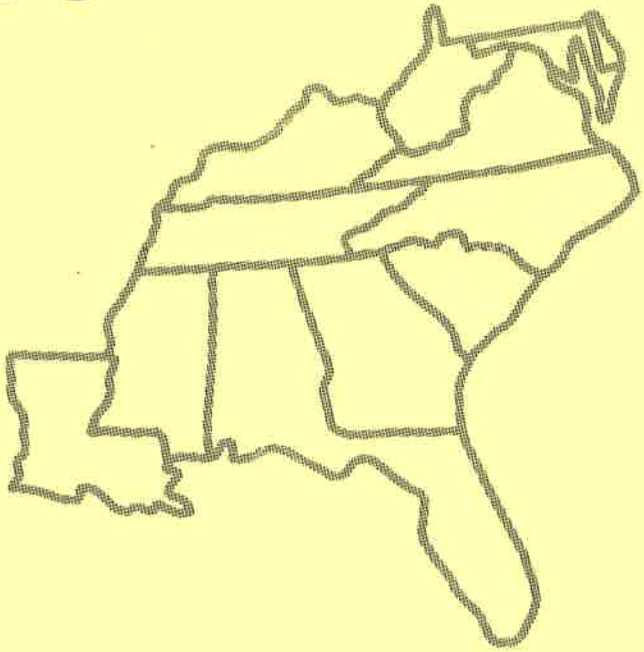
SOUTHEASTERN GEOLOGY



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IN-SITU STRESS AND ITS RELATIONSHIP TO JOINT FORMATION IN THE TOXAWAY GNEISS, NORTHWESTERN SOUTH CAROLINA

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ABSTRACT

In-situ stresses have been measured and the joint systems have been studied in a 511 m long exploratory tunnel in Toxaway Gneiss. It is concluded that in-situ stresses have a direct relationship to joint formation.

The maximum principal stress (σ_1) is compressive, oriented N57°E and has a magnitude of 29.3 MPa (4253 psi). The intermediate principal stress (σ_2) is compressive, oriented N32°W and has a magnitude of 18.4 MPa (2675 psi). The least principal stress (σ_3) is subvertical and has a magnitude of 10.2 MPa (1476 psi).

The dominant joint sets are approximately east-west, northwest (N45°W), and northeast (N60°E). The northwest and east-west sets have steep dips, while the northeast set dips moderately northwest. All of the sets are mineralized to some degree, but the northeast and particularly the east-west sets contain a greater percentage of mineralized joints than the northwest set. The number of continuous and discontinuous joints per unit distance diminishes rapidly beyond 183 m (Station 6+00) into the tunnel. This phenomenon is most evident among the more discontinuous northeast and northwest joint sets. The ratio of mineralized to nonmineralized joints increases toward the end of the tunnel. Disappearance of jointing with depth along with the interpretation of the in-situ stress data suggests that joints in the Toxaway Gneiss formed at different times. The youngest joints propagated from preexisting incipient fractures oriented subparallel to earlier tectonically formed joints due to the release of stored elastic strain energy

associated with the in-situ stresses. At depths of 100 m the overburden pressure is great enough to prevent the release of the stored elastic strain energy.

INTRODUCTION

As the major part of the subsurface exploratory program for Duke Power Company's Bad Creek Pumped Storage Project a pilot tunnel was driven into the proposed powerhouse location. The proposed project is a 1000 MW, hydroelectric pump storage station consisting of four reversible pump turbines located in a powerhouse 200 m underground. The purpose of the pilot tunnel was to explore the geology, measure the behavior of the rock mass in response to tunneling, and to conduct various in-situ tests. The results of these measurements and tests are being used as a basis for the design of the underground chambers and tunnels of the project.

The 511 m pilot tunnel follows the alignment of what will be the access tunnel and turns into the top of the powerhouse location (Figure 1). The tunnel section is a 3 m straight legged horseshoe with an enlarged test station in the powerhouse area. Numerous measurements of geologic features were made. The entire tunnel was mapped on a 1:240 (1 inch = 20 feet) scale and the powerhouse section was mapped in detail at a scale of 1:60 (1 inch = 5 feet). Figure 2 is a profile of the pilot tunnel showing the layout of the tunnel and the position of the land surface.

Acknowledgments

The authors would like to thank G. T. Christenbury for typing the manuscript and W. A. Hedrick for drafting the various figures. Photographs were provided by the Duke Power Construction Department. Critical review by D. T. Secor, Jr., N. J. Gilbert, D. N. MacLemore, III, and D. R. Privett has improved the manuscript considerably. Discussions with D. T. Secor, Jr., concerning the mechanisms of joint formation were most helpful. However, the authors are responsible for the interpretations presented in this paper.

REGIONAL GEOLOGY

The Bad Creek Pilot Tunnel is located immediately northwest of the Brevard Zone within the Toxaway Dome (Figure 1). The Toxaway Dome consists of a core of banded granitic gneiss known as the Toxaway Gneiss and a sliver of Tallulah Falls Formation. The dome is flanked on all sides by rocks of the Tallulah Falls Formation. The Toxaway Dome is an elongate feature that has a steeply dipping northwest limb and a more moderately inclined southeast limb. At the ends,

