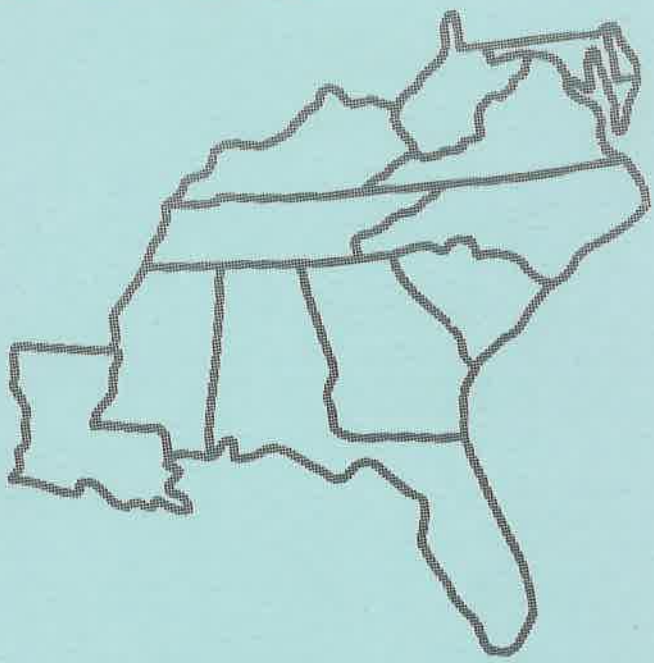
SOUTHEASTERN GEOLOGY



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SOUTHEASTERN GEOLOGY

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THE GEOLOGY OF THE NORTHERN PORTION OF THE WADESBORO TRIASSIC BASIN, NORTH CAROLINA

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ABSTRACT

Clastic sediments representing alluvial fan and other fluvial deposts were mapped in the northern portion of the Wadesboro Triassic basin. Complex faulting and fanglomerate deposits occur along the northwestern border of this basin, and a major normal fault forms the southeastern border. Post-depositional movement along this fault has given the Triassic beds a southeasterly dip. The southeastern border generally lacks a coarse facies but the presence of fanglomerate-filled Triassic outliers lying several miles to the southeast suggest the basin was once wider than it is today. Tectonism has apparently had a great influence on sedimentation. The Deep River basin, separated from the Wadesboro basin by surficial Coastal Plain deposits, contains a middle unit (Cumnock Formation) composed of dark shales and coal beds. Field evidence suggests the absence of the coal-bearing unit in the Wadesboro basin is probably related to cross faulting along the Pekin cross structure. This structure probably originated as an anticlinal warp in the basin floor but cross faulting has subsequently given the feature a fault block appearance. Post-Cumnock uplift and erosion may have removed 5,000 feet of Triassic sediment including the Cumnock Formation from this portion of the Wadesboro basin.

INTRODUCTION

General Statement

The Wadesboro basin extends from Chesterfield County, South Carolina to the village of Candor, Montgomery County, North Carolina (Figures 1 & 2). It is located in the Piedmont Physiographic Province and forms a gently rolling topographic lowland that ranges from 50 to 200 feet lower than the surrounding Piedmont and Coastal Plain surfaces.

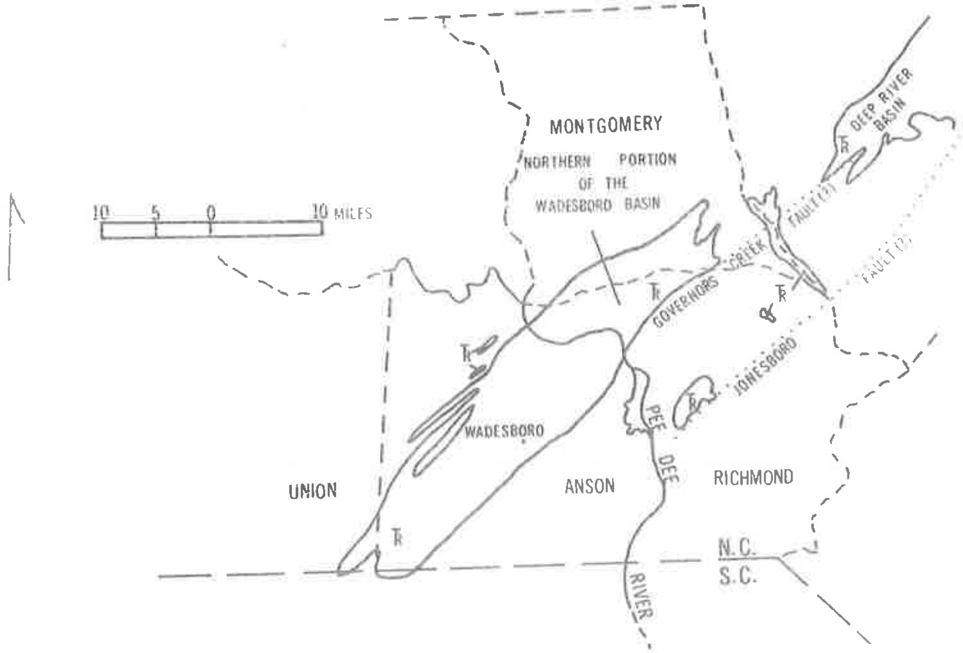

Figure 1. Index map of the Wadesboro Triassic basin showing outcropping Triassic areas and the possible extensions of the Jonesboro and Governors Creek faults (after Randazzo and others, 1970).

A prominent northwest facing escarpment (Pea Ridge) is present along the southeastern boundary.

The northern portion of the Wadesboro basin is the area lying north of the Pee Dee River and south of the sands of the Coastal Plain overlap which completely covers the basin near Candor. These sands, Cretaceous or younger in age, separate the Wadesboro basin from the Deep River basin which lies to the northeast (Figure 2).

Regional Geology

The Triassic sediments of the Wadesboro basin lie unconformably over and are faulted against Lower Paleozoic metamorphic rocks of the Carolina Slate Belt. To the southwest the basin terminates in South Carolina by a series of cross faults. The Crowburg basin in South Carolina (Bell and others, 1974) may be the southernmost outlier of the Wadesboro basin. Structural splintering along the northwestern border of the Wadesboro basin has developed a series of horsts and fanglomerate-filled graben. In the southeast the basin is bordered by a major fault dipping steeply to the northwest. This is possibly an extension of the Governors Creek fault (Randazzo and others, 1970, p.

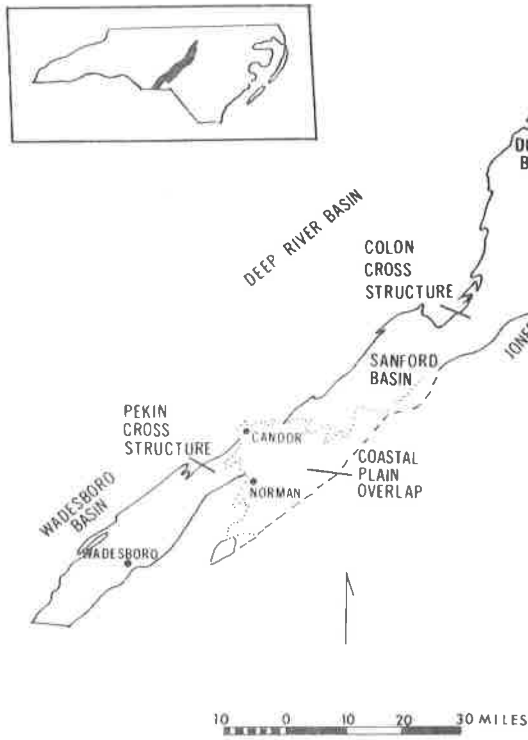

Figure 2. Map of the Deep River-Wadesboro basin showing the location of the five structural units: Durham basin, Colon cross structure, Sanford basin, Pekin cross structure and the Wadesboro basin (after Zablocki, 1959).

1002) which lies within the Deep River basin to the northeast, or a branch of the Jonesboro fault which forms the southeastern border of the Deep River basin (Figure 1).

Southeast of the Wadesboro basin a series of Triassic outliers occur. Some of the outliers contain fanglomerates and conglomerates and some are terminated by an extension of the Jonesboro fault (Zablocki, 1959, p. 38-41).

Pre-Triassic rocks in the Wadesboro area include a series of argillites and tuffaceous rocks of the Carolina Slate Belt (Randazzo, 1972), a granite batholith and associated contact aureole (Waskom and Butler, 1971), a thin belt of gneiss and schist southeast of the basin and numerous quartz veins which have intruded the bedded argillites. Triassic sedimentary rocks within the basin are composed of a thick sequence of nonmarine clastic sediments consisting of varicolored clays, red, brown and maroon claystones, siltstones, shales, sandstones, conglomerates and fanglomerates. Gray and black shales and coal are found in the Deep River basin but not in the Wadesboro basin. Triassic igneous rocks are dark gray and black dolerite which have intruded Triassic sedimentary rocks and the pre-Triassic lithologies principally as dikes. These intrusions vary in width from a few inches to over 300 feet and in length from a few feet to several miles. Gravel, coarse sand, claystones, siltstones and recent alluvium make up the post-Triassic sedimentary deposits of the area.

The Wadesboro basin south of the Pee Dee River was mapped by Swe (1963) and Randazzo (1965). Their works were integrated and expanded to include emphasis on the role of Triassic tectonism within the Wadesboro basin (Randazzo and others, 1970). Our paper focuses attention on the northern portion of the Wadesboro basin.

Acknowledgments

We wish to thank Walter H. Wheeler and J. Robert Butler for reviewing the original manuscript and making many helpful suggestions for its improvement. Paul A. Thayer generously provided us with a number of thin sections of sandstones from the Wadesboro basin. We are grateful for the encouragement and financial support received from the N. C. Department of Natural & Economic Resources, in particular Stephen Conrad and Eldon Allen. Ed Burt provided additional field information which has been included in Plate 1. The Society of the Sigma Xi provided additional funds for this study.

Scope of the Problem

Many of the Triassic basins of eastern North America have been stratigraphically subdivided into three units because a middle unit, usually composed of dark shales and fine-grained sandstones and occasionally containing coal or fresh water limestone, could be recognized and correlated in the field. This is true of the Bull Run Shale, Richmond basin, Virginia (Roberts, 1928), the Cumnock Formation, Deep River basin, North Carolina (Reinemund, 1955), and the Cow Branch Formation, Dan River basin, North Carolina (Thayer, 1970).

In the Deep River basin the lower unit (Pekin Formation) is lithologically similar to the upper unit (Sanford Formation). Both formations are clastic redbed sequences. The middle unit (Cumnock Formation) consists of gray and black shales, fine-grained sandstones and two beds of coal. Randazzo and others (1970), in a study of the Wadesboro basin south of the Pee Dee River, found that no coal-bearing unit (Cumnock Formation) occurred in this area. Furthermore, the present study also indicates that it is absent in the Wadesboro basin north of the Pee Dee River. Thus, it has not been possible to differentiate the Wadesboro basin sediments into stratigraphic formations. Triassic sediments along the northwest border of the Wadesboro basin are reported to belong to the Pekin Formation (Campbell and Kimball, 1923). Whether this formation extends to the southeast border is not known.

Why is the Cumnock Formation absent in the Wadesboro basin? Is the Wadesboro basin actually a southwestern extension of the Deep River basin in terms of sedimentological regimes? Has the Cumnock Formation been merely faulted out of the stratigraphic section? Does the middle coal unit exist, but only in the subsurface? Does the absence of the coal-bearing unit represent a facies change? In the area of the Coastal Plain overlap could a Triassic lava flow have dammed drainage of southwestward flowing streams, thus causing the formation of a lake or swamp in the Deep River area? Because of differential subsidence, could the Deep River basin have subsided at a greater rate than did the area to the southwest, causing a lake or swamp to develop there?

The Sanford and Durham segments of the Deep River basin are separated by an anticlinal warp (Colon cross structure) where a thin deposit of coarse-grained sediments exist. Reinemund (1955) believed that the coarse-grained sediments in this region represented a large fan or delta of Triassic age. Mann and Zablocki (1961) showed that a similar structure exists in the northern portion of the Wadesboro basin (Pekin cross structure). Could the Pekin cross structure also represent a region where a large Triassic fan or delta was deposited? Were coarse-grained sediments deposited in the area of the Pekin cross structure while finer-grained sediments were carried into topographically lower areas by streams flowing parallel to the basin? Geological mapping of the area, detailed hand specimen studies, and thin section analyses were employed to answer these questions.

In the report, reference made to the Deep River basin will include the Durham basin, the Colon cross structure and the Sanford basin. The Wadesboro basin will refer to the Wadesboro basin including the Pekin cross structure. The Deep River-Wadesboro basin will include the Durham, Sanford and Wadesboro basins along with the Colon and Pekin cross structures (Figure 2).

TRIASSIC SEDIMENTARY ROCKS

Fanglomerates, conglomerates, sandstones, siltstones, claystones and shales make up the Triassic sedimentary rocks of the Wadesboro basin. Reinemund's (1955) lithology definitions are used in this report except for the sandstones which follow Folk's (1968) classification.

Numerous fanglomerate and conglomerate deposits can be found along the northwest border of the northern portion of the Wadesboro basin. The fanglomerate deposits are composed of pebbles and cobbles of brown argillites and milky-white quartz. Argillite fragments tend to be elongate, angular to subangular and weather to a bright red color. Quartz fragments are less common and tend to be more rounded. Brown and maroon sand, silt, and clay form the matrix. Only one fanglomerate deposit has been observed along the southeast border fault.

Conglomerates are most commonly found near the northwestern border and become rare toward the center of the basin. They consist of rounded to subangular pebbles of argillites and quartz in a matrix of sand, silt, and clay. Weathered feldspar is occasionally present. The matrix generally has a maroon or light red color. If feldspar is present, the matrix tends to have a white color. Lenses of clay, silt, and sand within the conglomerates are common. Fanglomerates and conglomerates make up less than 10 percent of Triassic sedimentary rocks within the basin.

Sandstones of the study area were analyzed petrographically. Sandstone types included arkoses, lithic arkoses, feldspathic phyllarenites, subphyllarenites, and phyllarenites. These rocks all contain more than 15 percent matrix and cement and were texturally immature. The ratio of quartz to quartz plus feldspar in the framework of the sandstones was used as a basis for mineralogical maturity determination. These two minerals were chosen to represent mineral maturity because of their overwhelming abundance and their marked contrast of resistance to chemical weathering (Table 1).

Sandstones beds range in thickness from a few inches to a few feet and display small scale cross bedding and graded bedding. The vast majority are grain supported and tightly packed. Sutured and concavo-convex grain contacts are most common but some subphyllarenites are matrix supported and show point and floating grain contacts. The shape of the individual grains ranged from equant to elongate. Elongate grains were oriented to bedding planes. Modal compositional data are summarized in Figure 3.

Siltstones make up approximately one-third, while claystones comprise about one-tenth of the Triassic sedimentary rocks. Maroon fissile shale is present, but is rare. It is usually found in thin beds between layers of claystone and gritty siltstone and occasionally has a gray-green color. Claystones are much more abundant than are shales and they have predominantly the same minerals as sandstones. They

Table 1. Mineralogical Maturity Index.

Supermature	=	>0.95
Mature	=	0.75-0.95
Submature	=	0.50-0.75
Immature	=	<0.50

Maturity Index = Q/Q+F

Ratio of Silica (Q) to Silica Plus Feldspar (Q+F) in Sandstone.

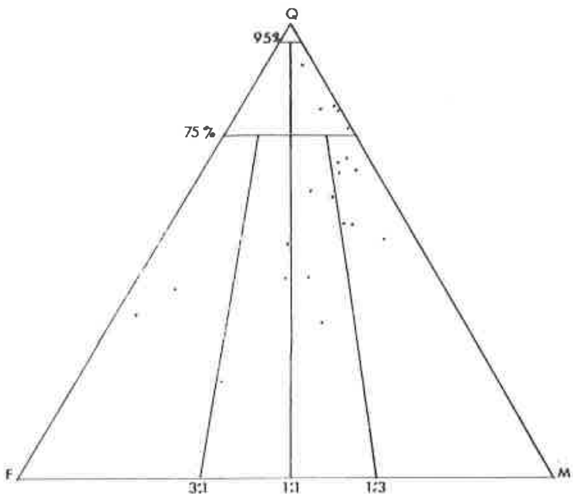

Figure 3. Compositional diagram of Newark sandstones of the northern portion of the Wadesboro basin based on point counts from thin sections (end members and class limits are from Folk's 1968 classification).

are recognized by their nonlaminated and nongritty appearance and usually have a maroon, brown or white color.

