addition to the iron compounds. It is possible that the values are slightly higher than they should be because this technique tends to exaggerate the presence of iron. No attempt has been made to apply a correction factor by checking the analyses on an atomic absorption spectrophotometer. It is considered that the values reported here reveal an accurate pattern of the relative variation in amounts of iron present in these different samples.

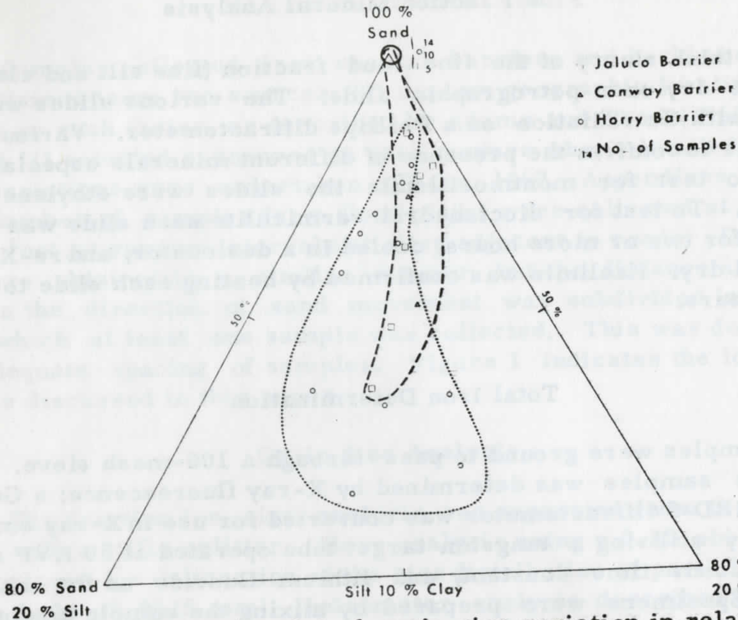

Figure 2. Triangle plot of grain size variation in relation to the different coastal barriers (see Figure 1 for location of samples).

ANALYSIS OF RESULTS

Grain Size Properties

Barrier sediments. A distinct difference in sand-silt-clay ratios contrasts the Recent and Myrtle barrier sediments (100 per cent sand) with those of the Jaluco, Conway and Horry barriers (sand with some clay and silt). While these are all classified as sands according to Shepard's (1954) classification, most of the older sediments contain a considerable accumulation of fines in the sampled horizon. These differences are illustrated in a triangle diagram (Figure 2). The plot shows a progressive increase in fines with age of barrier. If this increase in fines is related to weathering of unstable minerals in the horizon sampled, rather than to the addition of fines to the sampled A_2 horizon from the humic zone above, there should be a strong correlation between abundance of the unstable minerals epidote and hornblende and the value of mean grain size measured in phi units (M phi). Figure 3 illustrates that this is not the case for either of the two minerals. In fact, modern beach sediments have a greater range in abundance of these minerals than all older sediments combined, and they have almost as large a variation in M phi. The scatter plot (Figure 3) shows very little association between the mean size and abundance for either

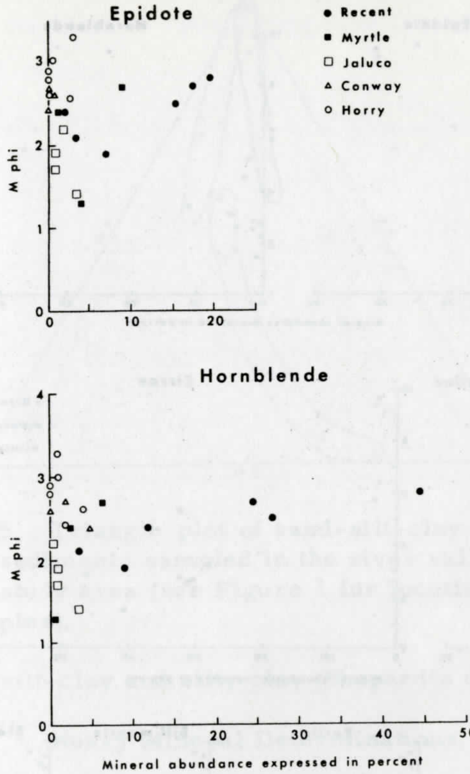

Figure 3. Relationship between M phi of barrier sediments and abundance of epidote and hornblende minerals.

mineral. The plot does suggest a marked decrease in abundance of the two unstable minerals in sediments older than the modern barrier, but lower abundance in older sediments bears no systematic relationship to size as reflected by M phi.

River valley sediments. Figure 4 is a plot of M phi versus mineral abundance for each of the significant mineral species differentiated according to river systems. The positive correlation and decrease in the size of M phi for both epidote and hornblende is striking in its contrast to the negative or poor correlation shown by all other minerals (increase in the value of M phi signifies a finer median diameter). The scatter is large for some mineral species, whereas it is tight for others. This possibly reflects some degree of mixing with local sources of sediment (e. g., by the erosion of coastal barriers).

The plots of sand-silt-clay ratios depicted in Figure 5 show a distinct break in size between the Great Pee Dee, a Piedmont river,

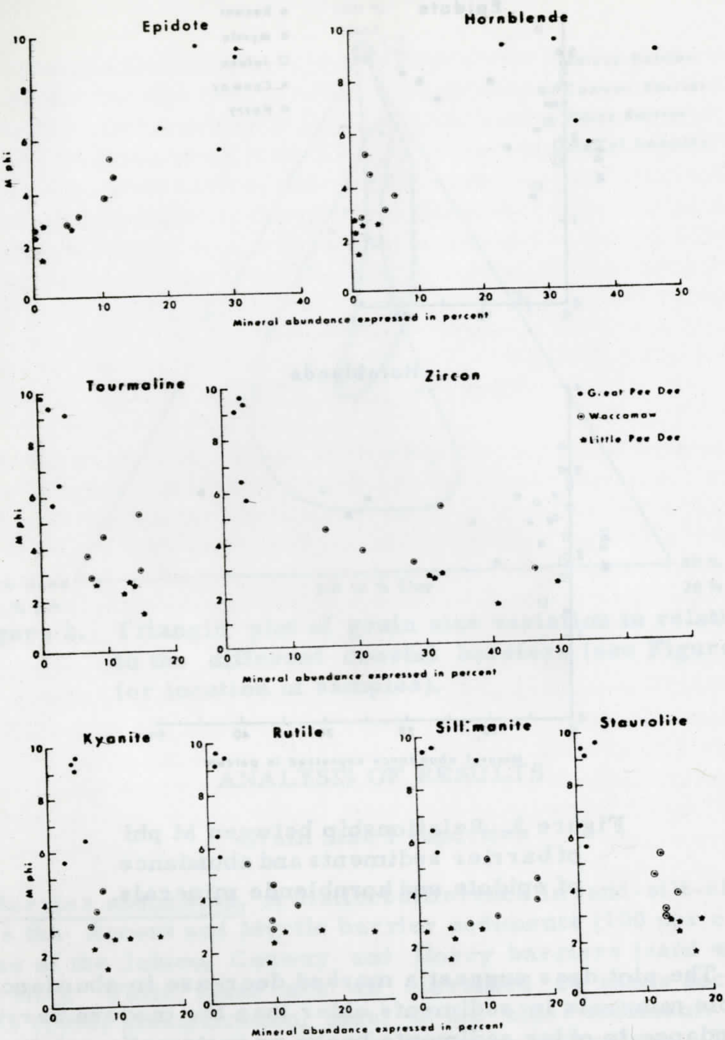

Figure 4. Relationship between M phi and mineral abundance for selected heavy minerals sampled from the river valleys in the study area.