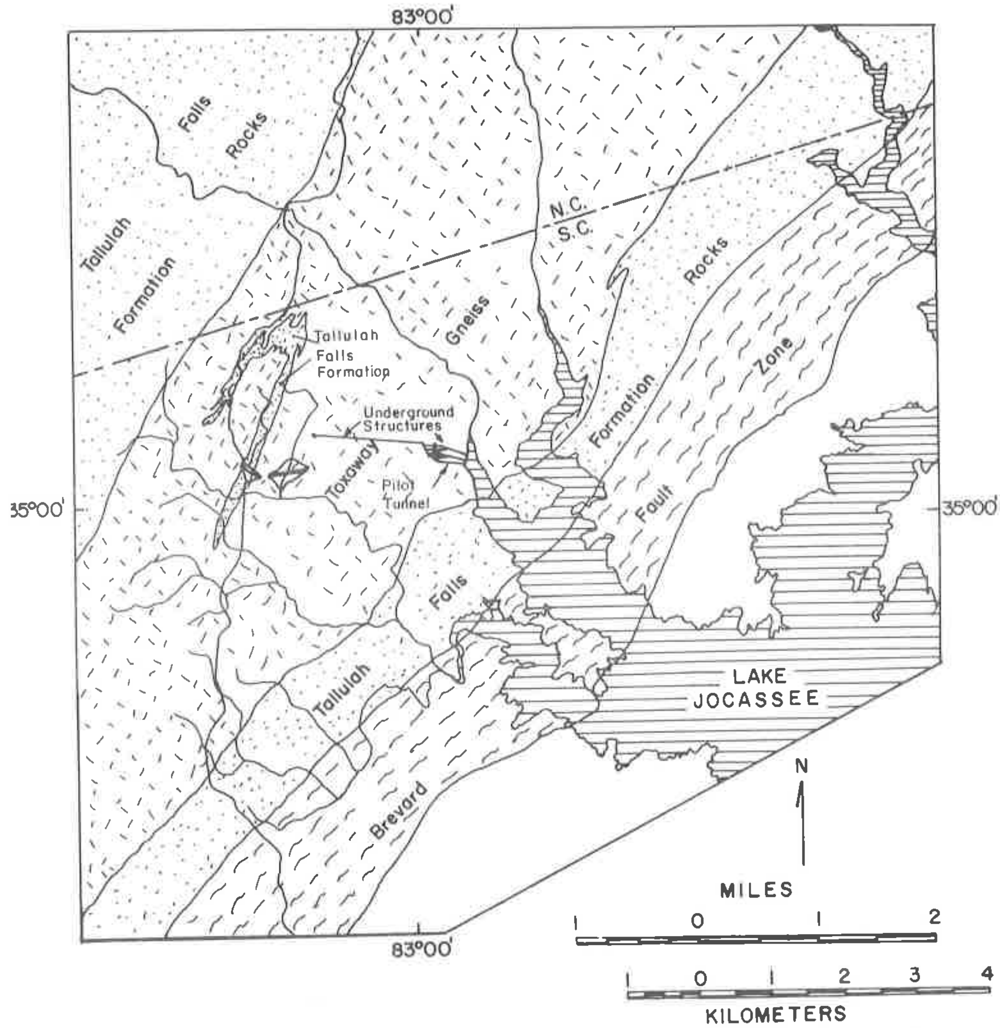


Figure 1. Location of Bad Creek Pilot Tunnel and generalized geologic map of the surrounding area.

the structure plunges gently northeast and southwest, resulting in a structural dome defined by upward arching of the S_1 foliation.

At least two episodes of flowage folding have been recognized in the Toxaway Dome by Hatcher (1977). The first set is isoclinal and recumbent, trending east to northeast, and verging north to northwest. The second set is more upright, isoclinal to open, trending northeast, and verging northwest. Later mesoscopic crenulation cleavage and macroscopic northeast- and northwest-trending folds are also present (Hatcher, 1977). The dominant northeast outcrop pattern of the Toxaway

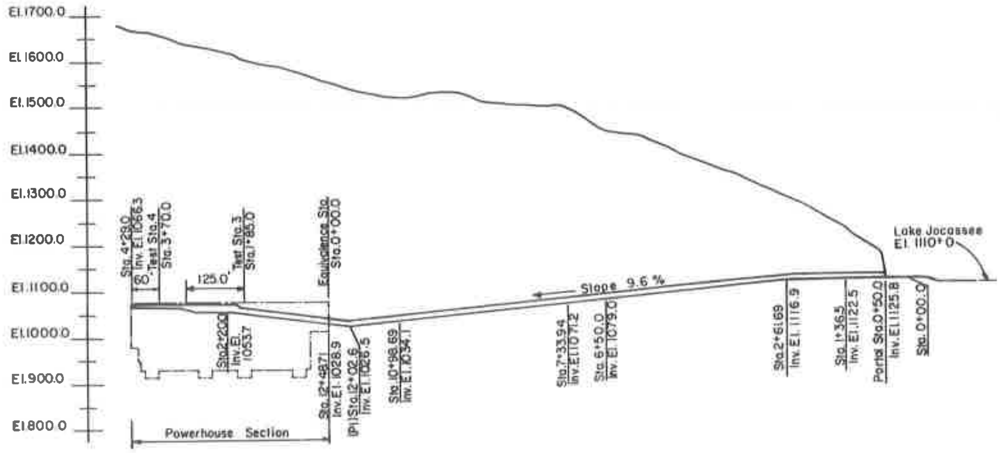


Figure 2. Profile of Bad Creek Pilot Tunnel and overlying topography. Stations in feet.