The composition and texture of the Triassic deposits change laterally both across and along the strike of the basin. Individual units are characterized by abrupt vertical and lateral changes in color and thickness of strata as well as in texture and composition.

Reinemund (1955) found that coarser-grained sediments in the Deep River basin generally occur along the basin margins while finer-grained sediments were generally located in the central portion of the basin. Randazzo and others (1970) found that in the southern portion of the Wadesboro basin, coarser-grained sediments were located along the northwestern border. Sediments became finer-grained toward the southeast. This trend, although subtle, continues in the northern portion of the Wadesboro basin (Figure 4).

Thickness of the Triassic strata in the Wadesboro basin is difficult to determine. This is due to (1) lack of key marker beds; (2) lack of adequate subsurface data; (3) poor exposure due to abundant vegetation and extreme weathering; and (4) unknown amounts of repetition of strata by concealed faulting. Reinemund (1955, p. 27) estimated the maximum thickness of the Deep River to be about 10,000 feet in the Durham basin, 4,000 to 5,000 feet in the Colon cross structure and 7,000 to 8,000 feet in the Sanford basin. Mann and Zablocki (1961) estimated the minimum thickness to be 3,000 to 6,000 feet in the Durham

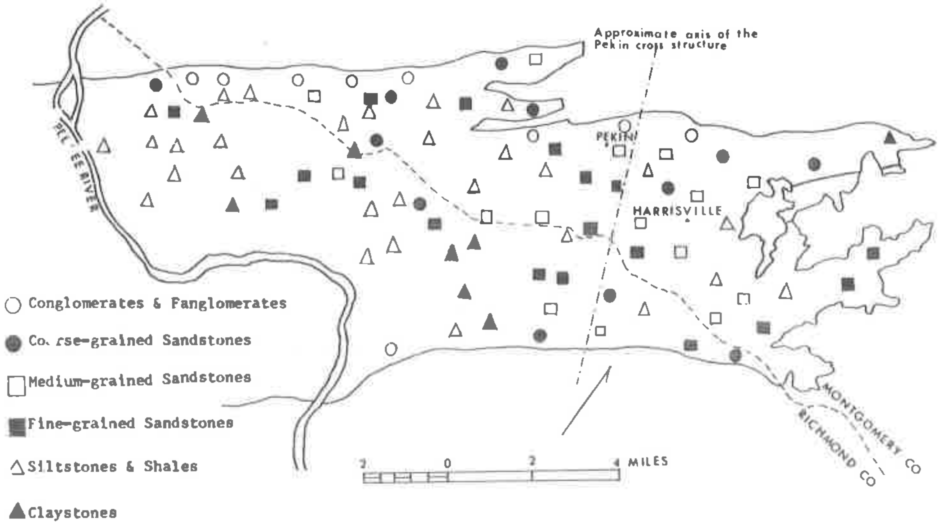

Figure 4. Distribution of sedimentary rocks as related to grain size.

basin, 2,000 feet in the Colon cross structure, about 6,000 feet in the Sanford basin and about 3,800 feet in the Wadesboro basin. There is some 5,000 feet of sediment fill in the deepest part of the Durham basin (Stewart and others, 1973; D. M. Stewart, Personal Communication, 1975).

The only fossils we found in the Wadesboro basin are petrified wood and root impressions. They are normally confined to the northwest border area but exceptions occur. Some of the petrified wood fragments are 12 inches or more in diameter. O. F. Patterson III and G. R. Ganis (Personal Communication, 1975) have found phytosaur teeth and bone fragments near the village of Candor.

Depositional Environment

The Triassic rocks of the Deep River-Wadesboro basin were deposited as alluvial fan, floodplain, lacustrine and swamp deposits (Reinemund, 1955; Randazzo and others, 1970). Fanglomerates and conglomerates are found along the northwestern borders of the Deep River and Wadesboro basins, along the southeastern border of the Deep River basin, and in Triassic outliers southeast of the Wadesboro basin. These coarse sediments suggest that fault scarps with steep gradients existed along the basin margins during Triassic time. The coarse-grained, poorly sorted, immature, and crudely stratified fanglomerates and conglomerates of the Pekin and Sanford Formations were deposited as alluvial fans. Bands of clay, silt and fine-grained sand found within the fanglomerates of the Pekin and Sanford Formations suggest failure of the scarps to be rejuvenated.

Three dimensional configurations of the various finer-grained sequences are lense-like and shoestring. Such deposit geometry along with graded bedding, fine scale cross bedding, desiccation marks, cut and fill structures, petrified wood fragments and root impressions indicate that the finer-grained, well stratified sediments in the central portion of the basin were deposited as stream channel and floodplain deposits. Reinemund (1955, p. 53), in his studies of the Pekin and Sanford Formations of the Deep River basin, observed irregular bedding, poor sorting, fluviatile cross bedding, and channel-like form of some of the coarse-grained sediments. The sedimentary structures noted by Reinemund (1955) and the sedimentary structures and fossils present in the Wadesboro basin are characteristic of fluvial environments as postulated by Visher (1965, p. 49).

Rhythmic and horizontal laminations, uniform fine-grained size, abundant vertebrate and invertebrate fossils, carbonaceous shale, and coal beds led Reinemund (1955) to believe that the Cumnock Formation of the Deep River basin represented lacustrine and swamp deposits. Conley (1962) thought the Cumnock Formation represented a period of relatively slow sedimentation in a shallow lake which supported a lush growth of vegetation. He believed that organic debris, accumulated at the bottom of a lake, was protected from oxidation by burial and eventually formed coal.

Lakes and swamps could probably have formed by the damming of longitudinal drainage in the basin but the cause of this disruptive drainage is not entirely understood. Reinemund (1955, p. 53) noted that ponding of water in the basin could have been caused by warping of the basin floor by differential subsidence or by the growth of fans or deltas which became obstacles to drainage.

Conley (1962) pointed out that the Cumnock Formation is limited to the central and west central portion of the Deep River basin. The red beds deposited along the basin margins during Cumnock time probably belong to the Pekin and Sanford Formations. During Cumnock time, the Pekin and Sanford Formations represented sedimentary facies rather than time-stratigraphic units.

STRUCTURE

Faults

The southeastern border of the Wadesboro basin is formed by a high angle, northeast trending, normal fault which dips to the northwest. A prominent fault scarp (Pea Ridge) makes this fault easily

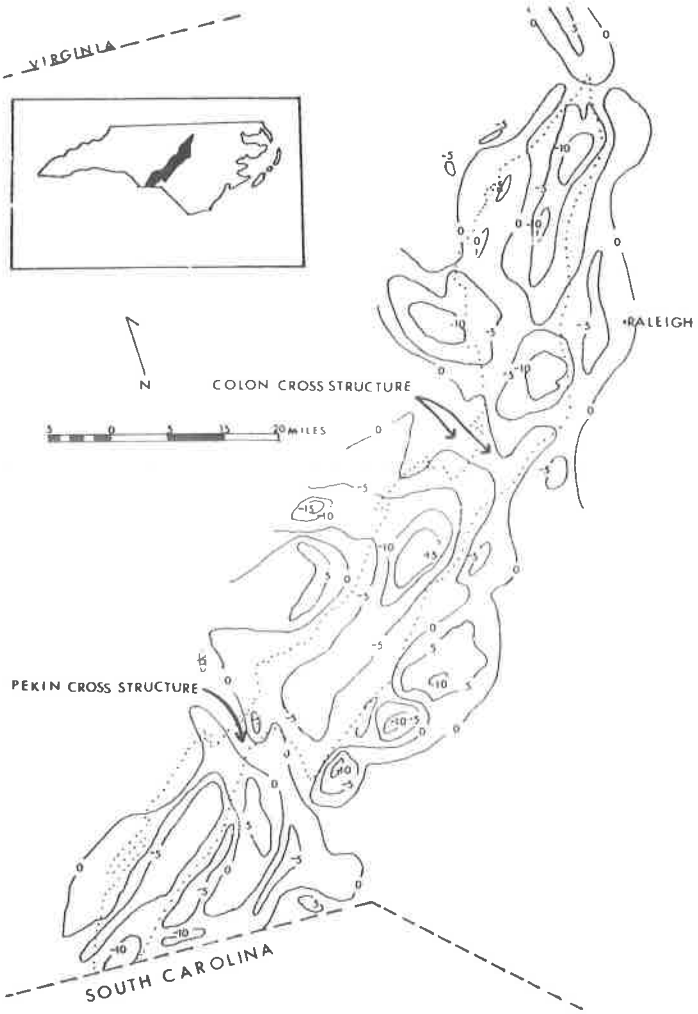

Figure 5. Residual anomaly gravity map of the Deep-River-Wadesboro Triassic basin area of North Carolina showing the location of the Colon and Pekin cross structures (after Zablocki, 1959).

development than a change in dip along the limbs would be expected. Because cross faults, northwest-striking dikes, and many of the joint patterns have a strike similar to the axis of the Pekin cross structure and because the direction of dip does not change along the supposed limbs of the cross structures, we believe that the Pekin cross structure more closely resembles a fault block than a cross fold. The northeastern edge of the fault block may be represented by the inferred cross fault beneath the Coastal Plain overlap west of Norman or by one or

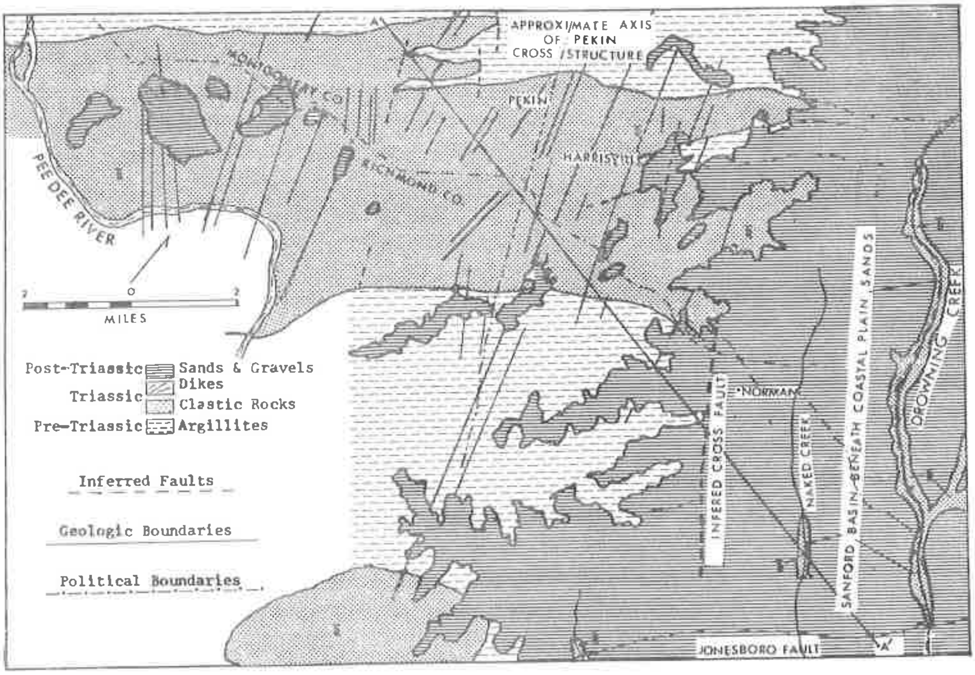

Figure 6. Map of the northern portion of the Wadesboro basin showing the relationship to southeastern Triassic outliers, Jonesboro fault, inferred cross fault, and the Pekin cross structure.

several of the northwest-striking dikes in the extreme northeastern portion of the Wadesboro basin. The southwestern edge of the Pekin cross structure may be represented by one or more of the northwest-striking dikes found in the vicinity of the Pee Dee River.

Reinemund (1955, p. 75) postulated that folding played the most important role in the development of the Colon cross structure of the Deep River basin and that cross faulting later accentuated the fold. The Pekin cross structure probably also began as an anticlinal warp in the basin floor, but cross faulting has apparently played the major role in its development.

Structural Development

The structural development of the various Triassic basins of eastern North America has been discussed by numerous authors (Roberts, 1928; Stose and Stose, 1948; Klein, 1969; Faill, 1973). Reinemund (1955) and Conley (1962) have discussed the development of the Deep River basin.

Reinemund (1955, p. 83) envisioned that downwarping of the earth's crust initiated the development of the Deep River basin and

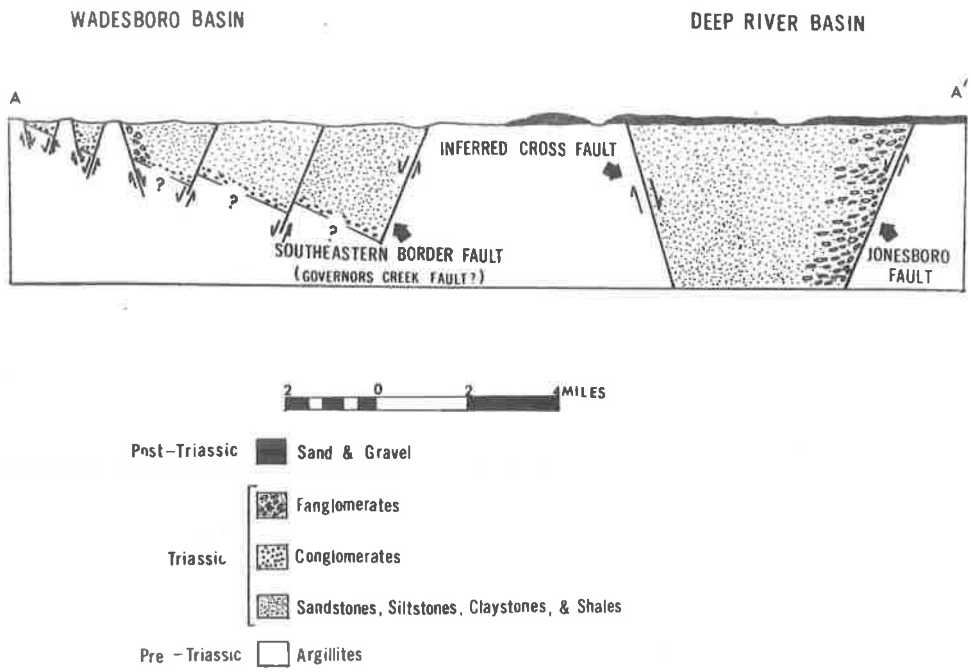

Figure 7. Deduced cross-section of the Wadesboro basin and the Sanford portion of the Deep River basin (section A-A' from figure 6).

faulting did not occur until after deposition of the basal conglomerate of the Pekin Formation. Conley (1962, p. 13) believed that the basin is a rift valley caused by down-faulting along the Jonesboro fault and the numerous western border faults. He also believed that these faults existed in pre-Triassic time and were reactivated during Triassic time to produce the basin.

The Deep River-Wadesboro basin as well as the other Triassic basins of eastern North America appear to be related to the opening of the Atlantic Ocean basin. Dewey and Bird (1970) believe that the formation of the graben of eastern North America was due to the rifting of North America from Europe and Africa in Late Triassic time. Dietz and Holden (1970) envisioned that the rift system of the Atlantic states was related to the breakup of the super continent of Pangaea in Late Triassic or Early Jurassic time. Justus and others (1970) thought that the development of the Triassic basins as well as the entire dike swarm of eastern North America was related to a Mesozoic fracture system and the opening of the Atlantic.