and the two Coastal Plain rivers, Waccamaw and Little Pee Dee. This is also reflected in the values of M phi as plotted in Figure 4. The study by Heron and others (1964, p. 6) showed that little silt or clay is available for transport by Coastal Plain streams in comparison with that from streams with Piedmont sources. This is well demonstrated by the Little Pee Dee sediments that largely consist of sands, clayey sands, and silty sands, whereas sediments in the Great Pee Dee Valley

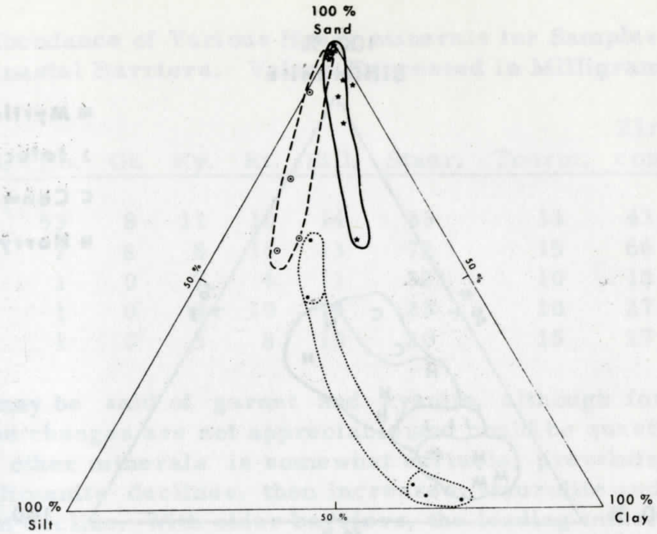

Figure 5. Triangle plot of sand-silt-clay ratios for sediments sampled in the river valleys of the study area (see Figure 1 for location of samples).

are chiefly sand-silt-clay and silty clay (Shepard's classification, 1954).

Heavy Mineral Determinations

Coastal barriers have been considered as a separate problem from the point of view of heavy mineral analysis. Modern beach sediments, as well as Myrtle, Jaluco, Conway, and Horry barriers, have been sampled. Although the numbers of samples are too few for definitive statements, they are sufficient to provide indications as to the value of a heavy mineral study in evaluating sedimentary history.

Mineralogical distinctions and interrelations are not as clear in barrier sediments as in fluvial sediments. It would be expected that the flood plains would have heavy mineral suites somewhat similar to those of the coastal barriers, which are constructed at least in part from the sediment delivered by these rivers. Thus, some smearing or blending effect would not be surprising. This is what the results show to a large degree.

Modern beach berm. The modern beach berm is characterized by a hornblende-zircon-staurolite-epidote suite. Such a suite shows the influence of Piedmont contributions (hornblende-epidote), but the amount of these minerals has already declined considerably as compared to that of the minerals of the river flood plains. This decline may be ascribed to either (a) the diluting effect of such Coastal Plain contributions as staurolite, (b) the chemical decay of these two unstables, or (c) some combination of both.

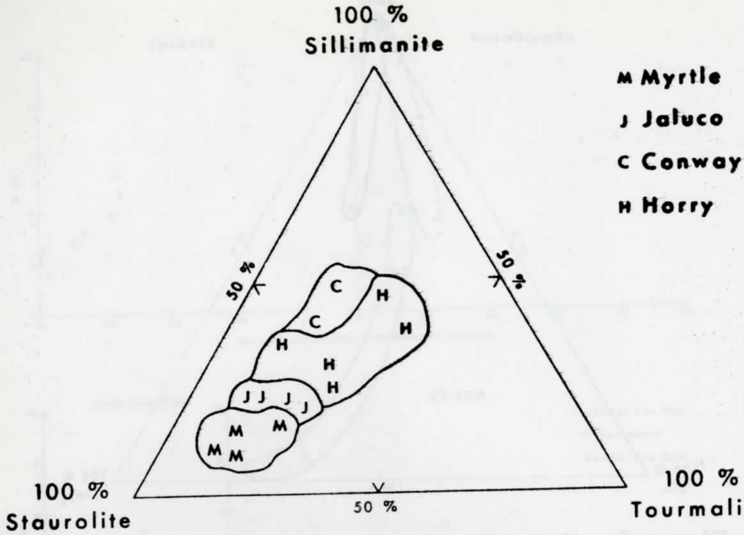

Figure 6. Triangle plot of the minerals tourmaline-staurolite-sillimanite for sediments sampled from the coastal barriers.

Older coastal barriers. The use of triangular diagrams to differentiate each of the four older coastal barriers was not completely successful. In most cases the fields overlapped. One conclusion that might be drawn from this is that the source of mineral supply to the beaches has not changed significantly since Horry time. However, a plot of the minerals tourmaline, staurolite, and sillimanite (Figure 6) on a triangular diagram shows that the barriers might be differentiated on the basis of progressive impoverishment of staurolite with time. Oddly enough, sillimanite, usually regarded as an unstable mineral (especially where it occurs in a fibrous form), is most common in Horry, the oldest of the barriers. The implication that sillimanite is relatively stable is also suggested by the work of Giles and Pilkey (1965). More samples would be necessary in order to determine if the distinctions implied in Figure 6 are valid. Progressive mineralogic changes in barriers from youngest to oldest can be shown by obtaining the average amount of each heavy mineral found in the samples collected from each barrier. Values can be expressed quantitatively in milligrams by multiplying the percentage obtained through counting by the weight of the total heavy-mineral crop. No adjustment for specific gravity was made. The results, shown in Table 1, indicate the approximate weight in milligrams of each heavy mineral in a unit volume of sediment (40 grams). This table shows that the unstable minerals epidote and hornblende decline drastically even in the Myrtle barrier and are virtually absent in still older barriers. It appears that chemical attack on these minerals is the factor responsible rather than any shift in the source of supply.

Table 1. Abundance of Various Heavy Minerals for Samples From the Coastal Barriers. Values Expressed in Milligrams.

	Ep.	Hb.	Gt.	Ky.	Rt.	Sill.	Staur.	Tourm.	Zir-con	Sample Size
Modern	27	53	8	11	10	14	33	11	43	6
Myrtle	8	7	8	8	16	13	72	15	66	4
Jaluco	2	1	0	7	4	11	32	10	18	4
Conway	1	1	0	8	10	14	23	10	27	3
Horry	1	1	0	5	8	19	20	15	27	6

The same may be said of garnet and kyanite, although for these twc minerals the changes are not appreciable and could be questioned. The behavior of other minerals is somewhat variable, preventing generalization. Sillimanite declines, then increases; staurolite and zircon increase, then decline. With older barriers, the leading mineral becomes a more stable variety: hornblende in the modern berm, staurolite in Myrtle and Jaluco, and finally zircon in Conway and Horry. However, the findings for Conway and Jaluco deposits are based on three and four samples respectively, and the Conway samples are not scattered but are localized.