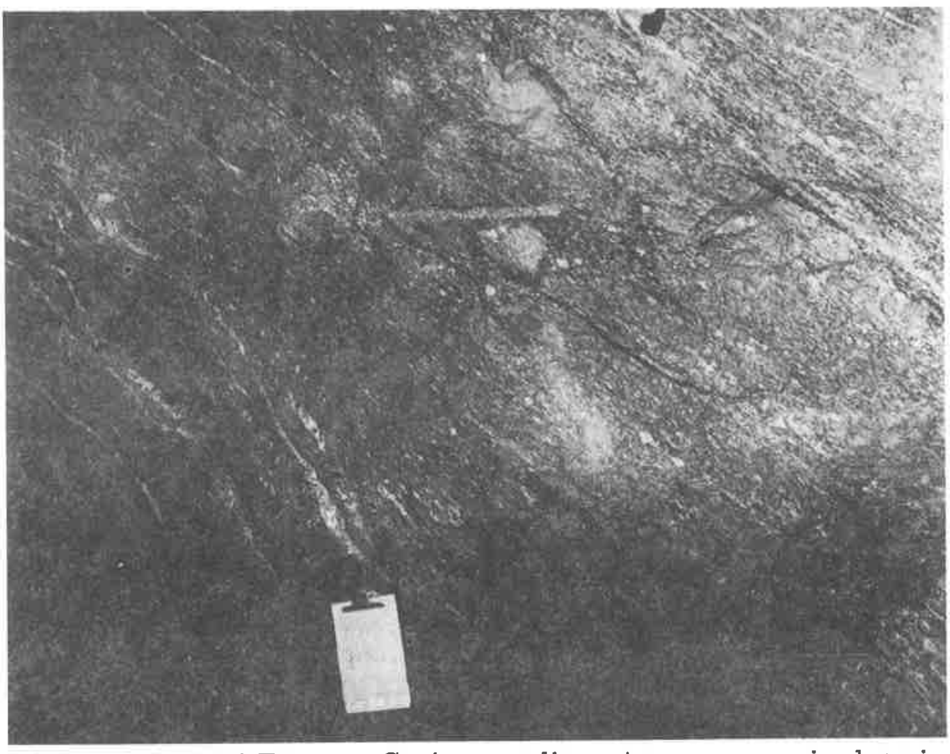


Figure 3. Banded Toxaway Gneiss, medium- to coarse-grained gneiss composed of light quartz-feldspar rich bands and dark biotite-quartz-feldspar bands.

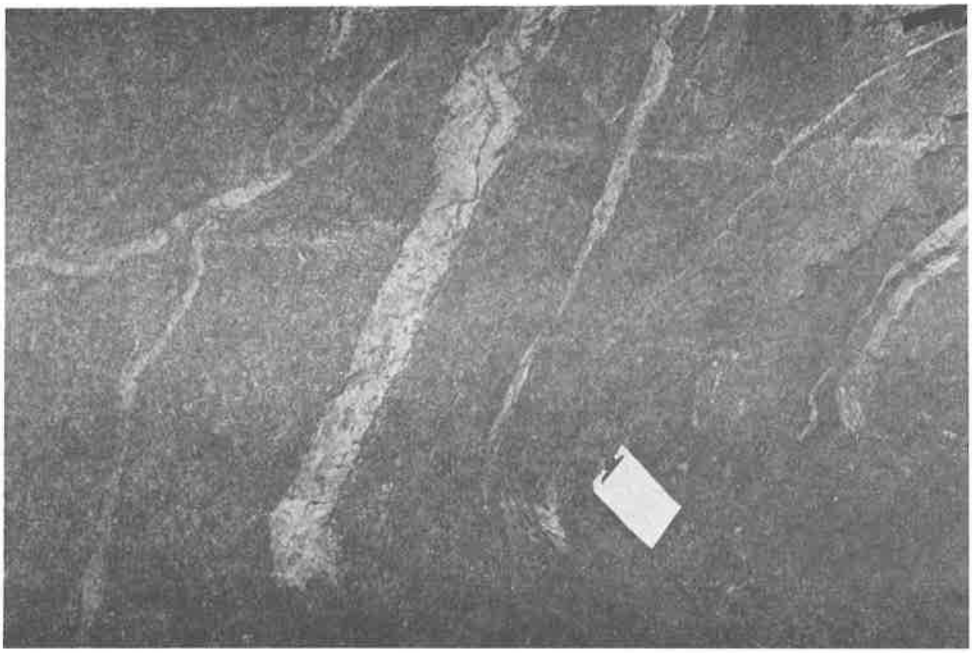


Figure 4. Augen Toxaway Gneiss, coarse-grained, poorly foliated feldspar-quartz-biotite gneiss with augen up to 3 cm across of feldspar or rare hornblende.

Gneiss and Tallulah Falls Formation is produced by the northeast trending structures, principally F_2 folds, but also F_4 folds (Hatcher, 1977).

The principal rock unit in the pilot tunnel is the Toxaway Gneiss. The Toxaway Gneiss can be divided into two types: 1) a banded, medium- to coarse-grained gneiss composed of alternating light quartz-feldspar rich bands and dark biotite-quartz-feldspar bands (Figure 3), and 2) an augen gneiss consisting of a coherent, coarse, massive, poorly foliated feldspar-quartz-biotite gneiss with feldspar and locally hornblende augen up to 3 cm across and a medium- to coarse-grained quartz-feldspar-biotite gneiss with a more distinct foliation containing feldspar augen up to 1 cm (Figure 4). Several small biotite-hornblende schist sills or dikes are present in the pilot tunnel. Their thicknesses range up to 5 m and orientation is parallel to the foliation in the Toxaway Gneiss. The biotite-hornblende schist is even-grained and without distinctive bands or other zones of differing composition. At least two generations of quartz-feldspar-mica pegmatite occur within the pilot tunnel in the Toxaway Gneiss. They are distinguished by the fact that the later generation is undeformed except by fracturing while the earlier is folded. Most of the early pegmatites parallel foliation while the later generation cuts across foliation. Small quartz veins fill some of the early fractures, but are not abundant.