With minor modification we believe that the development of the Wadesboro basin was very similar to the development of the Deep River basin as postulated by Conley (1962) and Reinemund (1955). The sequence of events is as follows:

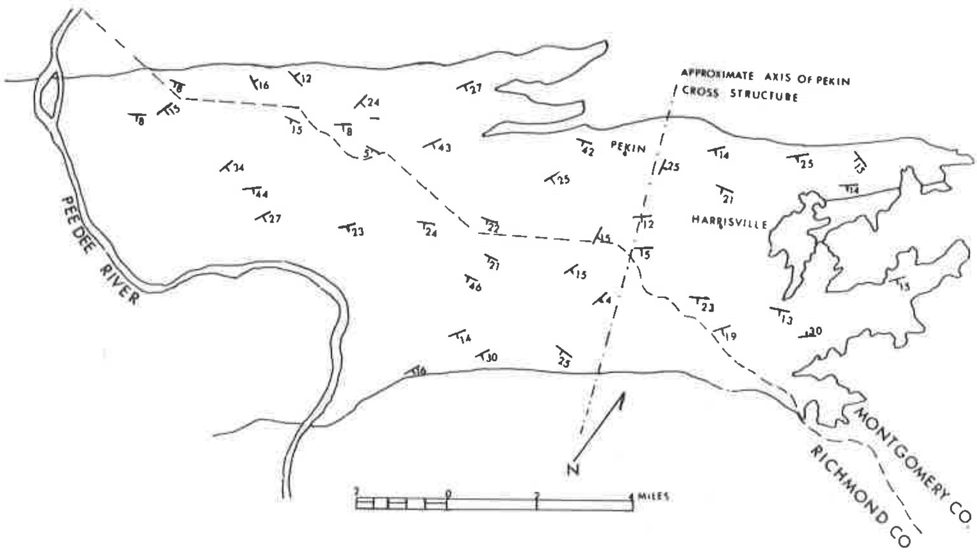

Figure 8. Dip and strike measurements in the northern portion of the Wadesboro basin.

(1) The opening of the Atlantic basin related to a Mesozoic fracture system established a pattern for subsequent episodes of tectonism.

(2) Movement along the Jonesboro fault and along the northwestern border faults caused the development of a graben. Penecontemporaneous to this was the development of the present day southeastern border fault of the basin proper and possibly other longitudinal faults.

(3) Metamorphic and igneous terrain surrounding the basin were eroded and sedimentation began.

(4) The Jonesboro fault became the most active fault. Coarse-grained sediments were deposited along the basin margins while finer-grained sediments were deposited in the central portion of the basin. The southeastern border fault could not keep up with sedimentation or became inactive. Finer-grained sediments were, thus, deposited in the region of the present day southeastern border fault.

(5) Differential subsidence began in the floor of the Deep River-Wadesboro basin along with development of the Colon and Pekin cross structures.

(6) Post-depositional movement of the Jonesboro fault and the Governors Creek fault (Reinemund, 1955, p. 82) in the Deep River basin and along the southeastern border fault in the Wadesboro basin occurred. The Jonesboro fault, which borders Triassic outliers southeast of the Wadesboro basin, apparently became less active than the present day southeastern border fault. Movement of these faults tilted the Triassic strata to the southeast. Cross faults developed and completion of the Colon and Pekin cross structures took place.

(7) New longitudinal faulting or a continuation of old longitudinal

faulting occurred after cross faulting had ceased.

(8) Dolerite dikes intruded Triassic sediments.

(9) Longitudinal faults reached final development.

(10) Triassic strata, between the southeastern border fault and the Triassic outliers lying some eight miles southeast of the present day Wadesboro basin, were eroded.

(11) Thousands of feet of Triassic sediments from the entire Deep River-Wadesboro basin were eroded.

ABSENCE OF THE CUMNOCK FORMATION

Was the Wadesboro basin actually a southwestern extension of the Deep River basin in terms of sedimentological regimes? Did the Triassic sediments of the Deep River and Wadesboro basins have the same provenance? Was the Wadesboro basin actually connected with the Deep River basin beneath the Coastal Plain sands or were the two basins separate entities? If it could be proved that the two basins were separate during the deposition of the Triassic sediments, then the sedimentological history of the Deep River basin would have been independent of the sedimentological history of the Wadesboro basin. The development of the Cumnock Formation in the Sanford basin would not have necessitated its development in the Wadesboro basin.

Our petrologic studies indicate a similar source for the Triassic sediments in the Wadesboro basin as Reinemund (1955) found for the Triassic sediments in the Deep River basin. Through the use of gravity studies, Zablocki (1959, p. 42) showed that the two basins are indeed connected beneath the Coastal Plain sands. The Wadesboro basin is a southwestern extension of the Deep River basin in terms of sedimentological regimes.

The absence of the Cumnock Formation in the Wadesboro basin can be explained in one of the following ways:

(1) The Cumnock Formation was never deposited in the Wadesboro basin.

(2) The Cumnock Formation was deposited in the Wadesboro basin but was in the original eastern half of the basin only and has since been eroded from the stratigraphic section.

(3) The Cumnock Formation was deposited in the Wadesboro basin but presently exists only in the subsurface.

If the Cumnock Formation were never deposited in the Wadesboro basin, then there are several possible explanations. One possibility is that a redbed facies, similar in lithology to the Pekin Formation, occurs in the Wadesboro area. Another possibility is that a natural barrier existed between the Sanford and Wadesboro basins in Cumnock time which created two distinct sedimentological regimes by temporarily disrupting drainage. Natural barriers such as an anticlinal warp in the basin floor, a lava flow or a large alluvial fan or delta are possible explanations.

Is a facies change responsible for the absence of the Cumnock Formation in the Wadesboro basin? It is possible that the Sanford basin was topographically lower than the Wadesboro basin in Cumnock time. The lake and swamp deposits of the Cumnock Formation may have graded laterally into the floodplain deposits of the Pekin and Sanford Formations in the vicinity of the Wadesboro basin. If the climate were dry enough in Cumnock time, then lake deposits may have graded laterally into playa deposits. Wheeler and Textoris (1971) believed these conditions to have existed in the Durham basin. Subsidence in the Sanford basin may have been slightly greater than subsidence in the Wadesboro or Durham basins during Cumnock time. This local subsidence could have been the cause of ponding within the Sanford basin but would not have been of a magnitude strong enough to develop anticlinal warps in the basin floor in the vicinity of the cross structure.

Could a Triassic lava flow, now lying beneath the Coastal Plain overlap have dammed southwestward flowing streams and, thus, caused the development of a lake or swamp within the Sanford basin? Lava flows are believed to have caused a ponding effect within the Connecticut basin (Krynine, 1950) and the Newark basin (Van Houten, 1965) resulting in lacustrine deposits within those Triassic basins.

The only igneous rocks found within the Wadesboro basin are dolerite dikes. Zablocki (1959) was able to locate the existence of subsurface dikes by geophysical methods. It, therefore, seems reasonable to assume that he would have been able to locate the existence of basaltic flows buried beneath the Coastal Plain sands. Zablocki (1959) reported no such peculiarities in his discussion of the Coastal Plain overlap.

Could an alluvial fan or delta have obstructed longitudinal drainage and prevented the development of the Cumnock Formation in the Wadesboro basin? In his discussions of the Colon cross structure, Reinemund (1955, p. 53) pointed out that considerable amounts of coarse-grained sediments were deposited there while finer-grained sediments were carried into both the Durham and Sanford basins by longitudinally flowing streams. He believed that in Cumnock time the Pekin and Sanford Formations were being deposited in the Colon cross structure region while the Cumnock Formation was being deposited in the Durham and Sanford basins. Was the northern portion of the Wadesboro basin also a locus of primary drainage in Triassic time? By comparing the grain size distribution of sedimentary rocks within the northern portion of the Wadesboro basin (Figure 4), it can be seen that an abundance of coarse-grained clastic rocks are not found in this region. If the northern Wadesboro basin were the site of a large alluvial fan or delta, then coarse-grained sediments should have been deposited there. The absence of the Cumnock Formation in this region does not appear to be related to the building up of an alluvial fan or delta in Cumnock time.

Could warping of the basin floor and the development of the

Pekin cross structure have obstructed drainage within the Deep River-Wadesboro basin in Cumnock time? Is the absence of the Cumnock Formation in the southwestern portion of the Sanford basin and the Wadesboro basin related to the development of the Pekin cross structure as an anticlinal fold? Reinemund (1955, p. 82) believed that the structural development of the Colon cross structure in the Deep River basin began during late stages of Triassic sedimentation. He stated that its development began after deposition of the Cumnock Formation had ceased. He believed that it was caused by differential subsidence of the Deep River basin floor and later emphasized by cross faulting. In all likelihood the forces which caused the formation of the Colon cross structure are related to those causing the formation of the Pekin cross structure so that both structures were initiated by differential subsidence and later emphasized by cross faulting. If the development of the Colon cross structure is "post-Cumnock", then the same is probably true for the development of the Pekin cross structure. As previously stated the Pekin cross structure more closely resembles a fault block than an anticlinal fold. Dikes, faults, and joints in the northern Wadesboro basin have a similar strike as does the axis of the Pekin cross structure and no change in direction of dip occurs along the supposed limbs (?) of the fold. Whereas Reinemund (1955) thought that folding played the major role in the formation of the Colon cross structure, we believe that folding only played a minor role in the formation of the Pekin cross structure. If warping of the basin floor only played a minor role in the formation of the Pekin structure, if the Pekin cross structure is "post-Cumnock" and since cross faulting followed sedimentation, then the development of the Pekin cross structure as an anticlinal warp in the basin floor probably had little or no effect on the development of the Cumnock Formation.

Was the Cumnock Formation originally deposited in the eastern portion of the Wadesboro basin and since been removed by erosion? It can be seen that the town of Norman, just south of Candor (Figure 6), lies upon the Coastal Plain overlap which separates the Sanford and Wadesboro basins. To the north of Norman lies the extreme northeastern portion of the Wadesboro basin. To the east, Drowning Creek has cut through the Coastal Plain sands and has exposed Triassic sediments of the Sanford basin. Triassic sediments have also been exposed along Naked Creek south of Norman. The unique feature is that just west of Norman lie pre-Triassic metamorphic rocks of the Carolina Slate Belt. Southeast of these argillites, there exist small Triassic outliers which are terminated by a major fault. This fault is possibly a southeastern extension of the Jonesboro fault (Stuckey and Conrad, 1958). Zablocki (1959, p. 38) states that the minimum thickness of the Triassic sediments beneath these Coastal Plain sands is 7,700 feet. A major cross fault or fault zone is inferred buried beneath the Coastal Plain sands in the vicinity of Norman (Figures 6 and 7). This inference is based on the occurrence of almost 8,000 feet of Triassic sediments along Drowning Creek and beneath the Coastal Plain overlap just east of Norman. Pre-Triassic argillites lie just west of Norman. These slate belt rocks are inferred to represent the upthrown fault block. This upthrown block lies in the vicinity of the Pekin cross structure which supports the suspicion that cross faulting has played a major role in the formation of the Pekin cross structure. If the thickness of the Triassic sediments (2,000 feet) within the northern Wadesboro basin in the vicinity of the Pekin cross structure is correct, and if the Pekin cross structure represents a major fault block, then at least 5,000 feet of Triassic sediments have been uplifted and eroded. Conley (1962, p. 10) stated that the presence of a coal prospect east of Norman indicated that the Cumnock Formation does exist in the extreme southwestern portion of the Sanford basin. The presence of the upthrown fault block (the Pekin cross structure and the pre-Triassic argillites lying west of Norman) suggests the Cumnock Formation was originally deposited in the northeastern portion of the Wadesboro basin but has been uplifted and eroded from the stratigraphic section.

Was the Cumnock Formation deposited in the Wadesboro basin, but presently exists only in the subsurface? Reinemund (1955, p. 68) believed that many normal faults, both longitudinal and cross, have broken the Deep River basin into many discordant fault blocks. Randazzo and others (1970, p. 1003) pointed out the existence of many longitudinal and cross faults in the southern portion of the Wadesboro basin. A similar occurrence of faulting exists in the Wadesboro basin north of the Pee Dee River (Plate 1). If the Cumnock Formation had been deposited in the Wadesboro basin, then it would be probable that the faulting within the basin would have, somewhere, exposed the Cumnock Formation. The Pekin cross structure, which most likely represents a major fault block, would have been responsible for the removal of the Cumnock Formation in the northeastern portion of the Wadesboro basin. The Triassic sediments in the southern Wadesboro basin are believed to be much thicker than in the northern portion of the basin (Zablocki, 1959, p. 42) and the Pekin cross structure may be responsible for this.

The Cumnock Formation has been removed from the stratigraphic section in the northeastern portion of the Wadesboro basin, but it possibly exists in the subsurface of the southern portion of the basin. Further subsurface investigations in the southern Wadesboro basin are needed to ascertain the validity of this hypothesis. Since faulting should have exposed the Cumnock Formation in the southern region of the basin, a more plausible suggestion is needed to explain its absence in that region. It seems probable that the Cumnock Formation graded laterally from dark lacustrine deposits in the Sanford basin to red floodplain deposits of the Pekin and Sanford Formations in the southern Wadesboro basin. The Cumnock Formation was probably present at one time in the northeastern portion of the Wadesboro basin but has since been uplifted and eroded out of the stratigraphic section by the development of the Pekin cross structure.

Another possible reason as to the absence of the Cumnock Formation is that it was originally deposited in the area represented by the upthrown block of pre-Triassic argillites lying west of Norman (Figure 6) but was not deposited in the northern Wadesboro basin. The Wadesboro basin was probably much wider in Triassic time than it is today. It may have extended some distance southeast of the basin as indicated by the presence of small Triassic outliers lying southeast of the present border and at least as far as the Jonesboro fault lying southeast of the basin. If the present day southeastern border fault is actually an extension of the Governors Creek fault (Figure 1) as proposed by Randazzo and others (1970, p. 1002) and if the lakes responsible for the development of the Cumnock Formation were present only in the central portion of the basin in Cumnock time, then the Cumnock Formation may have been deposited in the central portion of the original Wadesboro basin but not along its margins as represented by the present day northern Wadesboro basin. The presence of the Cumnock Formation just east of Norman (Conley, 1962, p. 10) suggests that it was originally deposited in the area of the pre-Triassic argillites lying just west of Norman. The occurrence of an upthrown block (Pekin cross structure) was probably responsible for the erosion of the Cumnock Formation in this area.

SUMMARY AND CONCLUSIONS

The Wadesboro Basin is an asymmetrical fault trough located in the Piedmont Province and is, in places, covered by Coastal Plain sediments. It is bordered along the southeast by a major, high angle, normal fault which dips to the northwest. This fault lacks a coarse facies which is typical of major border faults of other Triassic basins. Post-depositional movement along this fault has given the Upper Triassic non-marine strata a southeasterly dip. Structural "splintering" has resulted in the complex development of a number of horsts and fanglomerate filled-graben along the northwestern border of the basin. Cross faulting truncates the basin near the South Carolina state line and forms the basin's southernmost boundary. In the northeast, Coastal Plain sands separate the Wadesboro basin from the much larger Deep River basin.

Maroon, brown and yellow claystones, siltstones, sandstones, conglomerates and fanglomerates make up the bulk of sedimentary rocks found within the basin. The geometry of the deposits and sedimentary structures such as cross bedding, graded bedding and cut and fill, indicate that these sediments were deposited as coalescing alluvial fans and as stream-channel and floodplain deposits.

Numerous longitudinal and cross faults have cut the basin into a number of discordant fault blocks. Cross faults are, in general, older than longitudinal faults. A major feature of the basin is the Pekin cross structure. Although this feature possibly originated as an anticlinal warp in the basin floor, cross faulting has subsequently given it a fault block appearance. Several small cross folds, related to faulting and diabase intrusions, have been observed within the basin.

Small Triassic outliers lie several miles to the southeast of the Wadesboro basin and fanglomerates occur in some of them. They are terminated by a southwestern continuance of the Jonesboro fault which forms the southeastern border of the Deep River basin. The presence of these outliers containing a coarse facies and the lack of a coarse facies along the southeastern border fault of the present day Wadesboro basin suggest that the Wadesboro basin was at one time wider than it is today.