Fluvial sediments. Samples taken for heavy mineral analysis from the flood plains show the following important minerals in each flood plain:

(a) Great Pee Dee: hornblende-epidote-kyanite
(b) Little Pee Dee: zircon-staurolite-tourmaline
(c) Waccamaw: zircon-staurolite-sillimanite

Hornblende and epidote are almost entirely lacking in sediments of the Little Pee Dee, while those of the Waccamaw contain only minor amounts. Aside from this obvious distinction between these rivers and the Great Pee Dee, the three rivers can be fairly well distinguished by minerals other than hornblende and epidote, which are plotted on triangular diagrams. Figure 7 shows a tourmaline-staurolite-sillimanite plot in which the three river flood plains occupy almost completely separate fields. One sample from the Great Pee Dee contains no tourmaline and lies outside all fields. It would appear that the Great Pee Dee is delivering very little staurolite to the Coastal Plain, whereas the other two rivers are handling appreciable quantities of this mineral. Figure 8 shows a sillimanite-kyanite-zircon plot for these same environments. Once again there is a fairly good separation of fields, but the Waccamaw and Little Pee Dee overlap slightly. Almost complete overlap of the Waccamaw and Little Pee Dee fields is apparent when kyanite-staurolite-zircon are plotted (Figure 9). These data indicate that the sediments of three river flood plains can be separated mineralogically. However, those of the Waccamaw and Little Pee Dee have more in common. It can be further noted that the Waccamaw-Little

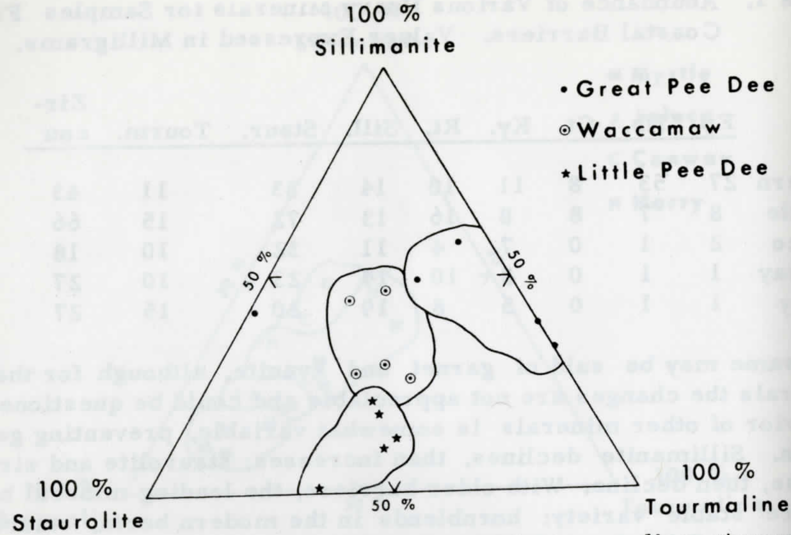

Figure 7. Triangle plot of the minerals tourmaline-staurolite-sillimanite for sediments sampled from the modern flood plains.

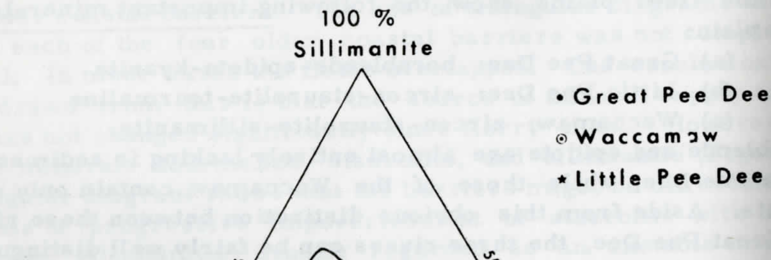

Figure 8. Triangle plot of the minerals sillimanite-kyanite-zircon for sediments sampled from the modern flood plains.

Pee Dee fields in Figures 8 and 9 tend to lie closer to the zircon corner, indicating an enrichment in this stable mineral in comparison to that in

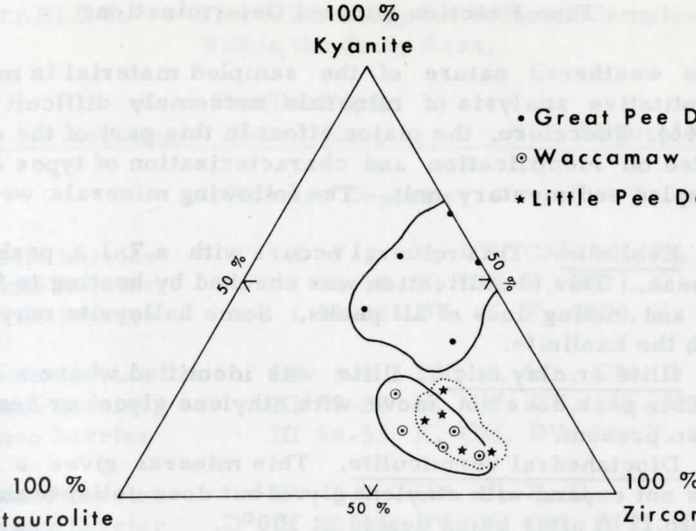

Figure 9. Triangle plot of the minerals kyanite-staurolite-zircon for sediments sampled from the modern flood plains.

less stable minerals, perhaps as a result of reworking of already deposited Coastal Plain sediments by the Waccamaw and Little Pee Dee Rivers. Again these findings corroborate studies by Giles and Pilkey (1965).

Dune sediments. The dune samples from the Great Pee Dee are closely related in their mineral suite to the Great Pee Dee flood plain sediments. A direct comparison of the average number of milligrams of different minerals in flood plain and dunes is shown in Table 2. Although the two suites are closely similar, there is no indication of sorting by wind action. It might be expected that the relatively light mineral hornblende would tend to concentrate to some extent in the dunes. Actually, it drops to second in order of abundance and is quantitatively reduced by 50 per cent from flood plain to dune. On the Isle of Palms hornblende becomes enriched from beach to dune (Neiheisel, 1958).

Table 2. Abundance of Various Heavy Minerals for Samples from the Sand Dunes Within the Great Pee Dee Valley and Adjacent Flood Plain Sediments. Values Expressed in Milligrams.

	Ep.	Hb.	Gt.	Ky.	Rt.	Sill.	Staur.	Tourm.	Zircon	Sample Size
Gr. P. D.	101	141	3	14	5	9	2	8	11	5
Dunes	86	75	5	16	8	11	7	10	17	3

Fine Fraction Mineral Determinations

The weathered nature of the sampled material in most cases made quantitative analysis of minerals extremely difficult (van der Marel, 1966). Therefore, the major effort in this part of the study was concentrated on identification and characterization of types occurring on the sampled sedimentary unit. The following minerals were identified:

1. Kaolinite. This mineral occurs with a 7.1 A peak and with a 3.56 A peak. This identification was checked by heating to 550°C for five hours and noting loss of all peaks. Some halloysite may be associated with the kaolinite.

2. Illite or clay mica. Illite was identified where a 10 A peak occurs. This peak does not move with ethylene glycol or heat. A 5 A peak is also present.

3. Dioctahedral vermiculite. This mineral gives a 14 A peak which does not expand with ethylene glycol but does collapse and "smear out" to 11 to 12 A after being heated at 350°C.

4. Vermiculite. This is the same as the dioctahedral vermiculite except that organic matter has complexed the Al so that it is not interlayered. The result is that it collapses to 10 A with heat (S. Weed, personal communication).

5. Montmorillonite. This mineral expands to about 17 A with ethylene glycol and collapses to 10 A with heat. The basal spacing of the untreated material is usually near 14 A, depending in part on the saturating ion.

6. Mixed layer. This material probably represents some non-random stacking of illite and montmorillonite or perhaps vermiculite.

7. Quartz. This mineral gives 3.35 A and 4.26 A peaks. It is present in all samples.

8. Gibbsite. Although this is a difficult mineral to detect with X-ray diffraction, it was thought to be present when a 4.8 A peak was noted.

9. Feldspar. Peaks at 3.20 A and 3.25 A were assigned to feldspar.

10. Goethite. This is another difficult mineral to detect in relatively low quantities; goethite was identified if a 4.18 A peak developed.

Table 3 summarizes the clay mineral identifications for samples taken in the study area. Dioctahedral vermiculite is the most common of the soil clay minerals of the barrier sediments, commonly occurring with kaolinite. This is not surprising since the mineral is characteristic of most humid soil types such as occur in the southeastern United States (Fiskell et al., 1970). It is an uncommon mineral in unweathered ancient marine sediments of the Coastal Plain (Heron, et al., 1965). Dioctahedral vermiculite does occur in modern river sediments. Clay mica (mostly illite) is most abundant in backbarrier flat

TABLE 3. Clay Mineralogy for Selected Samples Within the Study Area.