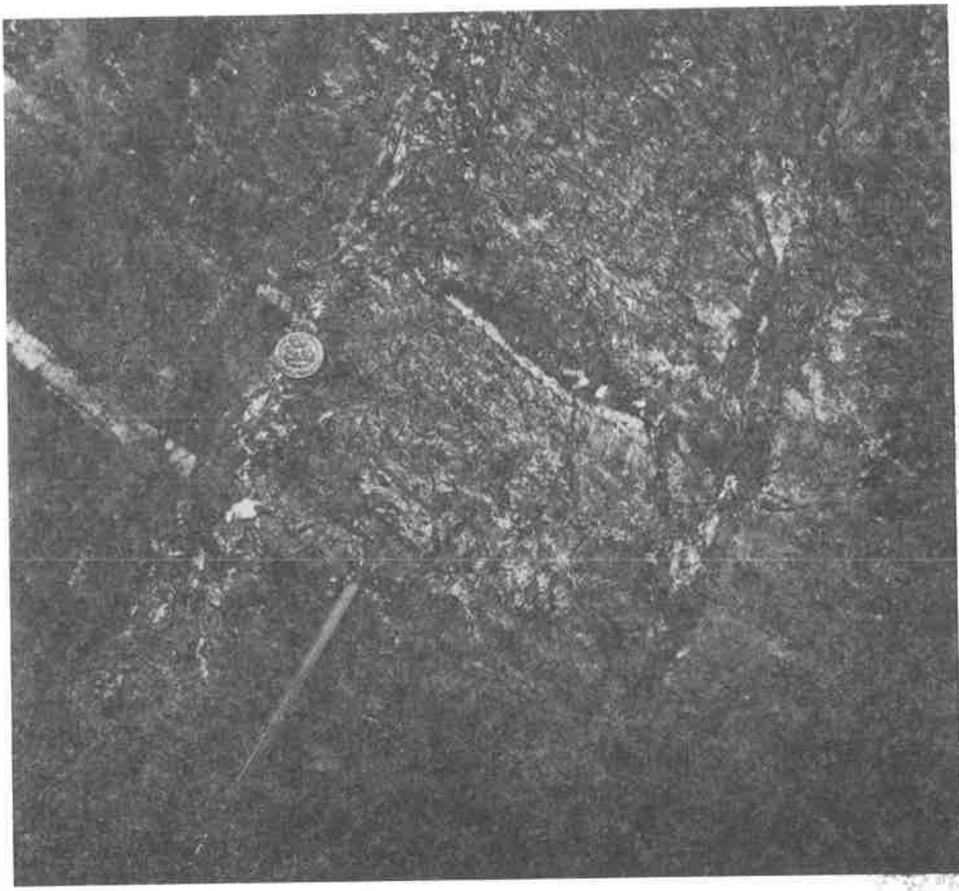
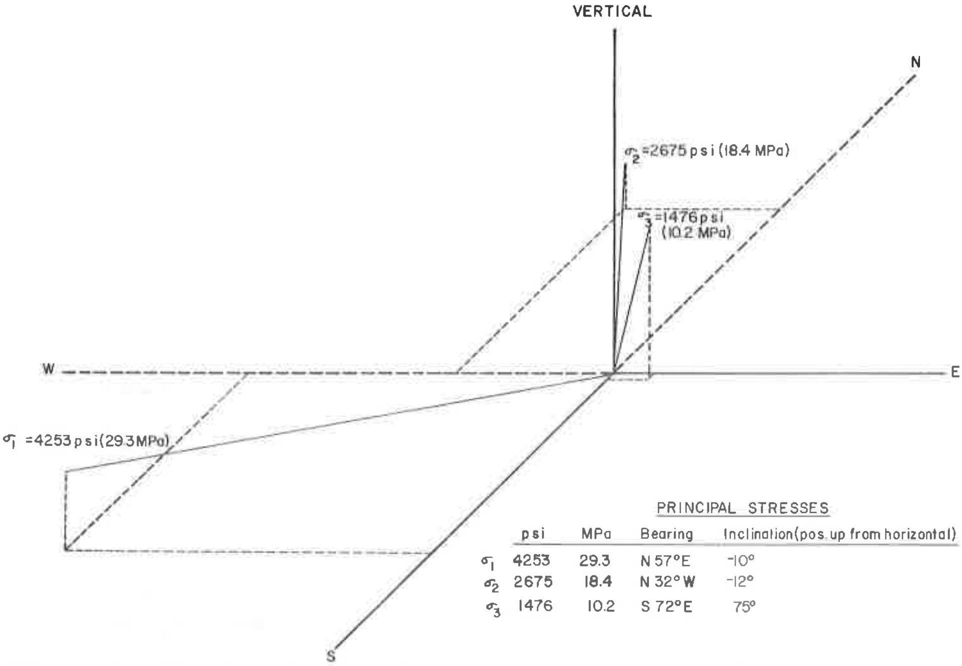


Figure 5. Shear zone, note disrupted foliation, shears generally healed by quartz, feldspar, and/or chlorite.

Structural features in the tunnel include foliation, minor folds, shears, and joints. Foliation in the Toxaway Gneiss is defined by parallel orientation of platy minerals and by layers of differing composition. The average orientation of foliation is N35°E with a dip of 30°SE. Minor folds are present, some of which fold the foliation while others lie within the foliation. Zones where many folds occur are found in several places throughout the tunnel but are confined to the banded Toxaway Gneiss. Several small shear zones have apparent offsets of up to one meter and altered zones associated (Figure 5). These small shear zones are generally filled with chlorite, chlorite-epidote, or chlorite-epidote-calcite-quartz. Several joint sets have been recognized and will be discussed in detail following the discussion of the in-situ stress data.



PRINCIPAL STRESSES

	psi	MPa	Bearing	Inclination (pos. up from horizontal)
σ_1	4253	29.3	N 57°E	-10°
σ_2	2675	18.4	N 32°W	-12°
σ_3	1476	10.2	S 72°E	75°

Figure 6. Graph and table of in-situ stresses determined by the over-coring stress relief technique.

IN-SITU STRESS DATA

In-situ stress measurements using the overcoring stress relief technique were made in the powerhouse section (between stations 0+00 and 4+29, Figure 2) of the pilot tunnel approximately 200 m below the ground surface. The test procedure is similar to that described by Hooker and Bickel (1974). The basic procedure consists of drilling a 38 mm (1.5 inch) diameter hole, inserting a borehole gage which can measure the diameter change on three axes in microstrains, i.e., 2.54 x 10^{-5} mm (1 x 10^{-6} inches). The small hole is then overcored with a 152 mm (6 inch) bit. The in-situ stresses are relieved as the larger bit passes over the borehole gage causing the diameter of the small hole to change. The changes are recorded and, when analyzed with data from overcore holes drilled in different orientations, will yield the in-situ stress ellipsoid for the rock mass. The computer program used to determine the stress ellipsoid was developed by the Bureau of Mines based on work by Panek (1966).

The magnitude and direction of the in-situ stresses determined by the over-coring technique are as follows: the maximum principal stress (σ_1), 29.3 MPa (4253 psi) at N57°E, the intermediate principal stress (σ_2), 18.4 MPa (2675 psi) at N32°W, and the least principal stress (σ_3), 10.2 MPa (1476 psi) subvertical. All stresses are compressive. Figure 6 is a plot and listing of the principal compressive stress magnitudes and directions.