Dolerite dikes have intruded both the Triassic sedimentary rocks and the pre-Triassic metamorphic rocks which surround the basin. They tend to follow fault and joint planes.

The Wadesboro basin is a southwestern extension of the Deep River basin. The Cumnock Formation, which includes dark shale and coal beds, is found in the Deep River basin but does not crop out in the Wadesboro basin. Field evidence suggests the Cumnock Formation was originally deposited in the northern portion of the Wadesboro basin. Its absence there is probably related to the Pekin cross structure. Reinemund (1955) believed that the Colon cross structure of the Deep River basin was caused by differential subsidence in the basin floor and that it was later emphasized by cross faulting. Zablocki (1959) postulated that the Pekin cross structure of the northern Wadesboro basin and the Colon cross structure of the Deep River basin were similar structures. The forces which caused the formation of the Colon cross structure were probably related to the forces which caused the formation of the Pekin cross structure. Whereas folding apparently played the major role in the formation of the Colon cross structure (Reinemund, 1955), we believe faulting played the major role in the formation of the Pekin cross structure.

A major cross fault has been inferred to exist near the village of Norman (Figure 6). This inference is based on the fact that some 7,700 feet of Triassic sediments occur beneath the Coastal Plain overlap just east of Norman (Zablocki, 1959, p. 38). West of Norman, a distance of less than four miles, pre-Triassic rocks are exposed. The Triassic sediments of the northern portion of the Wadesboro basin have been calculated to be only some 2,000 feet thick (Zablocki, 1959). We believe the pre-Triassic rocks lying west of Norman and the northern portion of the Wadesboro basin represent the upthrown block of the Pekin cross structure. The direction of dip of Triassic sediments in the vicinity of the Pekin cross structure does not change along the supposed limbs of the fold (?). Many cross faults, joints, and dikes have a strike direction very closely related to the trend of the Pekin cross structure. For these reasons we believe that the Pekin cross structure more closely resembles a fault block than a cross fold.

South of the Pee Dee River, Triassic sediments are believed to be only some 3,800 feet thick (Zablocki, 1959, p. 41). Both the Sanford basin beneath the Coastal Plain overlap and the southern portion of the Wadesboro basin have probably been downthrown relative to the Pekin cross structure. Since 7,700 feet of Triassic sediment are thought to exist beneath the Coastal Plain overlap and since the northern Wadesboro basin may only contain some 2,000 feet of Triassic sediment, then at least 5,000 feet of Triassic sediment has probably been eroded from the upthrown stratigraphic section. Since the Cumnock Formation exists beneath the Coastal Plain sands near the village of Norman and since at least 5,000 feet of Triassic sediments have apparently been removed from the stratigraphic section in the northeastern portion of the Wadesboro basin, then the Cumnock Formation has probably been eroded from the stratigraphic section in this area.

The Cumnock Formation probably does not exist in the subsurface in the southern Wadesboro basin. Further subsurface investigations are needed to ascertain whether or not this is true. Since faulting should have, somewhere, exposed the Cumnock Formation, its absence in the southern Wadesboro basin probably represents a facies change.

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A NEW ORDOVICIAN CAMERATE CRINOID FROM KENTUCKY

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ABSTRACT

A new rhodocrinitacean crinoid genus, Simplococrinus, from the Trentonian (Middle Ordovician) Cynthiana Formation of Kentucky, is characterized by an array of primitive calyx features such as very small concealed infrabasals, laterally disrupted basal circlet, strong median ray ridges, and by the absence of median interray ridges and of an anitaxis. The free arms, four per ray, are completely biserial distally and densely pinnulate. The crown of Simplococrinus persculptus presents a complex of characters combining features both morphologically primitive and advanced, making familial assignment difficult. The genus is tentatively assigned to the family Archaeocrinidae. The evolutionary history of related predominantly Ordovician crinoids of the orders Monobathrida and Diplobathrida is reviewed.

INTRODUCTION

Exposures of Middle and Upper Ordovician strata on the flanks of the Cincinnati Arch, especially in Ohio and Kentucky, have long been known as a prolific source of well preserved crinoids, including many morphologically primitive camerates. Both monobathrids and diplobathrids are well represented; in fact no other area has yielded as great a variety of Ordovician Camerata. The great majority of species, and at least two previously known genera (Gaurocrinus and Compsocrinus), are endemic to the region. A multiplicity of names, many of them probably synonyms, has been applied to these forms. Moreover no comprehensive revision of the superfamily Rhodocrinitacea, to which many of them belong, has been attempted in recent years.

These two factors, and additionally the general rarity of Lower

Ordovician crinoids, especially camerates, has rendered the supra-generic classification of these crinoids uncertain. At present the fossil record seems to indicate an explosion of forms in the Upper Ordovician. This is particularly true in the area surrounding the Cincinnati Arch; but is is probably simply an artifact of preservation rather than an accurate reflection of taxonomic diversity. Regardless of its origins the basic effect has been to obscure phylogenetic relations of the group. In the discussion which follows the authors have elected to accept the phylogeny of the Rhodocrinitacea outlined by Moore & Laudon (1943), supplemented by more recent information when available, because a full revision of the superfamily is beyond the scope of this paper.

Familial placement of the crinoid described below as Simplococrinus persculptus, n. gen., n. sp., is rather arbitrary, as the genus presents a peculiar combination of morphologically primitive and advanced characters that are reminiscent of genera in at least two rhodocrinitacean families. Inclusion in the Archaeocrinidae is based mostly on the resemblance of the calyx of Simplococrinus to primitive archaeocrinids such as Rhaphanocrinus. Like Rhaphanocrinus, the new crinoid has a flat base and strong median ray ridges but lacks a median ridge in the anal interray. The radials are separated all around by interradials and the interradials are more or less regularly polygonal, at least in the lower half of the calyx. However the separated basals of Simplococrinus are judged to represent retention of an extremely primitive feature whereas its lack of an antiaxis is most nearly paralleled in the evolutionarily advanced Rhodocrinitidae. The phylogenetic relationships of the relevant families within the Rhodocrinitacea are discussed below. Given the problematic nature of the current classification of diplobathrid camerates we prefer at present to place Simplococrinus within the somewhat broadly interpreted and generalized Archaeocrinidae of Moore & Laudon (1943), rather than to subdivide the family or to erect a new one for a single genus.

Acknowledgments

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SYSTEMATIC DESCRIPTION

Class CAMERATA Wachsmuth & Springer, 1890

Order DIPLOBATHRIDA Moore & Laudon, 1943

Suborder EUDIPLOBATHRINA Ubaghs, 1953

Superfamily RHODOCRINITACEA Bassler, 1938

Diagnosis. - Radials separated from each other laterally by interradials that adjoin basals. M. Ord. -L. Carb. (Mississippian).

Family ARCHAEOCRINIDAE Moore & Laudon, 1943

Diagnosis. - Calyx subglobular to conical; median ray ridges generally present, faint to well marked; fixed interradials and intersecundibrachs in lateral interrays more or less regular, numerous, depressed; aborally becoming smaller, less regular, and indistinguishable from tegminals. Posterior interray wider, typically with extra plates, frequently with anitaxis but without sagittal anal ridge. Free arms uniserial to biserial, branched or unbranched. Column round. Range: Middle-Upper Ordovician.

Genus SIMPLOCOCRINUS, new genus

Type species. - *Simplococrinus persculptus*, new species.

Definition. - As genus is at present coextensive with the genotype, its definition is deferred to the diagnosis of the type species below.

Range and distribution. - Upper Middle Ordovician (Champlainian), Kentucky.

Derivation of name. - From *simplokos* (Greek); complex, involved; in reference to the genus' calyx plate configuration.

Simplococrinus persculptus, new species

Plate 1, Figures 1-4

Diagnosis. - Archaeocrinid with subconical calyx; base flat, lacking basal concavity; infrabasals very small, not visible in side view, concealed by column. Median ray ridges prominent, no ridge in anal interarea. Basal circlet disrupted by intercalation of radials above and small quadrate accessory plates laterally. Lateral interrays regular, each first interbach followed by two and then three large plates; remainder of interrays occupied by a regular series of small plates passing onto tegmen. Posterior interray less regular, primanal followed by three plates; sagittal ridge and anitaxis lacking, rest of posterior interray plates not arranged regularly, gradually decreasing in size aborally. Free arms proximally cuneate, distally biserial, unbranched four arms per ray. Column round.

Material. - One specimen (holotype), consisting of one nearly complete crown and arms and part of calyx of a second crown, Miami University 14164, and an unfigured crown designated a paratype, MU202T.

Locality and horizon. - Roadcut along secondary road 0 5 mi.

NW of road fork just E of Sadieville, Scott County, Kentucky: upper Bromley Shale Member, Cynthiana Formation, Trenton Group (uppermost Champlainian), Middle Ordovician. The holotype was collected by W. H. Shideler.

Description. - Calyx originally depressed subconical in shape, now compressed laterally, wider than high. Base of calyx flat; no basal concavity developed. Infrabasals five, small, equal, subtriangular in shape with rounded base. Infrabasal circlet circular, diameter less than that of calyx base, not visible in side view, hidden by column. Basals hexagonal in side view, five in number, with distal sides merging at bottom of cup to form a planar base under which the infrabasals and proximal columnal are situated; basals flaring strongly upward from cup base, not in contact with each other laterally, separated by radials and by small square accessory plates immediately beneath the radials. Radials five, hexagonal in shape, about equal in surface area to the basals, intercalated to one-half their total height into the basal circlet. All radials widely separated from each other by contiguous basals and adjacent interradials. Interrays narrow, somewhat depressed below level of arm plates, composed of two discrete groups of moderately numerous regular plates in all rays except for CD (posterior) interrray; the latter contains extra plates and is distally wider and less regular in plate arrangement than remaining interrays. Interradials below secundibrachs large, arranged in three series of one, followed by two, followed by three plates (proceeding aborally from cup base) in the four lateral interrays, forming a rough triangle with apex pointing downward and occupying about one-half the total area of the interray. Upper triangle of interradials above the level of the primibrachs elongate, with apex pointing toward tegmen; component plates in all interrays considerably smaller than those of lower triangle, more randomly arranged, merging into tegmen aborally and becoming indistinguishable from marginal tegminals. Interradial plate sequence in posterior interray not resolved into distinct groups, irregular throughout, decreasing in average size and increasing in number gradationally from bottom to top. Primanal approximately equal in size to equivalent interradials, not succeeded above by a sagittal series of anal plates; anal ray ridge totally absent. All interradials ornamented with ridges normal to plate sides and with central raised nodes. Ridges are predominant ornament on larger, regular series of interradials but become threadlike and are displaced in importance by nodes on upper triangle irregular groups in lateral interrays. Transition in ornament type abrupt in each lateral, but gradual in posterior, interray. Plate disposition in normal and CD interrays is compared in text-figure 1 a-b (see also plate 1, figures 1-4).

First two arm branchings within calyx isotomous, producing four equal trunks. Primibrachs two per ray, equal in size to distalmost interradials. Primaxil hexagonal, supporting first secundibrachs and central narrow series of twelve depressed polygonal interprimibrachs

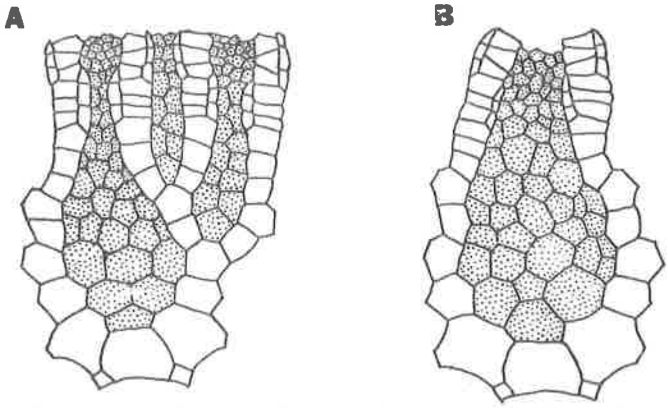

Figure 1. Camera lucida drawings of the holotype of Simplococrinus persculptus (n. g., n. sp.), x6. A-lateral view of normal interray (DE) and part of E ray arm: B-lateral view of posterior (CD) interray.

ornamented peripherally with narrow ridges through midpoint of plate sides and centrally with low rounded nodes occupying most of plate surface. Secundibrachs two in each arm division, barely smaller than primibrachs in overall dimensions. Intersecundibrachs twelve, nodose, arranged in a double column above each secundaxil, decreasing in size approaching the upper cup margin, running onto tegmen.

Remaining two arm branches bilaterally heterotomous, still occuring within the calyx. Tertiaxil and quartaxil are subaxils bearing the main ramus on larger distal facet and an unbranched uniserial ramulus composed of three brachials on smaller facet. Tertibrachs three per ray, slightly smaller than secundibrachs. Quartibrachs three per ray, uniserial like preceeding brachials. First few quintibrachs cuneate, fixed in calyx by calicinal or tegminal plates at immediately adjacent upper border of calyx. Arms completely free by sixth quintibrach, grading from cuneate to totally biserial between sixth and twelfth quintibrach. Biserial portions of arms unbranched, made up of about ninety additional brachials in best preserved arm. Diameter of arms uniform throughout most of their length, decreasing rapidly in last brachials retained on arms of holotype. Free arms extend a minimum of 24 mm beyond the free calyx: distal terminus not seen. Adoral sides of free brachials evenly rounded; brachials incorporated in calyx elevated above surrounding plates by strong broadly rounded median ray ridges (text-figure 1a).

Pinnules dense, subequal in length throughout the whole span of free arm, uniserial, composed of fifteen or more elongate aborally subcylindrical pinnulars. Dimensions of pinnulars subequal, with length barely exceeding width in lateral view. Tegmen largely hidden from sight by arms on more complete crown. A few scattered tegminals are visible on the exposed inner surface of the partial crown (Plate 1, figure 3); they are very small, thinner than calicinal plates, and irregularly polygonal.

Dimensions of holotype (nearly entire crown): crown length 42.8 mm; calyx laterally compressed, maximum width 21.7 mm, minimum 13.9 mm, average width 17.8 mm. Height of calyx 12.8 mm; width of column facet 5.4 mm. The most completely preserved arm has a free portion about 34 mm in length; the longest observed pinnule measures 7.3 mm lengthwise.

DISCUSSION

The systematic position of Simplococrinus (new genus) can be evaluated accurately only in relation to other early camerates. Accordingly the evolutionary history of other closely related diplobathrids and monobathrids is reviewed briefly here. The discussion that follows is restricted for convenience to those families with significant representation in the Ordovician. The suggested phylogeny of these groups is illustrated diagrammatically in text-figure 3. As noted above the phylogeny does not differ substantially from that of Moore and Laudon (op. cit., pp. 76-101, figure 12), except for modification in the earliest lineage proposed by Brower (1974a, b) and herein.