Sample No.	Location	Profile No.	Depth (in.)	Clay Minerals
				Barrier Deposits
574	Myrtle barrier	I	7-8	K, DV, CM, ML(?), F, Q
575	Myrtle barrier	I	14-15	K, DV, F, G(?), Q
576	Myrtle barrier	I	32-33	DV, K, F, G(?), V, Q
626	Jaluco barrier	III	3-4	DV, K, ML, CM, Q
627	Jaluco barrier	III	28-29	K, CM, CV, ML, F, Q
628	Jaluco barrier	III	54-55	K, CM, DV, M, F, Q
636	Conway barrier	IV	6-7	DV, K, F, Q
637	Conway barrier	IV	13-14	DV, K, CM, G(?), Q
638	Conway barrier	IV	50-51	DV, K, CM, G, F, Q
639	Conway barrier	IV	61-62	DV, K, CM, F(Tr), G(?), Q
663	Horry barrier	VI	7-8	DV, ML, K, CM, F, G(?), Q
664	Horry barrier	VI	14-15	DV, K, CM, F, Q
665	Horry barrier	VI	32-35	DV, K, CM, F, G(?), Q
667	Horry barrier	VI	71-72	K, DV, CM, ML, F, G(?), Q
668	Horry barrier	VI	94-95	K, CM, DV, ML, Q
				Backbarrier Flat Deposits
356	Tidal marsh, Garden City		0-2	K, CM, V, M, ML, G(?), F, Q
357	Tidal marsh, Garden City		0-2	K, CM, V, M, ML, Q
410	Tidal marsh, Little River		0-2	K, M, V, CM, F, Q
411	Tidal marsh, Little River		0-2	K, CM, V, M, ML, F, Q
577	Myrtle backbarrier	II	5-6	CM, DV, K, ML, G(Tr), F, Q
578	Myrtle backbarrier	II	12-13	CM, K, DV, ML, G(?), F, Q
579	Myrtle backbarrier	II	20-21	K, CM, DV, G(?), Q
580	Myrtle backbarrier	II	30-31	K, CM, ML, F, GTh (Tr), Q
641	Conway backbarrier	V	9-10	K, DV, CM, ML, F, G(?), Q
642	Conway backbarrier	V	20-21	K, CM, F, Q
643	Conway backbarrier	V	56-57	K, CM, ML, F, G, Q
669	Horry backbarrier	VII	3-4	K, DV, CM, F, Q
670	Horry backbarrier	VII	7-8	DV, K, CM, ML, F, Q
671	Horry backbarrier	VII	32-33	K, CM, DV, ML, F, Q
672	Horry backbarrier	VII	53-54	CM, K, GTh (?), Hematite(?), Q

TABLE 3 continued

Sample No.	Location	Profile No.	Depth (in.)	Clay Minerals
673	Horry backbarrier	VII	91-92	CM, K, Q
	Fluvial Deposits (Collected 1 ft. below surface)			
831	Waccamaw R. (point bar)			K, DV, CM, M, Q
494	Waccamaw R. (levee)			K, DV, M, ML, CM, Ch, Q
499	Waccamaw R. (levee)			K, ML, V, CM, F, Q
528	Little Pee Dee R. (levee)			DV, V, K, G(?), Q
535	Little Pee Dee R. (levee)			K, M, DV, ML, O, CM(?)
536	Lumber R. (levee)			DV, K, G(?), CM, F, Q
561	Great Pee Dee R. (levee)			K, DV, CM, M, G(Tr), Q
562	Great Pee Dee R. (levee)			K, DV, CM, ML, G, F, Q
563	Great Pee Dee R. (flood basin)			K, DV, CM, ML, M, Ch(?), G(?), Q
564	Great Pee Dee R. (flood basin)			K, DV, CM, ML, G(?), F, Q
565	Great Pee Dee R. (levee)			K, DV, CM, ML, G(?), F, Q
566	Great Pee Dee R. (levee)			K, DV, CM, ML, G(?), Q
835	Great Pee Dee R. (flood basin)			K, CM, V, ML, G(Tr), Q
553	Terrace II			K, V, CM, M, or ML, F, Q
558	Terrace I (point bar)			DV, K, CM, F, Q

K: kaolite; CM: clay mica (illite); DV: dioctahedral vermiculite; V: vermiculite; ML: mixed layer; G: gibbsite; F: feldspar (?); Q: quartz; M: montmorillonite; and GTh: goethite. Minerals are listed in approximate order of abundance (except for Q). Trace amounts (usually means a small peak) are indicated with Tr.

sediments. Montmorillonite is not common and has probably weathered to kaolinite or dioctahedral vermiculite. Assemblages of montmorillonite-kaolinite and kaolinite-vermiculite-montmorillonite were found in tidal marshes. The river flood plains are characterized by high kaolinite, along with dioctahedral vermiculite and some illite.

Total Iron Determination

Total iron was measured on samples from nine soil profiles in the study area (Table 4). The profiles are located on surfaces of different ages (for location, see Figure 1 and Thom, 1967, Appendix) and on different facies. For barrier profiles (examples I, III, V, VII, VIII)

TABLE 4. Total Iron Determination for Selected Soil Profile Samples Within the Study Area (see Figure 1 for location of profiles I - IX).

Sample No.	Location	Profile No.	Soil Horizon	Depth (in.)	Percentage Iron
			Barrier Deposits		
574	Myrtle barrier	I	A_2	7-8	0.19
575	Myrtle barrier	I	B_1	14-15	0.35
576	Myrtle barrier	I	B_2	32-33	0.35
626	Jaluco barrier	III	A_2	3-4	0.15
627	Jaluco barrier	III	B_1	28-29	0.39
628	Jaluco barrier	III	B_2	54-55	1.28
636	Conway barrier	V	A_2	6-7	0.10
895	Conway barrier	V	B_{11}	8-9	0.29
637	Conway barrier	V	B_{12}	13-14	0.38
896	Conway barrier	V	B_2	31-32	0.70
897	Conway barrier	V	B_3	40-42	1.42
639	Conway barrier	V	C	61-62	0.09
663	Horry barrier (1)	VII	A_2	7-8	0.08
664	Horry barrier	VII	B_1	14-15	0.275
665	Horry barrier	VII	B_2	32-35	0.20
667	Horry barrier	VII	B_3	71-72	5.10
668	Horry barrier	VII	C	94-95	3.20
898	Horry barrier (2)	VIII	A_2	5-7	0.35
899	Horry barrier	VIII	B_1	16-17	0.34
900	Horry barrier	VIII	B_2	20-22	3.52
901	Horry barrier	VIII	B_{31}	25-27	3.52
902	Horry barrier	VIII	B_{32}	34-35	2.20
			Backbarrier Flat Deposits		
577	Myrtle backbarrier	II	A_2	5-6	2.04
905	Myrtle backbarrier	II	B_{11}	11-12	5.40
578	Myrtle backbarrier	II	B_{12}	12-13	6.20
579	Myrtle backbarrier	II	B_2	20-21	7.90
580	Myrtle backbarrier	II	B_3	30-31	20.00
904	Myrtle backbarrier	II	C	51-52	5.40
894	Jaluco backbarrier	IV	A_2	3-4	2.81
903	Jaluco backbarrier	IV	B_2	21-23	6.20

TABLE 4 continued

Sample No.	Location	Profile No.	Soil Horizon	Depth (in.)	Percentage Iron
641	Conway backbarrier	VI	A_2	9-10	4.77
891	Conway backbarrier	VI	B_1	13-14	3.40
892	Conway backbarrier	VI	B_{21}	16-17	6.20
642	Conway backbarrier	VI	B_{22}	20-21	9.00
893	Conway backbarrier	VI	B_{31}	30-31	8.30
643	Conway backbarrier	VI	B_{32}	56-57	12.20
669	Horry backbarrier	IX	A_2	3-4	1.30
670	Horry backbarrier	IX	B_1	7-8	8.50
671	Horry backbarrier	IX	B_2	32-33	7.80
672	Horry backbarrier	IX	B_3	53-54	6.40
673	Horry backbarrier	IX	C	91-92	4.40

samples have been taken in each soil horizon. In the case of more deeply weathered sediments (VII, VIII), the C horizon continued into fine-textured bedding interlayered with barrier sand. In the case of backbarrier flat soil profiles (examples II, IV, VI, IX) the initial fine-textured deposit was possibly more stratified than in barrier sediments.