Other stress determinations in the Bad Creek vicinity have been made by Haimson (1975) using the hydrofracturing technique in a core hole on the Bad Creek site and by Talwani (1977) based on composite fault plane solutions for earthquake activity in the Jocassee reservoir area. Haimson's (1975) measurements were made approximately 80 m west of and 30 m deeper than the overcoring measurements in the pilot tunnel. He determined that the orientation of the maximum principal stress (compressive) is N60^{o}E. His magnitudes (σ_1=22.75 MPa (3300 psi), σ_2=15.86 MPa (2300 psi) and σ_3=6.21 MPa (900 psi)) are lower than those obtained in this study. The difference between Haimson's (1975) values and those reported in this paper is probably because he assumed a vertical stress component equal to that of the overburden. At the overcoring test elevation this value would be about 4.76 MPa (690 psi). The stress measured by overcoring of 10.2 MPa (1476 psi) is more than twice this value. If this adjustment is taken into account along with his computed errors (σ_1, $\pm$5.52 MPa (800 psi) and σ_2, $\pm$2.41 MPa (350 psi)) the results are similar. Talwani (1977) inferred from the composite fault plane solutions that the maximum compressive stress is oriented northwest-southeast, or about 90 degrees different from the direction obtained by overcoring and hydrofracturing. Based on revised data, Talwani (personal communication) states that the maximum principal stress is oriented northeast-southwest. The orientation of the maximum principal stress determined by overcoring, hydrofracturing, and composite fault plane solutions is subparallel to Blue Ridge structure and is similar to results reported by Hooker and Johnson (1969) from overcoring data for other portions of the Appalachians.

Excess stresses above those expected for gravitational loading could be the result of elastic strains locked in during tectonic events and only partially released by natural processes or from present-day straining of the earth's crust. Voight (1969) has suggested that convective flow associated with the spreading of the Mid-Atlantic Ridge could produce horizontal compressive stresses within the North American plate if the plate is moving more slowly than the mantle. Sbar and Sykes (1973) favor an east-west trend for compressive stresses. Their conclusions are derived principally from fault plane solutions of earthquakes, in-situ stress measurements, and geologic observations. They note that stress measurements in the Appalachians may vary from the east-west direction due to modifications by local effects such as structure, topography, and uplift.

To arrive at the present stress directions in the Toxaway Gneiss based on a tectonic stress component it would be necessary for the northwest-southeast component, which was the maximum stress direction during deformation, to decrease such that the northeast-southwest component would become the maximum compressive stress direction. This could be produced by release parallel to the northeast fold axes in the Toxaway Dome following the major tectonic event. The stress

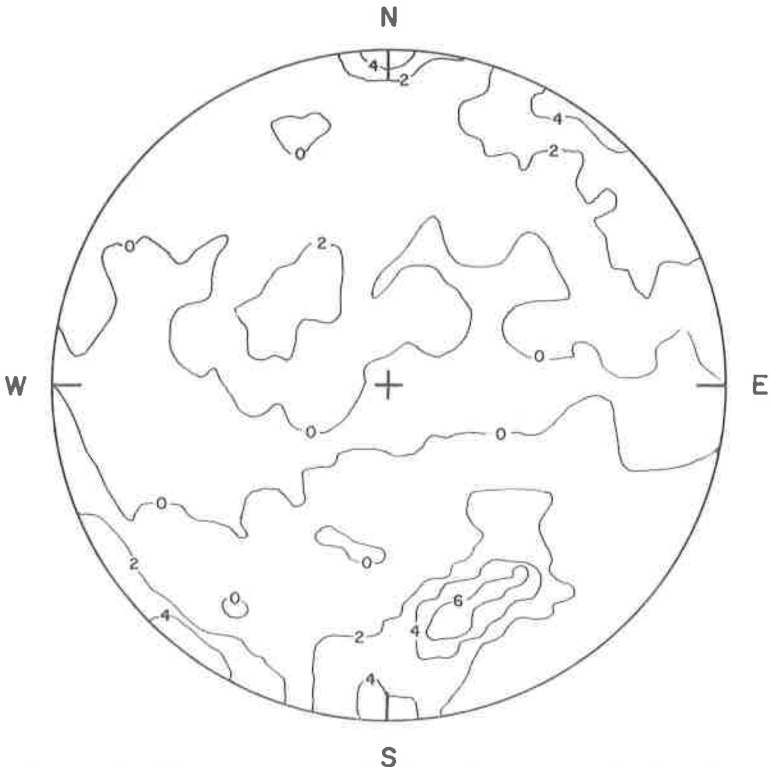


Figure 7. Equal area projection (lower hemisphere) of poles to planes of discontinuous joints. 377 observations.

directions could also be explained by Voight's (1969) hypothesis if the stress field has been modified by local effects.

JOINT DATA

Previous work by Acker and Hatcher (1970) in the Chauga belt and the Inner Piedmont belt indicates that major joint sets trend N30°-50°E and N40°-50°W. These sets tend to be parallel and perpendicular to the regional structural trend. Measurements of joints in the Inner Piedmont by Griffin (1973) indicate three major joint sets oriented N36°W, N45°-55°E, and N85°E, respectively.

The present investigation established three prominent joint sets in the Toxaway Gneiss: 1) a predominant east-west set that varies between N70°W and N70°E with steep north and south dips, 2) a N60°E set with moderate to steep northwest dips, and 3) a N45°W set with steep southwest dips (Figures 7 and 8). All of the above sets show some degree of mineralization, but the northeast and particularly the