According to these authors the dominant trends in evolutionary development in both the Diplobathrida and Monobathrida are essentially identical. Basically these involve progressive modifications from a many-plated calyx characterized by structural weakness due to large interradial areas and bilateral symmetry through the so-called madreporite plane (Ubaghs, 1953). Both the inadunates and camerates would be derived in this scheme from a hypothetical form incorporating at least some lower brachials and interbrachials in the cup (Moore & Laudon, 1943, p. 77). The main course of evolution in the Camerata would be toward a more simple organization of calyx plates achieved by 1) reduction in the number of fixed interbrachials; 2) elimination of lower brachials from the calyx; and 3) achievement of complete pentamerous symmetry. The most primitive forms recognized in each order were Reteocrinus (dicyclic; diplobathrid) and Xenocrinus (monocyclic; monobathrid). These two genera, as has been noted by several authors, bear a considerable superficial resemblance to each other, aside from the fundamental difference in regard to the presence or absence of the infrabasal circlet. So close is this resemblance, in fact, that at least one diplobathrid (Opsiocrinus) was originally described as a monobathrid (Kier, 1952) and only subsequently recognized as a valid member of the Diplobathrida (Kier, 1958). Moore & Laudon (1943) clearly regarded Xenocrinus and Reteocrinus as being closely related phyletically to each other by divergence from an ancestral form not far removed from either

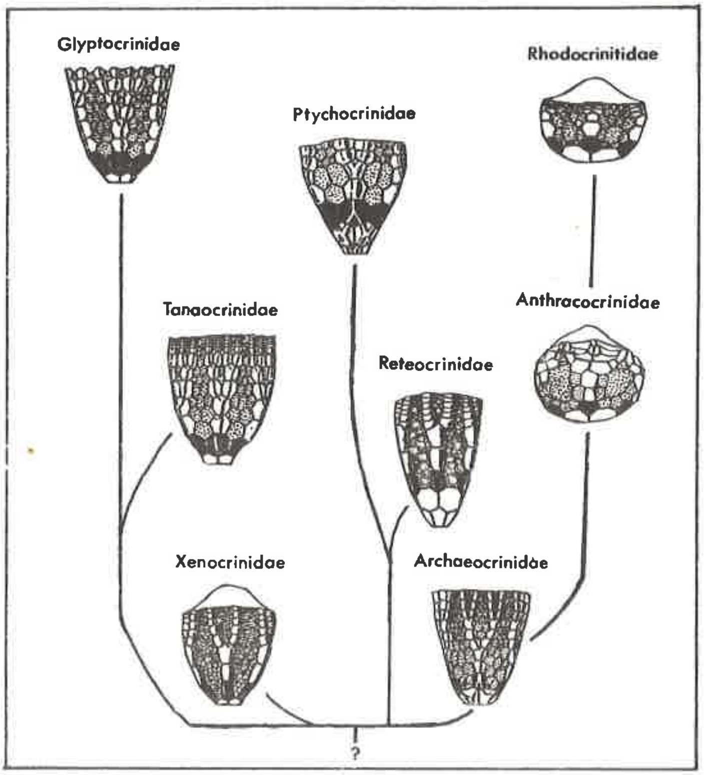

Figure 2. Tentative phylogeny of representative genera and families of Ordovician Camerata. Monobathrida; Xenocrinidae, Xenocrinus; Tanaocrinidae, Tanocrinus; Glyptocrinidae, Glyptocrinus. Diplobathrida: Archaeocrinidae, Rhaphanocrinus; Reteocrinidae, Reteocrinus; Ptychocrinidae, Ptychocrinus; Anthracocrinidae, Hercocrinus; Rhodocrinitidae, Diabolocrinus.

of them in appearance and morphological features. They derived all later families of their respective orders directly from them. The main calicinal features common to the two (numerous small irregular interbrachials and interbrachials located between the radials in most interrays) are regarded as primitive.

This position has recently been challenged by Brower (1974a, p. 2) who prefers to derive at least the early monobathrids (Xenocrinacea) from a crinoid with relatively few, large, and more regular interbrachials. The characters cited by Moore & Laudon (1943) as indicating the primitive nature of Xenocrinus and Reteocrinus are taken by Brower to represent "a specialized adaptation, perhaps in response to respiration" (Brower, 1974b, p. 19). Brower (1974a) relies on two main lines of evidence to substantiate his contention: the general ontogeny of the camerates and statistical analysis of calyx growth, and morphological analysis of some early camerate forms described since the publication of Moore & Laudon's (1943) work. Conclusions from the first-mentioned studies are somewhat vitiated by several factors. Comparatively little work has been done on camerate ontogeny in general and almost nothing on stratigraphically older forms. Examples cited by Brower from the literature mostly involve either dubious material (Cyttarocrinus eriensis (Hall)) or rather atypical forms (Eucalyptocrinites and Platycrinites bozemanensis). Their relevancy to the early history of the Xenocrinacea and Rhodocrinitacea is questionable. The usefulness of the xenocrinid and glyptocrinid growth series (Brower, 1974a, pp. 12-20) is somewhat reduced by the small number of individuals employed. The second line of evidence is more convincing. Substantial additions to the early diplobathrid lineage have been made by Strimple & Watkins (1955) and to the early monobathrid lineage by Brower (1974) and Strimple & McGinnis (1972). The Lower Ordovician Proexenocrinus Strimple & McGinnis, a form with large, relatively few interbrachials and uniserial arms, is judged to be ancestral to later monobathrids. Demonstration by Brower that most specimens of Xenocrinus do have laterally conjoining radials in at least some rays reduces the taxonomic distance between the two genera. Discrimination of the family Anthracocrinidae and the addition of new genera to the early Archaeocrinidae has clarified the origins of the Rhodocrinitidae and established the early divergence of the Archaeocrinidae from the common monobathrid-diplobathrid line.

Acceptance of Brower's derivation of the Xenocrinacea (and, by extension, the Rhodocrinitacea), from a Proexenocrinus-like ancestor, plus his interpretation of Xenocrinus and Reteocrinus as morphologically advanced forms, necessitates reevaluation of the entire Ordovician lineages of both orders. A tentative phylogeny along these lines is presented in text-figure 2. Evolution of both the Monobathrida and the Diplobathrida from an unknown pre-Ordovician inadunate form is assumed. The Xenocrinidae are treated as a dead-end offshoot from an as yet largely undiscovered main monobathrid line. The Reteocrinidae are believed to represent an early side branch from the principal diplobathrid line that gave rise to the Archaeocrinidae and all later forms with the exception of the Ptychocrinidae, which probably diverged directly from the Reteocrinidae.

The presence of numerous interradial interbrachials in Simplococrinus, as well as the accessory plates separating the basals and the reduced infrabasals, are judged to be advanced features and probably served a similar function. Other characteristics of the genus, such as the biserial free arms, the degree of incorporation of the arms, the partly intercalated basal and radial circlets, and the lack of an anitaxis and associated anal ridge may also be taken to indicate the evolutionarily advanced nature of Simplococrinus. Conversely the regular large interradial interprimibrachs, only slightly depressed interrays, four free arms per ray, and flat base may be retained archaic features more strictly typical of the ancestral diplobathrid line.

A rather unusual feature of Simplococrinus, the small pumilate ramuli fixed in the calyx, needs further comment. Incorporation of pinnules into the calyx is a widespread tendency among early Diplobathrida and is especially common among genera of the Reteocrinidae, Archaeocrinidae, and Anthracocrinidae. It is in fact a diagnostic character of the latter (Strimple & Watkins, 1955, p. 348). Fixed pinnules in these groups generally are short, are invariably in lateral contact with fixed pinules originating on other fixed main ramus brachials, have axillary facets with a width that is only a small fraction of that of the main ramus, and are given off alternately from every calyx brachial in the main arm series once pinnulation has begun. The small fixed arm appendages of Simplococrinus are similar except that they occur only on every third main arm brachial, have axillary facets subequal in width to those of the main arm, have first brachials equal in size to adjoining main ramus brachials, and are not in lateral contact with other such appendages (Plate 1, figure 4); they are best interpreted as reduced major arm branches (ramuli) rather than pinnules.

The fixed pinnulate condition is most often encountered in Middle and early Upper Ordovician forms and is most likely to be absent in advanced families such as the Rhodocrinitidae. Early Rhodocrinitacea, with or without incorporated pinnules, tend to have the rays separated by interposed interradial interbrachs that continue aborally without interuption from the calyx onto the tegmen. In the process of reducing the number of fixed calyx plates, probably as a direct consequence, later Rhodocrinitacea often bring the rays into lateral contact, isolating the interradials from the tegmen. As Simplococrinus is unique, according to traditional interpretation, among rhodocrinitacean camerates in having fixed ramules the evolutionary significance of the character is not easy to assess. The separated rays and most other arm features mentioned in the foregoing discussion seem to indicate that the overall aspect of the genus' arms is decidedly primitive. Alternatively it is possible that the distribution of fixed ramuli among the families of the Diplobathrida is more widespread than hitherto supposed. Undoubted fixed pinnules are present in Gaurocrinus (Reteocrinidae) but the "pinnules" incorporated into the calices of members of the archaeocrinid-anthracocrinid line (e. g. Hercocrinus and Deocrinus) may represent ramules that have secondarily come to resemble pinnules. Elimination of nonaxillary main arm brachials and reduction of ramulus brachials' dimensions and number would produce such a result. In this case incorporated pinnules being given off by a single main arm would constitute a significant advance over those forms in which survive evidence of several extra main arm branchings in the form of diminutive fixed ramuli.

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THE ORIGIN OF CALCIUM SILICATE AND ASSOCIATED MINERALS FROM A TRIASSIC SILL NEAR DURHAM, NORTH CAROLINA

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ABSTRACT

A suite of four minerals in the system CaO-SiO_2-H_2O, together with a variety of associated silicate minerals, occurs as alteration or subsequent crystallization products along and in a fracture system in a deuterically altered gabbro sill-like unit. Xonotlite, tobermorite, gyrolite, and okenite formed under low pressure and temperature. Their depositional sequence was modified by the ionic composition and ratios of the depositing solutions.

INTRODUCTION

The deuteric alteration by hydrothermal solutions along a fracture system in a gabbroic sill-like body and the mineralization that followed was studied in an attempt to delineate the possible controlling factors involved in the process and to illustrate the significance of solution content during the episode. Overlap of mineral species deposition pointed to controlls other than temperature and pressure.

Authors such as Surdam (1973) and Liou (1971) have strongly suggested that ionic activity could be an important factor in the anomalous variations that occur in low grade metamorphic terraines. If this is true, verification of such assumptions might lead to a better understanding of some of the problems that now exist in this area.

GEOLOGIC SETTING

A series of massive sill-like gabbroic bodies either shallowly underlie or are partially exposed along the northwestern border of the Durham, North Carolina, Triassic basin. In the area of this study they bear both an intrusive relationship at varying stratigraphic levels within their Triassic sedimentary host rocks or are exposed at the surface by erosion.

In reconstructing the total stratigraphic column of the western border of the Durham Triassic Basin, Harrington (1948) estimated approximately five hundred meters for the minimum depth of sediments to have been removed by erosion from the study area. Under these circumstances the minimum lithostratic pressure would have been approximately 0.15 kb. With reasonable assumptions of increased load the lithostatic pressure would still remain very low.

Though no evidence of extrusive nature was found in this area, Koch (1967) has suggested that field data indicate an extrusive nature for much of the diabase along the northwestern boundary of the basin in the area to the northeast of the study area. If Koch's (1967) assumptions of extrusion are correct for the area directly to the northeast of the study area, the above estimate of overburden depth might be reduced.

The Triassic sediments of the study area, into which the sills bear a semi-conformable intrusive relationship, lie unconformably upon a rugged erosional surface of nearly vertically dipping meta-volcanic rocks of the Piedmont Provinance. They consist of basal conglomerates and conglomerates with interlayered feldspathic and argillaceous sandstones.

At the Nello Teer crushed rock quarry, which is located north of Durham, North Carolina, on Durham County road 1641 and from which the minerals of this study were obtained, the integral relationship of the three units can readily be seen in section. An inlier of Piedmont 'slate' rises above the surrounding Triassic basin rocks. The higher sections of its irregular profile are flanked with Triassic conglomerates and coarse sediments and its lower expression is overlain by them. A large gabbroic unit, that appears to originate as a feeder dike, cuts through the meta-volcanic unit and rises approximately forty-five degrees to the west-northwest in the lower level of the quarry. In the higher levels of the quarry it assumes a lateral expression and spreads horizontally to the west-northwest into the host Triassic sediments.

The gabbroic unit is characterized by a fine-grained, chilled border zone at its contact with the hornfels of the meta-volcanic host rock. Textures throughout the rest of the gabbroic unit range from random pegmatitic to diabasic to normal gabbroic. Superimposed upon both the gabbroic unit and host meta-volcanic unit is a fracture system that cuts both units and all textures. It is in this setting that the minerals under discussion occur, along and in that portion of the fracture system confined to the coarser textured part of the gabbroic unit.

Prehnitization and Laumontization

Two massive basic changes occur along and in conjunction with fractures in the basic gabbro host rock. As deuteric solutions reacted with the gabbro sill, which contains predominantly plagioclase feldspar, diopside or augite and some olivene, one of two processes occurred. The basic gabbro host rock was altered to a predominantly prehnite-chlorite rock or to a laumontite-chlorite rock. Either type may contain a later sequence of one or more of a variety of low temperature silicate minerals (Table 1) as end products of the particular process that took place.

Calcium Silicate Mineralogy

Tabulation of the total mineral suite and their important relationships is shown (Table 1). Only the minerals belonging to the system CaO-SiO_2-H_2O and those of unusual occurrance will, however, be discussed separately. Identification of all minerals was by x-ray diffraction.

Okenite ($CaSi_2O_4(OH)_2 \cdot H_2O$) was a minor calcium silicate mineral, although it was abundant in the specific specimens on the few occasions when it was found. It was always associated with and after (Table 2) prehnite in fractures and vugs. When it did not fill the openings in which it was found it consisted of radiating to matt-like crystal groups. When it filled the openings it was composed of solid massive radiating acicular to matted crystal groups. The color was white to cream with a silky luster. Many times the crystals that ended in non-filled openings appeared as though they had been fused and bent.

Walker (1971) found okenite associated with other calcium silicate minerals low in the Mull and Morvern basalt pile where he concluded they formed at relatively high temperature. The central core of the Mull volcano was surrounded by an inner epidote and an outer prehnite zone. Tschernich (1972) reported okenite in a vug, associated with earlier quartz and later apophyllite, from the Tertiary basalt flows at the Skookumchuch dam site. Stephens and Bray (1973) report okenite, with the other silicates, in the silicated and hydrothermally altered limestone members of the Oquirrh Formation in the southwestern portion of the Utah Copper Mine at Bingham, Utah. This alteration is thought to have resulted from the same hydrothermal activity that formed the ore minerals.

Xonolite ($5CaSi_3 \cdot H_2O$) occurs as vein fracture fillings, usually in a sequence after prehnite and before gyrolite (Table 2). Two textural types occur either separately or together: 1) flesh pink colored masses composed of radiating shieves of well crystallized material extending inward from the vein walls. 2) Light pink to powdery white, fine grained, massive material that usually follows the crystallization of textural type #1. Under the microscope it is seen to be composed of a fine crystalline fiberous mesh. Within type two, isolated small blebs or patches of type one occur and present a color and texture contrast.

Authors such as Berman (1937), Taylor (1954), Bilgrami and

Table 1. Crucial Minerals With Their Formulas and C/S Ratios.

Mineral	Formula	Moles CaO	Moles Al_2O_3	Moles SiO_2	C/S
Prehnite	$Ca_2Al_2Si_3O_{10}(OH)2$	2	1	3	.66
Laumontite	$CaAl_2Si_4O_{12} \cdot 4H_2O$	1	1	4	.25
Xonotlite	$5CaSiO_3 \cdot H_2O$	5	-	5	1.00
Tobermorite	$Ca_5Si_6O_{16}(OH)_2 \cdot 4H_2O$	5	-	6	.83
Gyrolite	$Ca_2Si_3O_7(OH)_2 \cdot H_2O$	2	-	3	.66
Okenite	$CaSi_2O_4(OH)_2 \cdot H_2O$	1	-	2	.50
Reyerite	$KCa_{14}Si_{24}O_{60}(OH)_5 \cdot 5H_2O$	14	-	24	.58
Apophyllite	$KCa_4Si_8O_{20}(F, OH) \cdot 8H_2O$	4	-	8	.50
Chabazite	$CaAl_2Si_4O_{12} \cdot 6H_2O$	1	1	4	.25
Natrolite	$Na_2Al_2Si_3O_{10} \cdot 2H_2O$	-	1	3	---
Saponite	$(Ca/2, Na)_{0.33}(Mg, Fe)_3(Si, Al)_4O_{10}(OH)_2 \cdot 4H_2O$				
Epidote	$Ca_2(Al, Fe)_3Si_3O_{12}(OH)$				
Corrensite	$Mg_8Al_3Si_6O_{20}(OH)_{10} \cdot 4H_2O$				
Chlorite	$(Mg, Fe^2, Fe^3, Mn)_6(AlSi_3)O_{10}(OH)_8$				
Thaumasite	$Ca_3(CO_3)(SiO_3)(SO_4) \cdot 14.5H_2O$				
Quartz	SiO_2				
Calcite	$CaCO_3$				

Howie (1960) and Speakman (1968), and others throughout the literature consider the Ca/Si ratio of the mineral to be constant at 1.0 but sharply disagree on the water content. From infra-red studies Kalousek and Roy (1957) indicated unbonded hydroxyl groups and Bilgrami and Howie (1960) suggested that if this is true the xonolite formula might, therefore, read $Ca_5Si_5O_{14}(OH)_2$. They also reported a H_2O content of from 2.65 percent to 2.85 percent for xonothlite specimens from various localities.