Age of surface appears to have some influence on iron content in the case of barrier, but not backbarrier, sediments (Table 4). It should be pointed out that the iron content of each sample includes iron from weathered and unweathered minerals and that the availability of these minerals in the primary deposits varies with the facies type. In surficial weathered sediments total iron also varies with the soil horizon. Backbarrier flats have more total iron than barriers, as shown by profiles II and VI. Table 4 shows that the A horizons have less total iron than B horizons.

CONCLUSIONS

1. Little association is apparent between mean size and mineral abundance for coastal barrier sediments. However, strong relationships, particularly for epidote and hornblende, exist for fluvial sediments, reflecting the mixture of fluvial sediments above the study area. Sediments along the coast represent mixing of individual stream contributions. It is possible that as a span of only 2+ phi separates mean grain size in barrier sediments whereas 9 phi variation is present in river sediments, the comparison is to some extent "unfair".

2. The sediments of the coastal barriers can only be grossly

separated on the basis of mean size, whereas those of the Coastal Plain and Piedmont rivers show a distinct separation on the basis of mean size.

3. Epidote and hornblende are dominant transparent heavy minerals in the modern beach berm. In older coastal barriers, more stable heavy minerals such as zircon, staurolite, and sillimanite are dominant, and unstable epidote and hornblende are essentially absent.

4. The Great Pee Dee River carries a suite of hornblende, epidote, and kyanite; the Little Pee Dee and Waccamaw Rivers carry a more stable assemblage of zircon and staurolite. Sediments from the three rivers studied can be distinguished mineralogically when plotted on triangular diagrams.

5. The abundance and distribution of heavy minerals in the Coastal Plain may be of considerable use as accessory information in detailing chronology of coastal constructional features based on relative stability.

6. The modern beach and Myrtle barrier seem to be the most likely areas for beach placer development because of the greater amounts of heavy minerals present.

7. Dioctahedral vermiculite is strikingly the most common clay mineral of barrier soils, illite is most abundant in backbarrier flat sediments, and kaolinite dominates in the river valleys.

8. Total iron appears to vary with age of deposit in the case of barrier sediments but more strikingly with facies type and soil horizon sampled.

9. The variations in textural and mineralogical properties discussed in this paper must be regarded as a first approximation. The study area is large (c. 1500 square miles) and sampling has of necessity been coarse. There is much room for refinement, not only in sampling (in particular, the subdivision of fluvial environments), but also in the mapping of sedimentary and soil units. However, it is apparent that in this area there is some sense to the regional mineralogic framework when consideration is taken of environment of primary deposition, source of sediment, age of deposit, degree of weathering and textural properties.

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VIRGINIA METAMICT MINERALS: EUXENITE AND PRIORITE

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ABSTRACT

Ideally euxenite and priorite are dimorphous forms of $YNbTiO_6$. In natural materials their compositions are more complex because of extensive isomorphism. When metamict they are amorphous and it is necessary to determine their morphology in order to differentiate between them. Heat treatment studies of amorphous metamict specimens show that the crystallization histories of both minerals are essentially identical, with the crystallization of a priorite phase at lower temperatures and a euxenite phase at higher temperatures. Indexed X-ray diffraction powder data are given for both of these phases. These may be accompanied by betafite-like, brannerite-like, and/or rutile phases. The temperatures of crystallization and the exact products formed vary some, and apparently depend upon the chemical composition and degree of metamictization of the individual specimens. Descriptions are given of specimens from several localities in Amelia, Bedford, and Powhatan Counties, Virginia.

INTRODUCTION

Although the chemical similarities between priorite and euxenite have been recognized for many years (Adamson, 1942), only recently it has been shown they are dimorphs of $YNbTiO_6$ (Komkov, 1963, 1964). The fact that both minerals are usually metamict thwarted this conclusion. Because metamict specimens of both minerals behave almost identically upon crystallization when heated in air, the morphology of a specimen is needed in order to identify the original mineral. The main purposes of this paper are to summarize the X-ray diffraction data for both dimorphs, to present the heat treatment studies of metamict specimens, and to briefly describe the occurrences of the material in Virginia.

Acknowledgments

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CHEMICAL DATA

Komkov (1963, 1964) showed that euxenite and priorite are the high (α) and low (β) temperature phases, respectively, of $YNbTiO_6$. Analyses of natural specimens depart from this ideal formula because of extensive isomorphism (Palache and others, 1944, p. 789, 795; Soboleva and Pudovkina, 1961, p. 351, 363; Vlasov, 1966, p. 472; George, 1949, p. 53, 57; Adademiia Nauk S. S. S. R., 1967, p. 351, 374). Compilations of many analyses show that in the general formula AB_2X_6, A = Y, Dy, Er, Yb, U, Th, Ca, Ce, Fe^2 (with very minor Mg, Mn, Pb, Na, K, Bi, Sc), B = Nb, Ti (with minor Ta, Fe^3, Al, Sn, Zr), and X = O (with minor OH). Studies of analyses have led some investigators (Soboleva and Pudovkina, 1961, p. 353) to conclude that euxenite contains more U than Th, while on the other hand priorite contains more Th than U. Komkov (1963, 1964), however, showed that this is not always true, and indicated the correlation is fortuitous. Much earlier Adamson (1942) expressed the opinion that euxenite and priorite are chemically identical and dimorphous. This idea was supported later by George (1949, p. 56), Aleksandrov and Pyatenko (1959), and Vlasov (1966, p. 470). Because specimens of both minerals are practically always metamict, a knowledge of the original morphology of a specimen is essential for its identification. The morphological data for the two minerals, given by Palache and others (1944, p. 788, 794), and Akademiia Nauk S. S. S. R. (1967, p. 349, 372), closely correspond to the structural data determined by Komkov (1959) and Komkov and others (1964).

Table 1 gives semiquantitative spectrographic analyses of two Virginian specimens. The results compare well with the chemical data for euxenite-priorite. The specimen from the Rutherford pegmatite, Amelia County, in which Th is greater than U, has the euxenite morphology. The one from the Nance pegmatite, Bedford County, is anhedral.

X-RAY DATA FOR EUXENITE AND PRIORITE

Indexed X-ray diffraction powder data for euxenite (heated to 1000° C) were published first by Arnott (1950). His orthorhombic cell, in which $\underline{a}$ = 5.52 A, $\underline{b}$ = 14.57 A, $\underline{c}$ = 5.17 A, is compatible with the morphological data. He found Z = 4, and considered the most probable space group to be Pcmn (or Pc2n, if hemimorphic). Subsequently unmetamict euxenite crystals reported by Nefedov (1956) were shown to have a cell of similar dimensions. Later Seifert and Beck (1960, 1961)

Table 1. Semiquantitative Spectrographic Data on Euxenite - Priorite from Virginia. Elements Reported as Oxides. Analyst: F. W. Barley, American Spectrographic Laboratories, San Francisco.