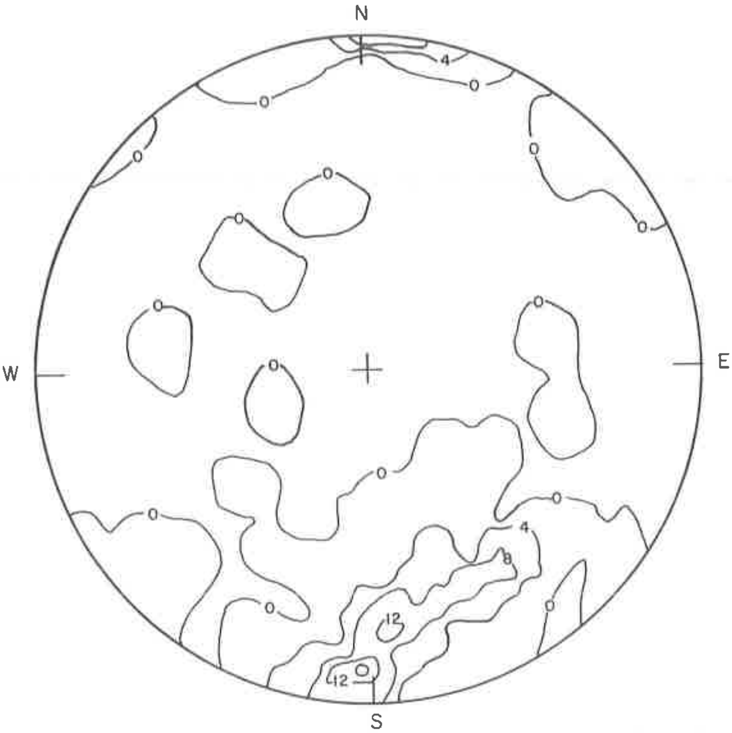


Figure 8. Equal area projection (lower hemisphere) of poles to planes of continuous or prominent joints. 127 observations.

east-west set contain a greater percentage of mineralized joints (Figure 9. The dominant mineral fillings are quartz, chlorite, epidote, and calcite in various combinations.

The determination of the continuity of joints in this study was based on whether a single joint plane could be traced across the approximately 3 m diameter tunnel. Figure 7 is a plot of all discontinuous joints and Figure 8 a plot of all prominent or continuous joints measured in the pilot tunnel. The orientation of discontinuous joints is similar to that of the continuous joints. There is more scatter in the data for the discontinuous joints. During the course of the investigation it was noted that many unfilled fractures were discontinuous while filled or mineralized joints were generally continuous. Since the orientations of the discontinuous joints appear to be consistent, they have been interpreted to be systematic joints produced in a regional stress field. The joint data are biased because of tunnel orientation; joints parallel to the tunnel alignment are not readily observed.

The number of joints decreases toward the end of the tunnel as the amount of overburden increases (Figures 2 and 10). There is a

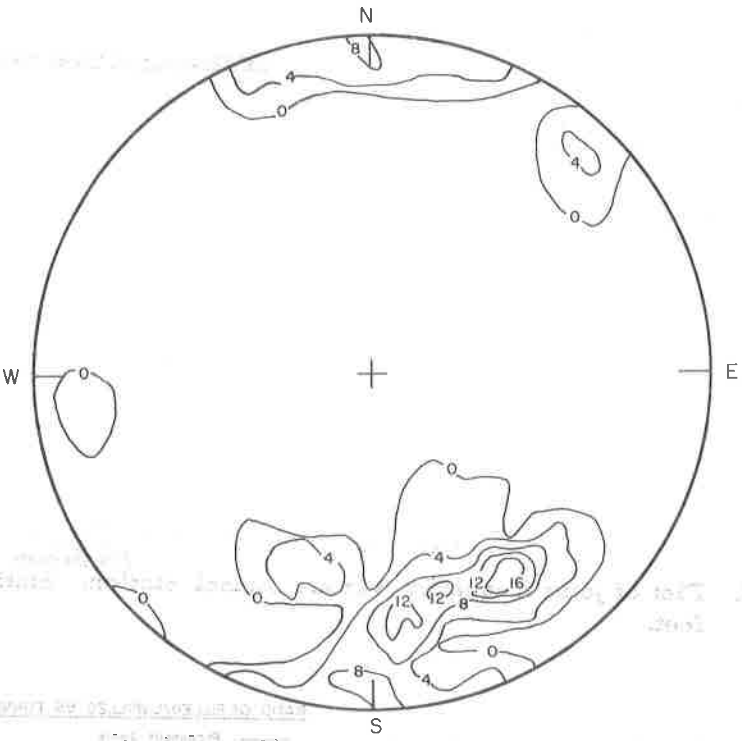


Figure 9. Equal area projection (lower hemisphere) of poles to planes of mineralized joints. 44 observations.

rapid decrease in joint frequency near Station 6+00 which roughly corresponds with the slight break in slope above the pilot tunnel (Figure 2). From the portal to near Station 6+00 the tunnel is in banded Toxaway Gneiss, from Station 6+00 to Station 11+50 in augen Toxaway Gneiss, and from Station 11+50 to the end of the tunnel in banded Toxaway Gneiss. The rapid decrease in joint frequency at Station 6+00 and the increase in joint frequency near Station 12+00 may be partly due to the change in rock types and their associated physical properties. Neglecting the effects of rock type, there is a decrease in joint frequency toward the end of the tunnel.

The ratio of filled to unfilled joints remains constant for continuous joints except between Station 8+00 and 12+00 which corresponds roughly to the unit of augen Toxaway Gneiss (Figure 11). The ratio increases for all joints (continuous and discontinuous) to the powerhouse area where it is approximately equal to the curve for prominent or continuous joints. These data confirm the field observation that the number of discontinuous joints decreases with progression into the tunnel. There is also a slight decrease in the number of continuous joints

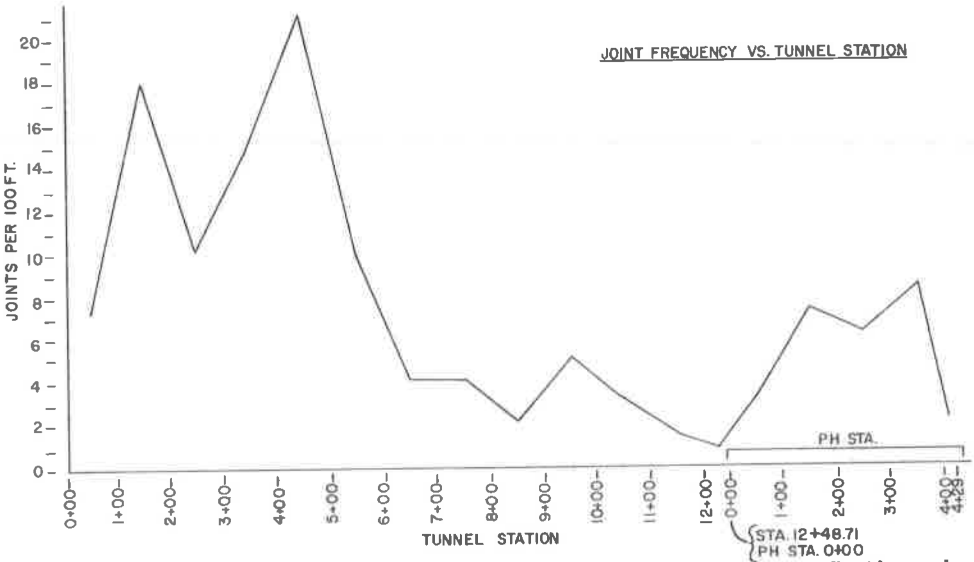


Figure 10. Plot of joint frequency versus tunnel station. Stations in feet.