Gyrolite ($Ca_2Si_3O_7(OH)_2 \cdot H_2O$) is one of the more abundant calcium silicates found in the deposit. It occurs as a vein filling and under these circumstances is composed of shieves of laminar crystals attached to the vein walls. It radiates into, and either partly or wholly fills the cavity. Color of the crystals range from pure white to shades of pale green. Gryolite was found to crystallize either along or simultaneously with a variety of silicates including prehnite, apophyllite and laumontite (Table 2).

Tobermorite ($Ca_5Si_6O_{16}(OH)_2 \cdot 4H_2O$). The 11A variety is

Table 2. Paragenetic sequence of vein mineral samples.

Vein Minerals of: Prehnitized Wall Rock	Vein Minerals of: Laumonitized Wall Rock
preh.	lau + sap
preh. → gyro	lau + sap → preh. → gyro
preh. → gyro + preh^{o}	lau + apo
preh. → gyro + preh^{o} → tobo	lau → apo + Ca
preh. → gyro + apo + lau	lau + ca → apo
preh. → gyro → apo	lau + ca → sap
preh. → gyro + muscovite → apo	lau → gyro
preh. → gyro + lau	lau → gyro → gyro + preh^{o}
preh. → gyro → xono	lau → gyro + apo
preh. → gyro → tobo + preh^{o}	lau → gyro + x + ca → tobo + preh^{o}
preh. → xono → gyro	lau + corr → corr → corr + preh^{o}
preh. → xono → gyro + Ca	lau → apo + lau → apo → thau
preh. → xono + Ca	
preh. → lau + sap	
preh. → lau + gyro → gyro + Ca + apo	
preh. → gyro + corr + preh^{o}	
preh. → oken	
preh. → rey	

Apo = apophyllite; ca = calcite; corr = corrensite; gyro = gyrolite; lau= laumontite; oken= okenite; preh = prehnite; sap =saponite; rey = reyerite; thau=thaumasite; tobo = tobermorite; xono=xonotlite; x = unknown; preh^{o} = prehnite as suspended, unattached particulate spheres within co-precepetate.

present and is quite common, though minor in amount. It is generally the last mineral of any sequence to crystallize and is usually found after gyrolite (Table 2). It usually occurs as a massive fine grained, sugar-like textured, white vein filling. In sharp contrast to this type of occurrence is crystallization in an open matt of fine, acicular crystals forming a product that looks like 'Angels Hair.' In either case, small prehnite crystalline spheres are sometimes found within the mass, having crystallized simultaneously with the tobomorite (Table 2).

Reyerite (See formula below) was found in minute quantity in only one sample as the end product of crystallization of a fracture vein containing prehnite.

Considerable controversy has existed as to the validity of reyerite as an individual species. Mackay and Taylor (1953) reported a specimen from Greenland to be a species distinct from gyrolite but questioned its relationship to truscotite. Stunz and Micheelsen (1958) considered reyerite and truscotite to be a single species while Chalmers et al. (1964) separated the two on optical and X-ray diffraction properties. The most recent work by Clement and Ribbe (1973), on material from Brunswick County, Virginia, compare the Virginia material with material from Greenland and with literature of reyerite from Scotland (Cann, 1965), gyrolite from Bombay (Mackay and Taylor, 1953) and Scotland (Cann, 1965) and truscotite from Sumatra (Mackay and Taylor, 1954), Japan (Minato and Kato, 1967) and concluded that gyrolite, truscotite and reyerite are each distinct mineral species. They gave the formula for the Virginia material as

$$(Na_{0.3}K_{0.7})_{3.8}Ca_{13.9}(Si_{0.9}Al_{0.1})_{24}O_{60}(OH)_{5.2} \cdot 5H_2O$$

Corrensite (1:1 Chlorite-vermiculite) normally occurs as a residual mineral from either clay or limestone. A single exception, reported by Gradusov (1969), is its occurrence in asbestos rock of the Zerevta deposit where it is found associated with talc, anthophyllite and chrysotile asbestos. At the Durham, North Carolina, location it is found as a hydrothermal alteration product of chlorite or as a hydrothermal vug or vein filling. In one case its formation pre-dated a vein filling of xonotlite; in another it crystallized after prehnite but simultaneously with gyrolite and suspended prehnite spheres. In a third case it progressed from simultaneous crystallization at the edge of a vug with laumontite to pure corrensite, to corrensite with suspended inclusions of prehnite spheres (Table 2).

Color varies from pure white to a very light green. It is composed of massive to radiating groups of foliate crystals.

DISCUSSION

Prehnitization of gabbro bodies, which has a parallel process in laumonitization, is generally attributed to the action of hydrothermal solutions derived from outside the host body. Bloxam (1954), considered such alteration of the gabbros in the Ayrshire district to have resulted from lime rich solutions and CO_2 that post-dated serpentinization. Watson (1953), related the prehnitization of gabbros to hydrothermal solutions derived from associated serpentine and (Watson, 1942) to hydrothermal solutions generated by a nearby granitic intrusion. The laumontitization and prehnitization of the Durham gabbro resulted, however, from the direct action of late deuteric solutions.

Huang (1962), compared the chemical composition of an altered prehnite-zoisite zone in gabbro from the Wichita Mountains, Oklahoma, with a portion that had been least altered. He found a large increase in lime and a decrease in silica, iron, magnesium and alkalic content in the altered materials. The altered zones of laumontite, prehnite-rock and their associated calcium aluminum silicate minerals at the Durham, North Carolina, location also reflect a high rise in CaO over their host gabbro rock. A list of formulae are given below to illustrate the possible changes induced in the major host rock minerals, to show the resultant minerals and to indicate the possible compositional change that could take place.

(an) (preh)

1) $2Ca\ Al_2Si_2O_8 + \underline{H_2O} \rightarrow Ca_2Al_2Si_3O_{10} \cdot (OH)_2 + \underline{Al_2O_3} + \underline{SiO_2}$

(an) (lau)

2) $CaAl_2Si_2O_8 + \underline{4H_2O} + \underline{2SiO_2} \rightarrow CaAl_2Si_4O_{12} \cdot 4H_2O$

(diop) (preh)

3) $2CaMgSi_2O_6 + \underline{Al_2O_3} + \underline{H_2O} \rightarrow Ca_2Al_2Si_3O_{10} \cdot (OH) + \underline{2MgO} + \underline{SiO_2}$

(diop) (lau)

4) $CaMgSi_2O_6 + \underline{2SiO_2} + \underline{Al_2O_3} + \underline{4H_2O} \rightarrow CaAl_2Si_4O_{12} \cdot 4H_2O + \underline{MgO}$

(diop) (chl)

5) $12CaMgSi_2O_6 + \underline{8H_2O} + \underline{Al_2O_3} \rightarrow 2Mg_6(Al, Si_3)O_{10} \cdot (OH)_8 + \underline{12Ca(OH)_2}$
$+ \underline{18SiO_2}$

These formulae are not given to represent specific or complete reactions but rather to reflect possible directions. It should be noted that calcium enrichment of the solution takes place when chlorite forms but that the total solid system does when either prehnite or laumontite are the end products.

Petrographic relationships of laumontite and prehnite in the Durham deposit do not show replacement of one by the other. Further, the paragenetic sequence of altered wall rock and veins shows that either mineral may follow the other (Table 2). If only the altered wall rock is considered, however, it is found that zones of laumontized and prehnitized wall rock occur within a short distance of one another and cannot, therefore, be totally temperature controlled unless the fracture system containing laumontite formed later than the associated prehnite alteration. If this were the case, prehnite would not be deposited after laumontite crystallization was initiated. This situation has its parallel in the point made by Surdam (1973), where wairakite and laumontite occur in adjacent amygdales in the flows of the Karmutsen Group. He suggested that control in this case was a small difference in the activities of Na+ and Ca++ because laumontite does not readily accept Na+ into its structure whereas wairakite does. Rusinov (1965), suggested that the transition from laumontite to prehnite takes place at a temperature of about 200°C. Eugster (1970), however, postulated that in metamorphic reactions involving a fluid phase, both ionic and thermal stability should also be considered.

Following this, Surdam (1973), noted that in the Triassic Karmutsen Group of Central Vancouver Islands laumontite and prehnite overlap in a (3,000 M) 10,000 ft. zone (100°C). He postulated that while such overlap is contrary to temperature condition, it can be explained in terms of ionic stability. He (Sudam, 1973), also compared data from the occurrence at Butte Lake, where Campbell and Fyfe (1965) estimated the upper limit of stability of laumontite to be 150°C + 50°C, with data from the New Zealand Geosyncline where Coombs et al. (1959) set the gradual transition of laumontite to prehnite at 300°C. Such an anomaly in thermal stability must be accounted for and Surdam suggested that a minor fluctuation in the activities of Ca^{++} and (or) H^+, and (or) SiO_2, in the aqueous solution can stabilize either laumontite or prehnite within a limited temperature overlap.

The results of the work of Speakman (1968) and of Harker (1960) can be related directly to the other mineral involved in the Durham deposit. Speakman (1968), in his work on the stability of tobermorite (between 0.35kb and 2.75 kb to 400°C), found that tobermorite and its associated minerals in the $CaO-SiO_2-H_2O$ system crystallized from starting material within definite CaO to SiO_2 (C/S) ratios and from this (Figure 1) concluded that all equilibria in the $CaO-SiO_2-H_2O$ system below C/S 1.0 can only be expressed in terms of truscotite (reyerite), gyrolite, tobermorite, xonotlite, silica, and H_2O (Speakman, 1968, fig. 2, p. 1102).

With small variations, the results of Speakman's work parallel those of Harker (1960) and others. Although Harker (1960) found that gyrolite plus silica converted to reyerite plus H_2O at less than 180°C, he also set the reaction of gyrolite to reyerite plus xonotlite plus H_2O at 220°C which is very similar to the results of Speakman (1968).

Little is known of the stability range for okenite except that Balitskiy and Gorbunov (1967) synthesized it within the experimentally restricted range of 0.4 kb between 320°C and 340°C. Within this range they found formation to be dependent upon the SiO_2/CaO ratio of the precipitating solution.

Mineral sequences of the Durham deposit (Table 2) where the C/S ratios are the same, such as in prehnite to gyrolite, represent a simple, continuous temperature drop and depletion of aluminum in the depositing solutions during crystallization. When the sequence moves from prehnite to gyrolite back to prehnite, however, there is a temperature overlap. The sequence is, in that case, dependent solely on the variation of the Al^{3+} concentration in the solution.

The upper limit for the temperature of stability of gyrolite has been documented for various pressures and ionic ratios and the subsequent temperature range is small (Figure 1). Any mineral that follows gyrolite in the depositional sequence must, therefore, have crystallized below the range of top temperature for the stability of gyrolite. Because of its abundance in the deposit, it becomes a very critical mineral to this study for temperature deliniations in the many vein sequences that contain it (Table 2).

A mineral sequence preceeded by laumontite also represents a low temperature sequence. Such a sequence that ends with gyrolite or gyrolite and prehnite can only be formed by a change of ionic ratios of Ca, Si, and Al. Although the sequence prehnite to okenite or prehnite to reyerite represents a higher temperature of formation than does a sequence containing laumontite as the initial crystallizing mineral, it can only be related to a change of Ca/SiO ratios and not temperature.

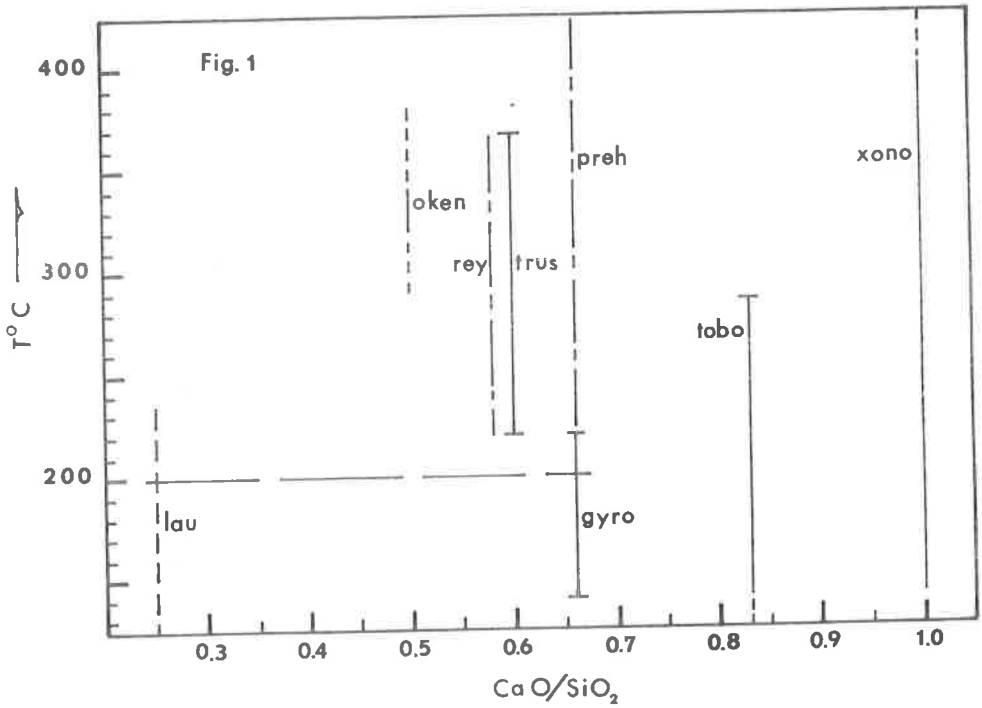

Figure 1. Approximate temperature stability ranges of some compounds in the CaO-SiO_2-H_2O, CaO-SiO_2-Al_2O_3-H_2O systems. Solid lines after Speakman. Okenite (oken.) after Gorbunov. Laumontite (lau.), reyerite (rey.), and prehnite (preh.) inserted on chemical formula C/S ratio. Horizontal indicator related to laumontite-prehnite after Rusinov.

Under these circumstances, a number of the vein filling sequences (Table 2) can only be rationalized by a fluctuation in concentration and ratios of the ions in the depositing fluids. Thus, though temperature was an important factor, it was overridden by fluctuating ionic concentration and ratios of the involved ionic species in solution.

CONCLUSIONS

Prehnitization and laumontization of the gabbroic host rock took place along fractures by deuteric action at low pressures and temperatures. The resulting solution was enriched in CaO and a series of vein filling minerals in the CaO-SiO_2-H_2O and associated systems formed.

The petrographic evidence indicates that deposition was controlled predominantly by the activity of ionic species such as Ca^{2+}, Si^{4+}, and Al^{3+}, which acted in conjunction with a constantly falling temperature control but which overrode it.

Corrensite developed from hydrothermal fluids by deuteric action within the temperature stability range of such minerals as prehnite and gyrolite.

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PALYNOLOGICAL STUDIES OF THE BULL CREEK PEAT, HORRY COUNTY, SOUTH CAROLINA: GEOMORPHOLOGICAL IMPLICATIONS

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ABSTRACT

A bluff exposure on the east bank of Bull Creek on the border of Horry and Georgetown Counties, northeastern South Carolina, contains two superimposed Pleistocene dune sheets underlain by a peat horizon. A soil and humate zone occurs on the surface of the lower dune sheet. There is a radiocarbon date of 16,900$\pm$320 (I-1897) from the humate and a suspect date of 36,200+3600,-2500 (I-1898) from the peat. Despite the finite date from the peat, extensive geomorphological work by Thom (1967, 1970) suggests that the lower dune sheet was formed during the earliest Wisconsin glaciations, that stabilization and soil formation occurred during the mid-Wisconsin Plum Point and Port Talbot interstadial, and that the upper dune sheet formed during the classical Wisconsin.