Element	Nance pegmatite, Bedford County WDR#77	Rutherford pegmatite, Amelia County V4393
Na	.75 %	<.5 %
Mg	.2	.1
Al	.75	.4
Si	2.5	.85
K	3.	-
Ca	.75	1.5
Ti	12.	10.
V	<.05	<.05
Mn	.2	.15
Fe	2.5	2.
Y	15.	12.
Zr	<.15	<.15
Nb	PC[a]	PC[a]
Mo	<.05	.1
Sn	.05	.04
Ba	.03	<.01
La	-	.5
Ce	.25	1.5
Nd	.15	.7
Eu	<.05	.05
Gd	.15	.1
Dy	2.	.75
Er	.1	<.1
Yb	.5	.1
Ta	1.	6.
Th	<1.	1.5
U	2.5	.3

Pr. Sm. Cr. Cu. Sr. not determinable, interference

[a] PC, principal constituent.

and Beck (1961), who showed synthetic $YNbTiO_6$ is closely related to euxenite, verified the cell, but determined the space group to be Pcan. Independently Komkov (1963), from a study of a series of synthetic rare earth compounds of the form $TRNbTiO_6$, arrived at the same conclusions. Indexed X-ray powder data for euxenite, based on this structural cell, have been published by Komkov (1959), Sokolova (1959), Aleksandrov and Pyatenko (1959), Hutton (1961), Gorzhevskaya and Sidorenko (1963, 1964), Lima-de-Faria (1964), Vlasov (1966, p. 479), Akademiia Nauk S. S. S. R. (1967, p. 354), and Korentova and others (1968).

The calculated interplanar specings for euxenite in Table 2 are based upon the unit cell for synthetic α -$YNbTiO_6$ in which $\underline{a}$ = 5.59 A, $\underline{b}$ = 14.65 A, $\underline{c}$ = 5.19 A, $\underline{a}:\underline{b}:\underline{c}$ = 0.382:1:0.354 (Komkov, 1959), Pcan. All possible reflections through 1.59 A are included. The measured data are from heat-treated euxenite from the Nance pegmatite, Bedford County, and represent averaged values from four films made in two cameras of 11.46 cm diameter with filtered copper radiation.

Komkov (1959) published indexed X-ray diffraction data for priorite (blomstrandine), and subsequently Komkov and others (1962, 1964) further verified the data from rare earth compounds of the form $TRNbTiO_6$, synthesized hydrothermally in the temperature range 250°-400° C. For β-$YNbTiO_6$ they reported a cell in which $\underline{a}$ = 5.185 A, $\underline{b}$ = 10.96 A, $\underline{c}$ = 7.415 A, $\underline{a}:\underline{b}:\underline{c}$ = 0.473:1:0.677, with space group Pbnm, Z = 4. Indexed X-ray diffraction data, based on this cell, have been published by Gorzhevskaya and Sidorenko (1963, 1964), Komkov and others (1964), Lima-de-Faria (1964), Vlasov (1966, p. 479), and Adademiia Nauk S. S. S. R. (1967, p. 376).

The calculated interplanar spacings for priorite in Table 3 were derived from the unit cell for synthetic β-$YNbTiO_6$ (Komkov and others, 1964) mentioned above. All possible reflections through 1.67 A are listed. The measured data for priorite in Table 3 are averaged values from four films made in two cameras of 11.46 cm diameter, using filtered copper radiation. Some of the more intense reflections in the priorite measured data can be assigned cubic indices and possibly indicate a betafite-like structure with $\underline{a}$ = 10.42 ± 0.01 A (Table 3). This is probably only a coincidence since the spacings for the reflections also correspond to priorite data, and a definite cubic phase with an $\underline{a}$ parameter this large has not been observed in heated euxenite-priorite before. The heat-treated specimens are from Amelia County, Virginia.

Unindexed data for euxenite and priorite, published by Berman (1955), Lima-de-Faria (1958, 1964), Soboleva and Pudovkina (1961, p. 361), and Heinrich (1965), now can be identified and indexed with certainty.

LITERATURE REVIEW OF THE HEAT TREATMENT STUDIES OF EUXENITE AND PRIORITE

In the following review there is some uncertainty regarding the identity of the metamict priorite or euxenite studied, because the morphologies of the specimens were not determined by all of the authors cited. The crystallization histories of both minerals when metamict are similar upon heating. Usually priorite forms at lower temperatures

Table 2. X-ray Powder Diffraction Data for Euxenite from Bedford County, Virginia. Filtered Copper Radiation.

hkl	d(calc.) A	d(meas.) A	I(obs.)	hkl	d(calc.) A	d(meas.) A	I(obs.)
020	7.33	7.26	w	052	1.94	1.93	vw
110	5.22	5.18	vvw	202	1.90	1.90	w
111, 130	3.68			212	1.89		
040	3.66	3.66	m	251	1.88		
121	3.38	3.36	w	310	1.85		
131	3.00	2.97	vs	222, 260, 152	1.84		
200	2.79	2.78	mw	171, 080	1.83	1.83	w
141	2.64			062	1.78		
220	2.61			232	1.77	1.77	mw
150, 002	2.60	2.60	w	311, 330	1.74		
012	2.56	2.56	mw	261	1.73	1.73	mw
201	2.46			321	1.71		
022	2.45	2.45	vw	162	1.70		
060	2.44			242	1.69	1.68	vvw
211	2.43	2.42	w	331, 181	1.65		
221	2.33	2.36	vvw	113	1.64	1.64	w
112, 151	2.32			072	1.63		
032	2.29	2.30	vw	123	1.61	1.61	vw
122	2.24			252	1.60		
240	2.22	2.21	vw	271	1.59	1.59	vvw
231	2.20	2.19	vw			1.56	vw
132, 042	2.12	2.11	w			1.49	w
161	2.06					1.48	vvw
241	2.04	2.03	vvw-			1.46	vw
142	1.98					1.44	vw
170	1.96	1.97	vw				

and euxenite at higher temperatures. Other phases formed by the decomposition of the minerals are a cubic betafite-like phase, a brannerite-like phase, and a rutile phase.

Arnott (1950) obtained a consistent euxenite pattern at 1000° C. At lower temperatures variations in X-ray patterns were frequent. An examination of a powder photograph published in his paper shows that he obtained the priorite phase at lower temperatures, a fact unknown to him. Komkov (1959) reported that either euxenite or priorite, along with a cubic phase, form in the range 450°-600° C, depending upon the original structure of the mineral and if the mineral is completely metamict or not. In the range 800°-1200° C all specimens crystallized to form a mixture of euxenite, a cubic phase, and rutile. Later, in studies of synthetic crystals, Komkov and others (1962) found that priorite

Table 3. X-ray Powder Diffraction Data for Priorite from Amelia County, Virginia. Filtered Copper Radiation.

hkl	d(calc.) A	d(meas.) A	I(obs.)	hkl	d(calc.) A	d(meas.) A	I(obs.)
020	5.48	5.50	vw	150	2.02		
110	4.69	4.74	m	222	1.98	1.99	mw
021	4.41	4.42	w	151	1.95	1.96	w
101	4.25	4.29	vw	133	1.90	1.91	w+
111	3.96	3.99	w	240	1.88		
120	3.77	3.80	vw	004	1.85	1.86	m-
002	3.71	3.71	ms	232,043	1.84	1.84	mw*
121	3.36	3.39	w	060,241	1.83		
022	3.07	3.08	w+	061,152,213	1.77	1.78	mw
130	2.99	3.01	vvs*	024	1.76	1.75	vw
112	2.91	2.92	vs	143	1.73	1.74	vw
131	2.77	2.80	w	114,160	1.72		
040	2.74	2.75	w	310	1.71	1.71	vvw
122	2.64	2.63	m*	223	1.70		
200	2.59			301,242,161	1.68	1.69	m
041	2.57	2.57	mw	250	1.67	1.67	vvw
210	2.52					1.65	vvw
140	2.42	2.43	w			1.62	vvw
211	2.39					1.58	ms
220	2.34	2.34	mw			1.57	mw*
132	2.33					1.55	vw
141	2.30					1.54	w
023	2.25	2.25	w			1.52	w+
221	2.24					1.50	mw*
103	2.23					1.48	vw
042	2.20	2.21	mw			1.45	vw
113	2.19					1.39	mw-
202	2.13	2.14	vw			1.35	vvw
230	2.11	2.11	vvw-			1.33	vw
212	2.09					1.31	vvw*
123	2.07	2.06	vw			1.29	vvw
231,142	2.03	2.03	vw			1.27	vvw