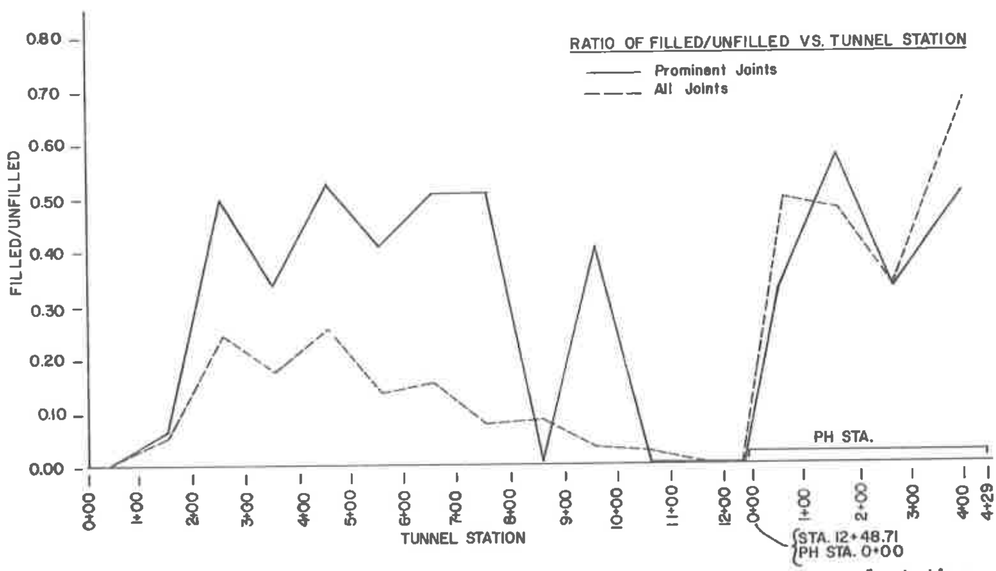


Figure 11. Plot of ratio of filled/unfilled joints verses tunnel station for prominent joints and all joints. Stations in feet.

towards the end of the tunnel, but the ratio of filled to unfilled joints remains almost constant.

A summary of our observations is that discontinuous joints have similar orientations to the continuous joint sets; that the northeast and northwest sets are more prominent near the tunnel portal and are generally discontinuous; and that some east-west joints are also discontinuous near the tunnel portal. From the above observations and data it is concluded that joints in the Toxaway Gneiss were formed at different times. The mineralized and possibly the continuous unfilled joints formed from tectonic stresses that were present during deformation. The minerals in the filled joints may have formed during the waning stages of metamorphism. The unfilled discontinuous joints and possibly the unfilled continuous joints formed later, either from residual tectonic stresses or from stresses unrelated to deformation. It is probable that all the unfilled joints did not develop at the same time.

INTERPRETATION

We interpret the large number of discontinuous joints near the surface to be younger than the mineralized joints and possibly younger than the unfilled continuous joints. The discontinuous joints are related to the large compressive in-situ stresses measured in the Toxaway Gneiss.

The formation of a joint plane requires the production of a discontinuity in what had been continuous rock. The process must involve separation of the rock with little of no offset. A tensile stress normal to the fracture plane would meet those conditions. According to Roberts (1961) the initial fracture may form parallel to the plane of maximum shear stress, but once the small incipient shear forms, the fracture may propogate by rapid separation of the rock due to release of stored elastic strain energy. This can explain the lack of movement on shear joints since the major direction of movement is normal, not parallel to the joint surface. Price (1959, 1966) has related the principal compressive stresses to the stored strain energy and has discussed certain conditions that must be met before the initial fractures can occur. These conditions involve Poisson's ratio, Young's modulus, compressive strength of the rock, orientation of principal stress directions, and most importantly the ratio of the greatest to least principal stress. In the Toxaway Gneiss under present stress conditions the ratio of greatest to least principal stress is apparently not great enough to cause fractures. Therefore, the discontinuous joints observed in this study may have formed or propogated from favorably oriented incipient fractures present before the development of the existing stress field.

There is more than one set of discontinuous joints and those joints have orientations similar to the mineralized and continuous joints. The discontinuous joints are old incipient or partly healed fractures that have propogated or reopened. The incipient fractures formed during the deformation that developed the mineralized and possibly the unfilled continuous joints, but did not propogate at that time because of insufficient elastic strain energy. Thus, the tectonic stresses must have been relieved before the incipient fractures could propogate. The present high stresses measured in the Toxaway Gneiss are not related to Paleozoic deformation but are possibly due to a mechanism related to crustal plate movements such as that described by Voight (1969) and summarized in the preceeding discussion of in-situ stress data. The propogation of the incipient fractures is related to the release of elastic strain energy, perhaps by differential expansion of the rock near the earth's surface. At depths greater than 100 m the static load is such that this energy cannot be released. Therefore, the number of discontinuous joints decreases with depth.

CONCLUSIONS

1. The high compressive in-situ stresses measured in the Toxaway Gneiss are not related to Paleozoic deformation, but have developed since the major deformation.

2. The joints in the Toxaway Gneiss formed at different times. The youngest joints formed in response to the release of stored elastic strain energy associated with the in-situ stresses. They propogate from pre-existing incipient fractures oriented subparallel to the earlier tectonically formed joints. At depths of 100 m the overburden pressure is great enough to prevent the release of the stored elastic strain energy.