Pollen analysis of the peat indicates that it was deposited in a fresh-water, probably riparian environment. This is suggested by the pollen of many aquatics, the occurrence of trees, shrubs and herbs characteristic of wet or boggy shores, and the absence of taxa indicative of brackish situations. The data also suggest an interglacial rather than interstadial age. This is inferred from the relative absence of "boreal" and "northern hardwoods" types, the presence of "austral" taxa (Cyrilla, Magnolia, Symplocos, Osmanthus), and the similarity of the pollen spectra to both modern and interglacial assemblages from the Coastal Plain. Thus the radiocarbon date from the peat appears to be too young, and Thom's geomorphological interpretations are supported.

INTRODUCTION

Extensive geomorphological studies in Horry and Marion Counties, northeastern South Carolina, have provided important new insights concerning the Pleistocene history of the area (Thom, 1967, 1970; Du Bar et al., 1974). It now appears as if the diverse landforms of the region have had a complex history both directly and indirectly related to the climatic and sea level changes of the late-Pleistocene. Within the drainage basins of the Waccamaw, Little Pee Dee, and Great Pee Dee (see Figure 1, and Figures 1-4 in Thom, 1967) Thom has been able to recognize a diversity of surface features (alluvial terraces, stabilized point bars, meander scours, valley dunes, barrier beaches, back-barrier flats, etc.), to map their areal extent, and to suggest their genetic and chronological relationships.

The relationship of fluvial and shoreline features is of particular interest. Thom has found that in each of the valley systems he was able to recognize three separate terrace levels, two (Terraces II and III) apparently relating to interglacial cycles of alluviation and the third (Terrace I) to a Mid-Wisconsin interstadial cycle. The two higher terraces (interglacial) can be traced downstream where they each merge with the lagoonal facies (back-barrier flat) of a Pleistocene shoreline complex. The back-barrier flats in turn grade into sandy barrier complexes. Thus the two highest alluvial terraces are both related to interglacial marine transgressions. Elevational expression, weathering profiles and the like provide an impression of the relative ages of the surfaces. The lowest alluvial terrace has no coastal correlate, as it is apparently related to a marine transgression with a sea level somewhat lower than the present.

In both the Great Pee Dee and Little Pee Dee valleys there are also two stabilized parabolic dune complexes. These two cycles of dunes were built of sands derived from the rivers during periods of degradation. The shape of the dunes suggest that the winds which formed them were from the southwest. The relationship of the dune sheets to the various terrace formations and the superposition of the dune complexes at at least two sites, suggest that (1) they were formed during periods of lower sea level (glacial advance), (2) the oldest dune sheet postdates Terrace II (interglacial) but predates Terrace I (interstadial), and (3) the youngest dune sheet postdates Terrace I.

The spatial relationships of these various Pleistocene features is suggested in Figure 1. The chronological and genetic relationships are indicated in Table 1.

Of interest in this investigation are the relative relationships and dating of Terraces I and II and the two dune sheets. In the Great Pee Dee valley Thom (1967) has found that some dunes have developed across the mouths of meander scars on the Terrace I surface, resulting in the partial burial of Terrace I channels and point bar sediments. This complex of eolian deposits is referred to as Dune Sheet I. In

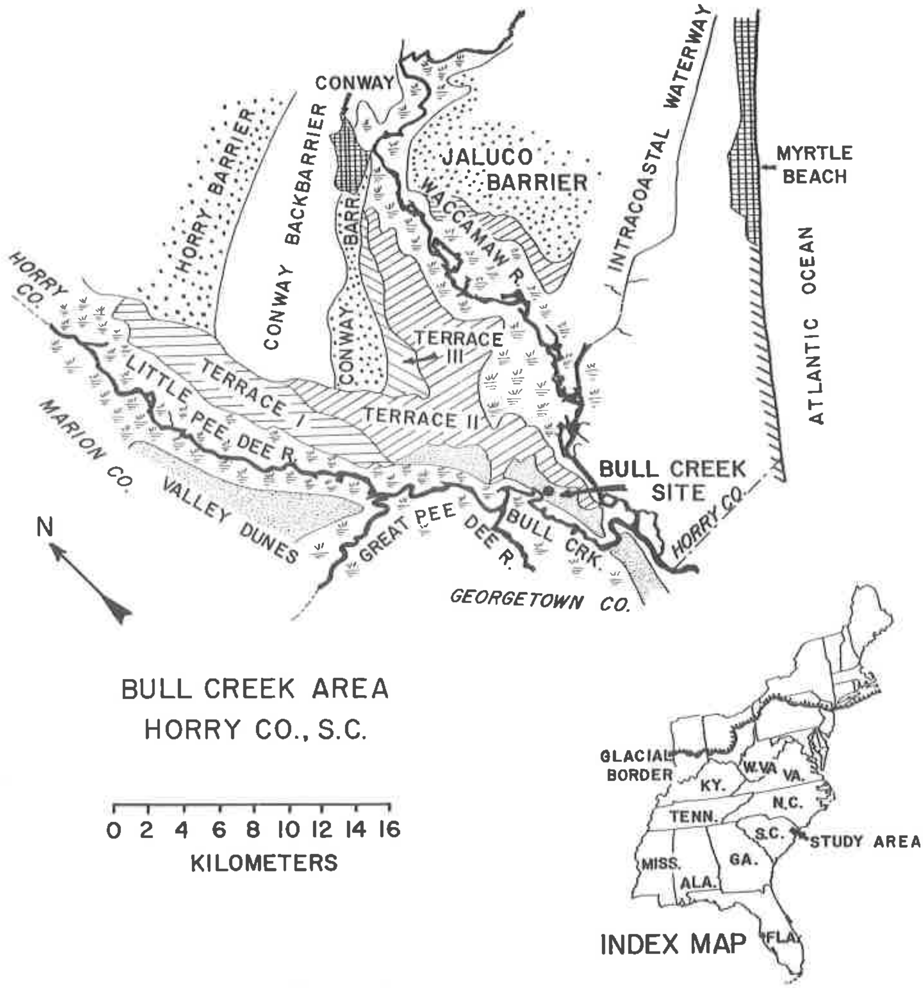

Figure 1. Map of Bull Creek area.

addition, there are a number of places where Terrace I meander scars have cut into an older set of dunes which in turn blanket Terrace II. There are at least two places, one of them Bull Creek (to be discussed) in this paper), where the two dune sheets are directly superimposed with a weathering profile on the surface of the underlying sheet.

The general suggestion is that Terrace II (and the associated barrier and back-barrier facies) was deposited during a high stand of the sea during the Sangamon interglacial. Dune Sheet II was probably formed during the earliest Wisconsin glaciations. Terrace I then developed in response to a new cycle of alluviation stimulated by a rise of sea level (to a maximum some 30 feet lower than the present) during

Table 1. Bull Creek Area--Surface Features and Relationships (from Thom, 1967).

Age	Morphostratigraphic unit		Environments
Quaternary			
L. Recent	Recent		coastal estuarine-deltaic fluvial
		(unconformity)	
E. Recent-L. Wisconsin	Dune Sheet 1		aeolian
		(unconformity)	
M. Wisconsin	Terrace I		fluvial
		(unconformity)	
E. Wisconsin	Dune Sheet 2		aeolian
		(unconformity)	
Sangamon	Myrtle-Terrace II		coastal estuarine-fluvial
		(unconformity)	
Yarmouth? or Aftonian?	Jaluco Conway	Terrace III	coastal estuarine-fluvial
		(unconformity)	
E. Pleistocene-Pliocene-Late Miocene(?)	Horry		coastal estuarine-fluvial

the Mid-Wisconsin (Plum Point-Port Talbot interstadials in the sense of Dreimanis, 1973). During this time interval Dune Sheet II was stabilized and a soil profile developed. During the classical Wisconsin glaciations Dune Sheet I was formed.

The relative chronology and geomorphic interpretations of these events are probably sound, but some questions remain concerning the absolute chronology. One of the major problems is the absence of radiocarbon control and the fact that some of the available dates are questionable.

A case in point is the interesting section exposed along Bull Creek (Figure 1) where the stream has cut into a valley dune complex revealing a section containing a peat at the base (at present river level) and two dune sheets separated by a soil and humate zone. The radiocarbon date from the humate complex at the top of the lower dune system is 16,900±320 (I-1897), the date from the peat is 36,200 (+3600, -2500; I-1898). If one takes these dates at face value, then the younger date is reasonably consistent with the age assignments suggested, but the older date is not. If the peat is in fact on the order of 36,000 years old, then the older dune complex could not date from the earliest Wisconsin. However, there is some reason to suspect the date from the peat (Thom, personal communication). Our own field work at the site demonstrated that the peak is riddled with roots, hence could easily be contaminated with relatively modern carbon.

Some of these problems might be resolved through an investigation of micro-fossils contained in the peat. For example, in the past decade it has become apparent that pollen assemblages from interglacial deposits in the Coastal Plain can, in general, be distinguished from those dating from the Mid-Wisconsin interstadial (e. g., Frey, 1951, 1952, 1953, 1955; Whitehead, 1965, 1967, 1973, 1975a, 1975b; Whitehead and Doyle, 1969; Whitehead and Davis, 1969). Pollen spectra from the interstadial sediments generally suggest a temperate forest complex with a variety of deciduous and coniferous taxa including some constituents of the present "northern hardwoods" forests and a variety of boreal taxa as well. The pollen assemblages from interglacial deposits indicate a forest more "austral" in character (probably very much like the modern forests of the area) with only rare occurrence of boreal taxa.

Accordingly, we feel that pollen analytical studies of the peat exposed at Bull Creek should enable us to suggest whether the peat is interglacial or interstadial. In addition, it should be possible to determine the depositional environment responsible for the peat. This would be of interest, because if the peat formed in a brackish environment, it would suggest a sea level close to or slightly above the present (at present slightly brackish marshes extend inland along the Waccamaw-Great Pee Dee drainage to within 5 km of the Bull Creek site).

This paper presents the results of our palynological study of the Bull Creek peat horizon.

Acknowledgments

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THE BULL CREEK SITE: LOCATION AND MODERN ENVIRONMENTS

Bull Creek is a meandering distributary of the Great Pee Dee River which connects the parent river with the Waccamaw along the border of Georgetown and Horry Counties in northeastern South Carolina (Figure 1). At one point on the east bank a meander has cut deeply into a valley dune complex revealing the complex stratigraphy referred to earlier. The exposure is situated 2.1 km (1.3 mi) west of the town of Bucksport; the coordinates are 33°39'00" North latitude, 79°07'25" West longitude.

The uppermost sediments exposed in the bluff are eolian sands, approximately 3.5-4.6 m (12-15 ft) thick. This sand unit correlates with Dune Sheet I which can be recognized at many points in the Great Pee Dee and Little Pee Dee drainages. The upper sand unit is underlain by another eolian sand complex about 4.3-4.6 m (14-15 ft) thick, the upper meter of which is weakly podzolized. On the uppermost surface of the podzolized zone there is a distinct humate horizon, probably the A_1 horizon of an ancient soil. Thom has correlated the lower sand with Dune Sheet II. The lower sand is in turn underlain by a peaty clay some 1.20 m (4 ft) thick. This unit is variable and tends to grade from peat to clay from top to bottom. It is underlain by a fluvial sand. The base of the peaty clay is at present river level and is in turn about one meter above present mean sea level.

As mentioned previously, there are two radiocarbon dates available from the section, one from the humate separating the two dune sheets and one from the peaty clay at the base. The date from the humate (16,900±320, I-1897) is from a composit of humic materials isolated from the A_1 horizon. The date from the peat is 36,200+3600, -2500 (I-1898).

Although the modern vegetation of the entire region is classified as "southeastern evergreen" forest by Braun (1950), it is actually exceedingly complex. Vegetation type clearly reflects the character of the soil and the availability of water. The forests on fine-grained alluvium along the west shore of Bull Creek are typical southern flood plain forest (e.g., Kuchler, 1964) and are dominated by cypress (Taxodium distichum), water gum (Nyssa aquatica), water ash (Fraxinus caroliniana), and a variety of oaks and other hardwood species. On the eolian deposits above the bluff the vegetation is open in structure and dominated by long-leaf pine (Pinus palustris), loblolly pine (P. taeda), and several scrub oaks. On the finer-grained upland soils (the higher alluvial terraces and the back-barrier flats) one finds a characteristic oak-pine community in which loblolly pine, many different oaks, hickories, sweet gum (Liquidambar styraciflua), black gum (Nyssa sylvatica), and other hardwoods are common. On the peat soils of the Carolina Bays and on a few other wet upland soils a "pocosin" community occurs. This community is typified by evergreen-leaved shrubs (many of them in the Ericaceae), loblolly bay (Gordonia lasianthus), red bay (Persea borbonia), and sweet bay (Magnolia virginiana). A variety of hardwoods and pines occasionally occur in the pocosin vegetation.

Climatically the area is typical of the southeastern Coastal Plain. Weather data from Conway (18 km north of Bull Creek) indicate an annual average temperature of 18°C (64°F), a January average of 9.2° C (48.6°F), and a July average of 26.9°C (80.5°F). The annual precipitation averages 128 cm (50.4 in) with the bulk concentrated in June, July and August (see Thom 1967, Appendix 2).

FIELD AND LABORATORY TECHNIQUES

The peat section was collected with the help of Bruce G. Thom on July 16, 1966. Samples were collected at selected intervals from a clean face excavated on the surface of the peat. Approximately 5 to 10 cc of material were collected from each level and then stored in tightly stoppered vials.

After collection of the pollen samples, the peat was excavated back about 1-1.5 m from the exposed face to determine whether the rootlets observed on the surface were present throughout the peat. All portions of the peat, whether on the surface or some distance back from the exposure, appeared to be riddled with fine roots. This certainly suggests that the radiocarbon date from the peat may be too young.

Qualitative observations were made on modern vegetation both at the site and on the diversity of soil types in the immediate vicinity.

Pollen samples were prepared by a standard procedure (KOH, HCl, HF, acetolysis, mounting in silicone oil) Counts were made with a Wild M20 microscope at a magnification of 400 diameters. Oil immersion and phase were employed for difficult grains. In every case the pollen sum (sum of herbs, shrubs, and trees, but excluding aquatics and pteridophytes) involves at least 500 grains.

INFERENCES FROM THE POLLEN DATA

The results of the palynological investigation are presented in Figure 2 and Table 2. It is evident that the bulk of the pollen sedimented into the peat was locally derived. This is suggested by the very low percentages of arboreal types, the high percentages of shrubs (particularly alder) and herbs (mostly grass, sedge, and composites), and the pollen (or spores) of many plants of wet habitats (e.g., Magnolia, birch, willow, sycamore, holly, buttonbush, Cyrilla, Sphagnum, Sarracenia (pitcher plant), Drosera (sundew), Osmunda cinnamomea and Osmunda regalis, and a variety of aquatics.