*Possible interference from a cubic betafite-like phase with a = 10.42 $\pm$ 0.01 A.

structures could be synthesized hydrothermally in the range 250°-400° C, and that these upon heating in air might be stable as high as 1000°C. Crystals synthesized at higher temperatures (Komkov, 1963) possessed the euxenite structure. Seifert and Beck (1961) showed that euxenite can form at temperatures as low as 650°-700° C under hydrothermal conditions. In heat treatment studies of metamict specimens they (Seifert and Beck, 1960, 1961, 1965; Beck, 1960) found a "low temperature form" above 400°-450° C, later recognized as priorite (Seifert and others, 1964), and euxenite above 750°-800° C. Gorzhevskaya and Sidorenko (1963, 1964, 1966) discussed the heat treatment of priorite and euxenite as separate metamict minerals. For priorite crystallization began at about 500° C and was the major phase through 900° C. A cubic phase with $\underline{a}$ from 10.11 to 10.15 A was present from 800°-900° C. At 900° C the euxenite phase, accompanied by a cubic phase with $\underline{a}$ = 10.16 to 10.22 A, began to form. For euxenite specimens crystallization began at about 600° C with the formation of a cubic phase with $\underline{a}$ = 10.09 to 10.16 A. At 800° C euxenite formed, accompanied by a cubic phase with $\underline{a}$ = 10.16 to 10.24 A. The presence of a priorite structure in heated euxenite was not mentioned.

Although the phases were not identified, Berman (1955) also recognized that X-ray data for specimens heated at low temperatures were different from data for specimens heated at higher temperatures. Aleksandrov and Pyatenko (1959) found that when euxenite and priorite (blomstandine) were heated to 1100° C they both yielded a euxenite phase accompanied by a cubic phase. Hutton (1961) found euxenite to form as low as 620° C. He did not observe the priorite phase, but did notice a cubic phase. Heat treatment studies of euxenite by Lima-de-Faria (1958, 1964) yielded a combination of euxenite and cubic phases, and rarely an $\underline{e}$ phase. For heated priorite he observed a combination of a cubic phase and the $\underline{e}$ phase. He suggested that the $\underline{e}$ phase is a disordered form while euxenite is an ordered form of the same compound. This $\underline{e}$ phase is now recognized to be identical to the priorite structure. He observed it to be most common in samples heated at lower temperatures (700° C). Vlasov (1966, p. 475) pointed out that priorite and euxenite behave in like manner when heat-treated and can not be distinguished by X-ray analysis. In the range 500°-900° C priorite, or a cubic phase, or both, crystallized from the metamict specimens. Above 900° C the euxenite phase formed often accompanied by a cubic phase. The formation of euxenite and priorite phases from the heat treatment of euxenite was acknowledged by Schröcke (1966). For Fauquier (1968) euxenite formed after treatment as low as 300° C. A cubic phase and unknown phases accompanied euxenite at higher temperatures. Priorite was not identified. Korentova and others (1968) found the cubic phase accompanied by euxenite in the range 800°-850° C, and pure euxenite at 1100° C. Van Wambeke (1970), in studies of altered and fresh euxenite and priorite, found that sometimes rutile, brannerite, UNb_2O_8, and cubic phases accompanied the major euxenite phase after heating in air at 1000° C.

The temperatures at which metamict euxenite and priorite begin to recrystallize vary over a wide range, and apparently depend on the chemical composition of the specimen, the degree of metamictization, and whether the original mineral was euxenite or priorite. Gorzhevskaya and Sidorenko (1963, 1964) reported that metamict euxenite began to crystallize at about 600° C while metamict priorite began at about 500° C. The crystallization of euxenite at 750° C was determined by Chudoba and Lange (1949). In differential thermal analyses different values for exothermic peaks, presumably representing crystallization of the metamict materials, have been reported by Kerr and Holland (1951), Kurath (1957), Sokolova (1959), Adler and Puig (1961), Soboleva and Pudovkina (1961, p. 350, 360). Gorzhevskaya and Sidorenko (1964), Seifert and Beck (1965), Vlasov (1966, p. 478), and Adademiia Nauk S. S. S. R. (1967, p. 353, 375). For euxenite these investigators frequently found a strong exothermic peak around 750° C, but this peak ranged from less than 700° to 780° C. For priorite they reported an exothermic peak in the range 470° to 540° C. In this current study of specimens from Virginia a very weak priorite X-ray pattern was detected after some specimens were heated in air at 200° C (one hour).

HEAT TREATMENT OF SPECIMENS FROM VIRGINIA

Unheated specimens from Virginia are either completely metamict or show a very weak broad reflection at approximately 3 A. Unfortunately this corresponds to the strongest reflection for euxenite, priorite, and the betafite-like phase, each of which may form when the specimens are heated. Heat treatment studies of numerous specimens show results which fall roughly into two possible series, generally corresponding to two geographical areas, the Amelia-Powhatan Counties area and the Bedford County area.

(a) Amelia-Powhatan Counties area. The priorite phase generally dominates heat-treated specimens from pegmatites of this area. After heating in air for one hour each, the following results were noticed: 200° C, a very weak priorite phase; 400° C, weak priorite; 600° C, medium weak priorite (after six hours the X-ray pattern was much improved); 700° C, medium priorite; 800° C, strong priorite plus weak euxenite (in one sample, betafite-like phase instead of euxenite); 900° C, strong priorite plus euxenite; 1000° C, either pure priorite, pure euxenite, or a mixture of the two in which priorite usually dominates (additional rutile at times). Heating for a longer period of time at 1000° C did not noticeably affect the ratio in these mixtures but the sharpness of the X-ray lines was enhanced.

(b) Bedford County area. The euxenite and betafite-like phases generally characterize heat-treated specimens from this area. After heating in air the following results were noticed: 600°-750° C (one to six hours), dominant betafite-like phase plus definite, but weaker, priorite phase; 800° C (one hour), euxenite plus less important priorite and betafite-like phases; 900° C (one hour), either pure euxenite, or euxenite with very minor betafite-like, priorite, rutile, and/or brannerite-like phases; 1000° C (one to six hours), well-developed euxenite, rarely with very minor impurities, e. g., rutile.

The cubic phase which forms when some specimens are heated is referred to here as a betafite-like phase, rather than pyrochlore phase, because the intensities of its X-ray diffraction reflections correspond more closely to those reported for betafite by Hogarth (1961). The data are close to those found in heated samarskite by Mitchell (1970), with reflections for 222, 400, 440, 622, 444, 800, 622, 840. However the lattice parameter $\underline{a}$ determined in the present study is smaller, varying from 10.15 ± 0.01 to 10.21 ± 0.01 A.

Table 1, which shows analyses of typical specimens from each of the two areas, indicates some differences in composition. For example, the priorite trending specimen from the Rutherford pegmatite (Amelia County) contains more Th than U, while the euxenite trending specimen from the Nance pegmatite (Bedford County) has more U than Th. That specimens from one area were originally priorite, while those from the other were euxenite is probably not the case. The euhedral specimen (#V4393) from the Rutherford pegmatite was shown to have a definite euxenite morphology, even though the priorite pattern persists for it in heat-treated samples.

VIRGINIA SPECIMEN OCCURRENCES

Occurrences of euxenite-priorite in Virginia have not been verified until now. Jahns and others (1952, p. 31) indicated that euxenite rarely occurs in the Amelia district and in the west-central part of the Virginia Piedmont, but they did not specify the pegmatites in which the mineral was found. The euxenite described by Dietrich (1961) from the Mitchell pegmatite, Bedford County, was shown to be samarskite by Mitchell (1970). Euxenite mentioned by Heinrich (1962) from Prince Edward County was shown later to be pyrochlore-microlite (Mitchell and Zulkiewicz, 1970). X-ray diffraction studies of metamict minerals from Virginia have verified the occurrence of specimens in Amelia, Bedford, and Powhatan Counties in seven definite pegmatites. With a lack of morphological data for specimens from Virginia, it is still problematic which were originally euxenite or originally priorite. The only specimen which could be determined with certainty was euxenite from the Rutherford pegmatite, Amelia County. For convenience the euxenite-priorite dimorphs simply will be referred to as euxenite in the following discussion.

Amelia County, Morefield Pegmatite

The only euxenite known from this deposit is a specimen measuring 15 by 22 by 30 mm in the U. S. National Museum (USNM no. 105564). It is dark brownish-black with a brilliant vitreous luster. The specimen was originally labelled "samarskite (?)" (Dietrich, 1963, p.

13). X-ray diffraction studies of a fragment loaned by the National Museum show a persistent priorite phase even at 1000° C, a situation typical for specimens from Amelia County. The mineral has not been reported previously from this deposit. Glass (1935) described the mineralogy of the pegmatite; and Lemke and others (1952, p. 127) described the geology.

Amelia County, John Patterson Pegmatite

Splendent subhedral dark brownish-black euxenite fragments (less than 5 mm across) were found in a dump rock at the John Patterson pegmatite. These crystals are embedded in grayish-white plagioclase along with muscovite and quartz. Although X-ray studies of specimens heated at lower temperatures showed a pure priorite phase, euxenite alone formed at 1000° C (one hour). Lemke and others (1952, p. 120) have described the pegmatite.

Amelia County, Rutherford Pegmatite

Two specimens of euxenite are known from the Rutherford pegmatite. One of these, collected by Sandra Knowlton, a student at the University of Virginia, is a brownish-black subhedral piece (3 mm across) embedded in a muscovite crystal. The second one is a large (23 by 12 by 8 mm) brownish-black euhedral crystal (#V4393) found by R. J. Bland, Jr. of Richmond.

The morphology of this large crystal was determined by using a one-circle reflecting goniometer. To heighten the reflectivity of some of the faces they were coated with a mixture of glue and water. Although the angular measurements were not extremely accurate they were close enough to show the crystal has the morphology of euxenite. The major form is {110} followed by {160}, {100} and {120} (?). Form {160} was first recognized on euxenite relatively recently by Sokolova (1959), and {120} is not in the literature. Attempts were made to see if the morphology could correspond to that of priorite. A good correlation was not found. Because priorite $\underline{a}$ is near euxenite $\underline{c}$, priorite $\underline{b}$ is near euxenite $\underline{a}$ x 2, and priorite $\underline{c}$ is near euxenite $\underline{b}$ x 1/2, the euxenite {110} is similar to a proposed {041} priorite form. This form has never been reported for priorite, and yet it is the major form on the crystal. Euxenite {160} is close to a possible priorite {023}, but this is uncommon for priorite. No correlation was found between the crystal measurements and other priorite orientations.

For the two specimens from this locality the priorite phase dominates at all temperatures of treatment. Minor amounts of euxenite are present in samples heated at 800° C and above, except in one sample heated six hours at 1000° C which yielded pure priorite.

The complex mineralogy of this deposit was described by Glass (1935). The geology was reviewed by Lemke and others (1952, p. 121).

Amelia County, Carl Wagner Property

Dark brownish-black subhedral fragments of euxenite were found as float by W. D. Baltzley, formerly of Amelia, on the property of Carl Wagner. This farm is north of Amelia on State Road 630, about 1.5 miles west of the intersection of State Road 630 with State Road 609. X-ray studies of heat-treated samples showed a persistent pure priorite phase.

Amelia County, Unspecified Locality

A very dark reddish-brown fragment of euxenite, from an unspecified pegmatite in Amelia County, was supplied for this study by W. D. Baltzley, formerly of Amelia. The fragment measures 10 by 7 by 5 mm. The priorite phase persists at all temperatures. A minor betafite-like phase ($\underline{a}$ = 10.21 $\pm$ 0.01 A) formed at 800° (one hour), and very minor euxenite and rutile formed at 1000° (one hour).

Bedford County, Mitchell Pegmatite

Tiny rectangular euxenite crystals, rarely over 3 mm long, are embedded, with muscovite and pink garnet, in gray plagioclase at the Mitchell pegmatite. The mineral is submetallic and black. X-ray studies showed a definite priorite phase at 750° C (one hour) and pure euxenite at 1000° C (one hour). The large "euxenite" crystals from this pegmatite, described by Dietrich (1961, p. 12), were shown by Mitchell (1970) to be intergrowths of samarskite and ferrocolumbite. Associated minerals have been listed by Mitchell (1966); and Brown (1962, p. 176) has discussed the geology of the deposit.

Bedford County, Nance Pegmatite

According to Riesmeyer (W. D. Riesmeyer, unpublished manuscript, 1969) euxenite is an abundant accessory mineral at the Nance pegmatite in rocks composed of quartz, plagioclase, muscovite, and garnet. It occurs as dark brown to honey-yellow rounded masses or elongated rough prisms and blades. In size this vitreous mineral averages 1/4 inch across. One rough crystal measures 25 by 8 by 7 mm. Most specimens are fractured and contain secondary thorogummite which is mixed at times with minor uranophane. Euxenite specimens from this pegmatite show typical X-ray diffraction data for the Bedford County area, summarized above, with betafite-like and priorite phases through 750° C, and euxenite the dominant phase at 800° C and above.

Riesmeyer (unpublished manuscript, 1969), who made an X-ray diffraction study of the rare minerals in the pegmatite, found monazite, rhabdophane (after allanite?), bastnaesite, thulite, clinozoisite, apatite, zircon, and pyrite, in addition to the euxenite and its associated thorogummite and uranophane. The pegmatite was described by Brown (1962, p. 177).

Bedford County, Young Pegmatite

Dark brown subhedral crystals of euxenite, less than 5 mm across, occur in rocks composed of white plagioclase, quartz, muscovite, and garnet at the Young pegmatite. In the matrix rock brown halos commonly surround euxenite. X-ray studies showed priorite and betafite-like phases at lower temperatures, and euxenite at 1000° C (one hour).

In the deposit the writer also found bastnaesite (after allanite?), pyrite, and clear smoky quartz masses. The rutile mentioned by Pegau (1932, p. 87) is possibly euxenite. Brown (1962, p. 179) and Griffitts and others (1953, p. 186) have described the pegmatite.

Powhatan County, White Peak No. 1 Pegmatite

Small fragments of somewhat altered euxenite were collected from weathered dump materials at the White Peak No. 1 pegmatite. Dark brown vitreous pieces contain zones, especially along fractures, which are waxy and yellowish. When heat-treated this euxenite showed an essentially pure priorite phase through 800° C (one hour). At 1000° C (one hour) the euxenite phase somewhat exceeded priorite in amount. Unheated earthy alteration crusts gave X-ray data for a betafite-like phase with $\underline{a} = 10.49 \pm 0.01$ A. Studies by others (Vlasov, 1966) of similar crusts on euxenite from other localities showed rutile and anatase.

Samarskite (Mitchell, 1970) and a mineral which maybe be an altered uranium-rich samarskite (Mitchell, 1965) have been found with the euxenite. The pegmatite was described by Brown (1962, p. 96) and Griffitts and others (1953, p. 178).

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