Among the tree types oak and pine are the dominants. Many other hardwood types are also represented. Measurements of intact pine grains suggest that no smaller pine types were represented (thus jack and/or red pine were not present; e.g., Whitehead, 1964). No white pine types were noted. Spruce occurs throughout the profile, but the percentages are extremely low and no intact grains were encountered. Among the "boreal" elements that characterize full-glacial,

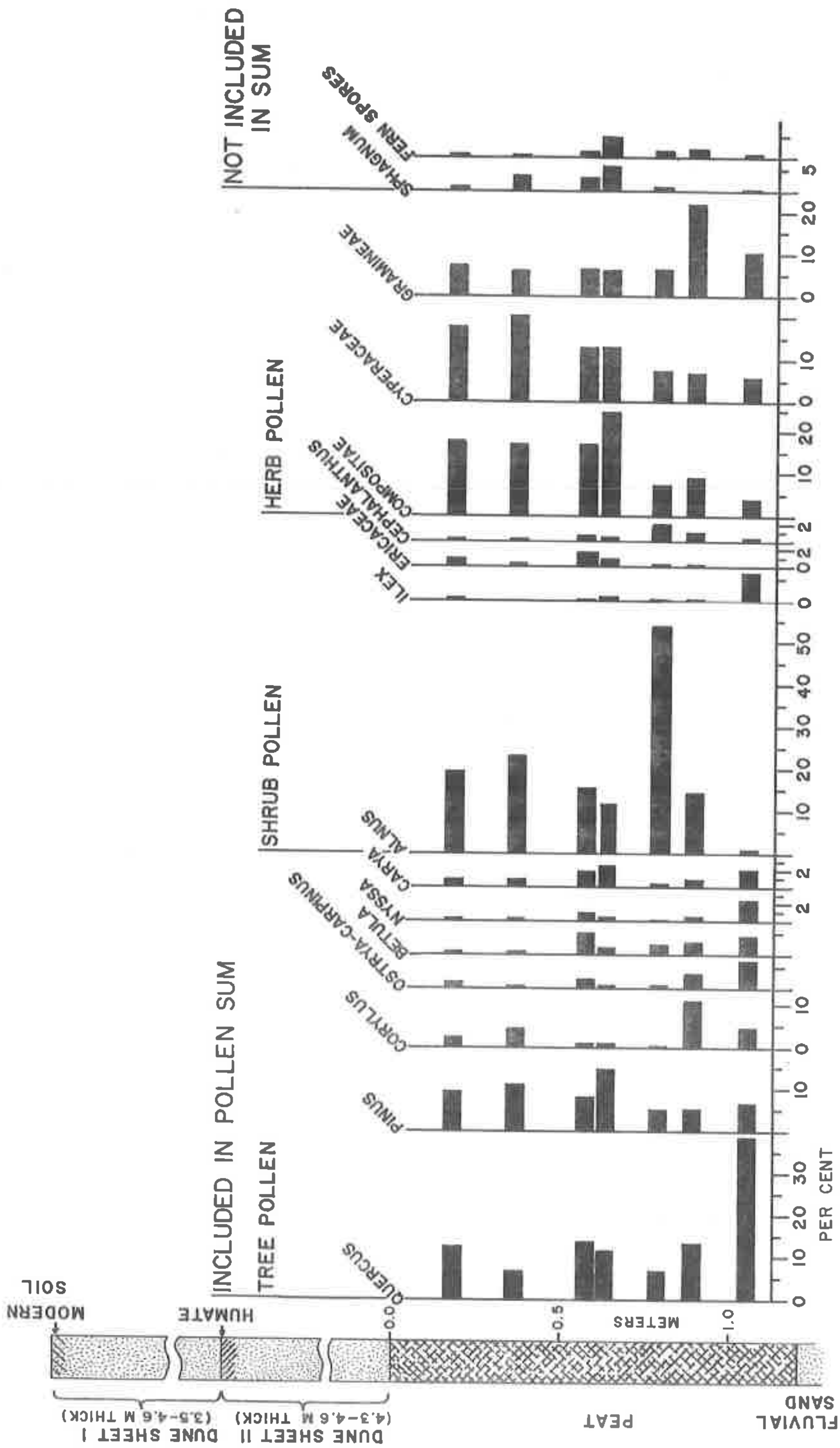

Figure 2. Pollen diagram from Bull Creek Peat.

Table 2. Bull Creek Peat Pollen Percentages.

	Arboreal Pollen								
Depth (cm)	Quercus	Pinus	Corylus	Ostrya-Carpinus	Betula	Nyssa	Carya	Salix	Castanea
18	12.48	9.40	2.42	1.43	0.68	0.34	0.89	0.56	1.32
36	6.94	11.31	4.45	0.69	0.91	0.36	1.04	0.26	0.22
57	13.92	8.27	1.02	2.18	5.14	0.94	1.92	0.38	0.34
63	11.69	14.76	0.91	0.92	1.80	0.54	2.70	0.42	0.30
79	6.85	5.27	0.46	0.88	2.50	0.22	0.53	0.41	0.46
89	13.68	5.32	11.14	3.58	2.96	0.58	0.91	1.23	0.66
105	38.69	6.60	4.61	6.63	4.40	2.56	2.16	0.26	0.15

	Arboreal Pollen (cont.)								
Depth (cm)	Fraxinus	Osmanthus	Fagus	Platanus	"Cupress"	Symplocos	Acer	Liquidambar	Tilia
18	0.22	0.04	-	0.04	0.18	-	-	0.04	0.04
36	0.28	0.08	0.04	-	0.20	-	-	0.04	-
57	0.52	0.10	0.16	0.17	0.34	0.04	0.02	0.07	0.06
63	0.52	0.53	0.08	-	0.04	0.04	-	0.18	-
79	0.05	-	-	0.05	0.04	-	-	-	-
89	0.28	0.15	0.04	-	0.12	0.10	0.02	0.12	-
105	0.24	0.12	0.34	-	0.18	0.17	-	0.43	0.10

	Arboreal Pollen (cont.)					Shrub Pollen			
Depth (cm)	Juglans	Picea	Ulmus	Magnolia	Σ AP	Alnus	Vitis	Ilex	Ericaceae
18	-	0.24	0.14	0.44	33.14	19.40	0.63	0.65	1.00
36	-	-	0.15	0.38	28.49	23.15	0.70	0.25	0.42
57	-	0.07	0.07	0.20	35.99	15.85	0.82	0.56	1.78
63	-	0.08	0.04	0.22	35.84	11.98	0.24	1.16	1.15
79	0.04	0.15	0.10	0.02	17.80	54.00	0.47	0.30	0.36
89	0.06	0.05	0.06	-	41.08	14.58	1.51	0.42	0.32
105	-	0.10	0.10	-	68.88	1.12	0.06	6.71	-

	Shrub Pollen (cont.)							
Depth (cm)	Cephalanthus	Cyrilla	Rhus	Myrica	Ceanothus	Viburnum	Aralia	Sambucus
18	0.36	0.94	-	-	-	0.05	-	-
36	0.35	0.43	0.18	-	-	0.14	-	-
57	0.74	2.40	0.04	0.05	0.18	0.48	-	0.09
63	0.61	1.73	-	0.10	0.26	0.68	-	-
79	2.14	0.81	0.04	0.05	0.22	0.24	-	-
89	1.03	0.86	0.62	0.37	0.98	0.33	0.04	-
105	0.40	0.43	0.18	0.15	-	0.26	-	-

late-glacial and to a lesser extent, Mid-Wisconsin interstadial sediments from the Southeast, only spruce, Schizaea pusilla (the curly grass fern), and Lycopodium lucidulum (a single spore) are represented. The latter two taxa are exceedingly rare.

The data are consistent with (1) a fresh-water depositional environment and (2) deposition during an interglacial rather than the Mid-Wisconsin Plum Point-Port Talbot interstadial (e. g. , Whitehead, 1973, 1975a).

Table 2. Continued.

Depth (cm)	Shrub Pollen (cont.) Rhus Toxicodendron	Σ SP	Herb Pollen Compositae	Cyperaceae	Gramineae	Rosaceae	Umbelliferae
18	0.04	23.12	17.80	17.87	7.54	0.19	0.62
36	-	25.68	17.20	20.44	6.20	0.10	0.06
57	-	22.74	17.21	13.34	6.79	0.40	0.24
63	-	17.88	24.64	13.14	6.22	0.77	-
79	-	58.62	7.15	7.45	6.41	0.26	0.58
89	-	18.58	9.04	6.92	22.11	0.36	0.52
105	-	9.44	4.00	5.99	10.35	0.36	0.24

Depth (cm)	Herb Pollen (cont.) Ambrosia	Caryophyllaceae	Cheno-Am.	Polygonum § Persicaria	Plantago	Labiatae	Thalictrum
18	0.68	0.05	0.14	-	-	0.18	0.14
36	0.76	0.04	0.28	-	-	0.16	0.11
57	0.86	0.02	0.29	-	-	0.04	0.24
63	0.61	-	0.20	-	-	0.28	0.26
79	0.42	-	0.19	-	-	-	0.05
89	0.59	0.02	0.69	0.05	0.60	0.08	0.25
105	0.26	-	-	0.15	-	-	0.19

Depth (cm)	Herb Pollen (cont.) Artemisia	Impatiens	Gentiana	Liliaceae	Sarracenia	Pyrola	Cruciferae
18	0.18	-	-	0.10	0.04	0.42	0.05
36	0.04	-	-	-	-	0.30	0.26
57	0.08	-	-	-	-	0.32	0.10
63	0.14	-	-	0.08	0.04	0.12	0.08
79	-	0.05	-	0.04	0.10	-	-
89	0.03	0.02	0.06	-	-	-	-
105	-	-	-	-	-	-	-

Depth (cm)	Herb Pollen (cont.) Drosera	Σ HP	Aquatics (not included in pollen sum) Nymphaea	Nuphar	Sagittaria	Proserpinaca	Isoëtes	Myriophyllum
18	-	45.96	0.36	-	0.74	-	0.24	0.97
36	-	45.90	-	-	0.20	0.11	-	1.14
57	0.02	41.02	0.23	0.04	0.06	0.02	0.02	0.06
63	-	30.85	0.92	-	0.16	0.04	0.04	-
79	-	23.28	-	-	0.04	0.08	0.10	-
89	-	40.32	0.09	0.02	0.18	0.07	0.09	-
105	-	21.69	0.06	0.06	-	-	-	0.18

The fresh-water environment is indicated by the types of shrubs, herbs, and aquatics that are present. It is true that a few of the aquatic types (e.g., Nuphar, Sagittaria) occasionally occur in brackish environments, but the others apparently do not (Radford et al., 1964). In addition, if the depositional environment had been brackish, we would expect far higher percentages of "cheno-am" pollen, lower percentages

Table 2. Continued.

Depth (cm)	Aquatics (cont.) Eriocaulon	Orontium	Pteridophytes (not included in pollen sum) Monolete Fern	Trilete Fern	Osmunda regalis	Osmunda cinnamomea	Polypodium	Schizaea pusilla
18	0.14	2.16	0.36	-	0.14	-	-	-
36	0.06	-	0.44	-	-	0.04	-	0.13
57	-	0.20	1.05	0.02	0.12	0.08	-	-
63	-	0.22	4.82	-	0.62	0.24	-	-
79	-	0.08	1.02	-	1.01	0.08	-	-
89	-	-	1.81	-	0.30	0.04	0.02	0.06
105	-	0.18	0.34	-	0.58	0.06	-	-

Depth (cm)	Pteridophytes (cont.) Lycopodium (unknown)	Lycopodium inundatum-alopecuroides	Pteridium	Lycopodium lucidulum	Miscellaneous Sphagnum	(not included in pollen sum) Unknown	Unidentifiable
18	-	-	1.09	0.04	1.14	0.62	0.30
36	0.11	-	-	-	3.74	0.30	-
57	0.02	-	-	-	3.18	0.30	1.10
63	-	-	-	-	5.92	0.32	1.00
79	-	0.04	-	-	0.60	0.19	0.96
89	0.06	-	-	-	-	0.12	1.00
105	-	-	-	-	0.10	0.78	1.77

of sedge, and pollen of typical salt marsh plants such as Iva frutescens. The presence of many typical riparian taxa (alder, buttonbush, birch, black gum, willow, ash, sycamore, "Cupressaceae," elm, magnolia, etc.), plus plants of boggy habitats and aquatics argues for deposition in a wet alluvial environment, conceivably in the slack water of a meander loop. The presence of considerable sand in the peaty clay suggests that there was periodic flooding.

The proximity of "pocosin" or evergreen bog-shrub vegetation, perhaps along the stream bank, is indicated by the pollen of magnolia, Ericaceae, Myrica, holly, and especially Cyrilla. Cyrilla is especially interesting, as work in southeastern North Carolina has shown that even in areas where it is extremely abundant, very little pollen is encountered in the modern pollen rain (Whitehead and Tan, 1969). The highest percentage of Cyrilla is 2.40 percent which indicates that it was overwhelmingly abundant locally. The presence of such peaty and/or acid habitats is also suggested by the occurrence of Sphagnum spores, pollen of pitcher plants and sundew, and the emergent aquatic Orontium (which generally occurs in reasonably acid waters).

The interglacial correlation is suggested by the general character of the pollen flora. Boreal elements are extremely rare. As mentioned previously, only spruce, Schizaea, and Lycopodium lucidulum occur and they are very uncommon. In addition, their presence must be interpreted with caution as there is always the possibility that they were water-transported for some distance. Bull Creek is part of the Great Pee Dee drainage which has its headwaters in the Blue Ridge Mountains of North Carolina (as the Yadkin River). Much of the pollen flora is quite "austral" in character (Magnolia, Cyrilla, Osmanthus, Symplocos) indicating conditions much like the present.

This tentative age assessment is strengthened by comparison with known interglacial and interstadial pollen assemblages from the Southeast. Relevant interglacial spectra are known from the Horry Clay near Myrtle Beach (Frey, 1952; Whitehead, unpublished data), the Flanner Beach locality on the Neuse River in North Carolina (Whitehead and Davis, 1969), from peats at the Lee Creek phosphate mine near Aurora, North Carolina (Whitehead, 1975b), and from the Kempsville Formation in southeastern Virginia (Whitehead, unpublished data). Mid-Wisconsin interstadial pollen assemblages are available from northwestern Georgia (Watts, 1973), from peats in an exposure along the intracoastal waterway near Long Beach, North Carolina (Whitehead and Doyle, 1969), from the Bay Lakes profiles in Bladen County, North Carolina (Frey, 1951, 1953, 1955; Whitehead, 1964, 1965, 1967), and from Rockyhock Bay in Chowan County, northeastern North Carolina (Whitehead, 1975a).

It is quite apparent that the greatest similarity of the Bull Creek spectra is with interglacial material. There are obvious differences in the representation of many pollen types, but almost all of these differences are attributable to the nature of the depositional environments (which in every case favor over-representation of plants growing locally) and local edaphic situations. The clear inference is of a vegetation and climate quite comparable to the present.

The Bull Creek assemblage differs from interstadial assemblages in a number of important ways. First, boreal elements (such as spruce, fir, northern pines, Sanguisorba canadensis, Schizaea pusilla, Arceuthobium, and several species of Lycopodium), although not frequent in the interstadial materials, are more common than at Bull Creek. The interstadial materials from both southeastern and northeastern North Carolina have some small pine pollen (hence jack and/or red pine), white pine types, plus pollen of some taxa now found in modern "northern hardwoods" forest (e. g. , Whitehead, 1975a).

It is probably significant that the interstadial spectra from both the Bay Lakes profiles and from Rockyhock Bay are quite similar to spectra dating from the transition from the late-glacial to the postglacial from the same profiles. The Bull Creek spectra bear little resemblance to late-glacial or early postglacial assemblages, but rather appear like modern ones from alluvial situations in the Coastal Plain. Distinctive "austral" types do not generally occur in the interstadial materials.

Thus the relative lack of boreal or northern hardwoods types, the presence of typical "southern" taxa (Magnolia, Symplocos, Cyrilla, etc.), the similarity of the overall peat assemblage to modern and interglacial assemblages, and the distinct contract with interstadial materials suggest that the Bull Creek Peat was deposited during interglacial time. Although it is impossible to indicate which interglacial is involved on palynological grounds, geomorphological data suggest the Sangamon (Thom, 1967).

Given these paleoecological assessements, it would appear that the radiocarbon date from the peat is far too young and that Thom's (1967) reconstruction of events for the section (and for the Pee Dee drainage) is correct: (1) Dune Sheet II formed during early Wisconsin glaciations; (2) the dune complex was stabilized and a soil profile developed during the Mid-Wisconsin Plum Point-Port Talbot interstadial; and (3) Dune Sheet I developed during the classical Wisconsin glaciations; and (4) Dune Sheet I stabilized and the modern soil profile formed during the postglacial.

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