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LATE CRETACEOUS PALYNOMORPHS FROM THE CAPE FEAR FORMATION OF NORTH CAROLINA

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ABSTRACT

The essentially unfossiliferous nature of the Cape Fear Formation of North Carolina has led to estimates of age based solely on physical criteria and stratigraphic position. Most workers regarded the formation as Early Cretaceous in age and considered it to be separated from the overlying Black Creek and Middendorf Formations (Upper Cretaceous) by a major unconformity.

Two samples of the Cape Fear Formation from the vicinity of its type locality near Fayetteville, North Carolina, contain four species of pollen that have previously been recorded only from the Eutaw Formation of Alabama and western Georgia and the upper part of the Raritan and all of the Magothy Formation of New Jersey. These species indicate that the Cape Fear Formation can be placed in pollen zone V of middle Turonian to late Santonian or early Campanian Age. Because all four species range throughout zone V, more refined age assignments of the Cape Fear Formation cannot be made at present. Both the Black Creek and Middendorf Formations can also be placed in zone V, which suggests that the hiatus separating the Cape Fear from the overlying units is smaller than had previously been thought.

By equating the Cape Fear Formation with pollen zone V, it is suggested that outcropping rocks of Early and early Late Cretaceous Age are absent over the axis of the Cape Fear arch. However, zone IV rocks of early Late Cretaceous Age have been observed both north and south of the arch, suggesting that the Cape Fear arch was a positive feature during early Late Cretaceous time and that it served to delineate the Albemarle embayment to the north from the Okefenokee embayment to the south.

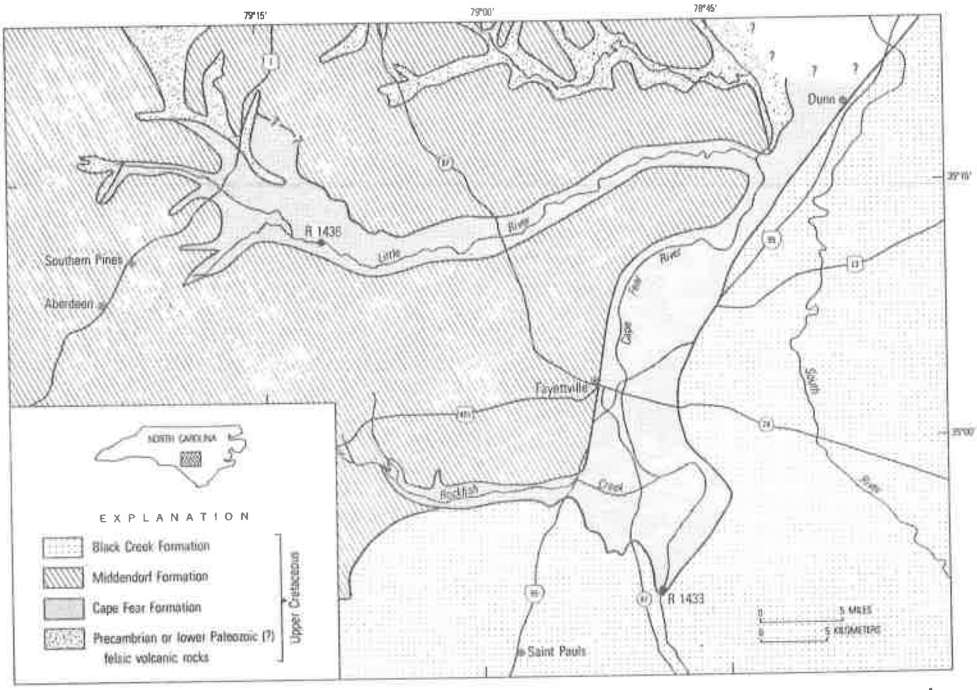


Figure 1. Generalized geologic map of the Cape Fear River region near Fayetteville, North Carolina, showing the two sample localities (modified from Heron and Wheeler, 1964).

INTRODUCTION

The Cape Fear Formation is the basal outcropping Cretaceous unit of the North Carolina Coastal Plain. It is well exposed in the Cape Fear River Valley, North Carolina (Figure 1), where it consists of a series of essentially unfossiliferous, poorly sorted sands and intercalated clays. The absence of calcareous mega- and microfossils has led to age assignments and correlations of the Cape Fear with other formations of the Atlantic and Gulf Coastal Plains based on physical criteria and/or stratigraphic position. As a result the unit has been variously referred to as the Cape Fear Formation (Stephenson, 1907; Heron, 1958; Heron and others, 1968; Swift and Heron, 1969), the Patuxent Formation (Stephenson, 1912, 1923), the Tuscaloosa Formation (Cooke, 1936; Spangler and Peterson, 1950; LeGrand and Brown, 1955), or simply as Lower Cretaceous? (Dorf, 1952). Ages proposed for the Cape Fear Formation range from Jurassic (Stephenson, 1907) to early Late Cretaceous (Cooke, 1936; Stephenson and others, 1942), although most workers regard the formation as Early Cretaceous in age (Stephenson, 1912, 1923; Cooke, 1926; Spangler and Peterson, 1950; Heron, 1958; Heron and others, 1968; Swift and Heron, 1969).

During a stratigraphic reconnaissance of the Cretaceous rocks of North Carolina, two samples of the Cape Fear Formation were examined that contain microfloral assemblages of low species diversity and generally poor preservation; however, they do contain several biostratigraphically important pollen types. These pollen types suggest a biostratigraphic correlation of the Cape Fear Formation with the Eutaw Formation of the Gulf Coast and with either the South Amboy Fire Clay Member of the Raritan Formation or with the Magothy Formation of New Jersey. The purpose of this paper is to present the palynological evidence that suggests these correlations and to suggest a possible age for the Cape Fear Formation.

Acknowledgments

The authors wish to express their appreciation to B. L. Blackwelder and L. W. Ward, U. S. Geological Survey, for their help in collecting the samples discussed in this report. Special thanks is expressed to Duncan Heron, Duke University, who accompanied the authors on a traverse along the Cape Fear and Tar Rivers and who provided insight into the Cretaceous stratigraphy of the Carolinas.

EVOLUTION OF THE STRATIGRAPHIC TERMINOLOGY APPLIED TO THE BASAL OUTCROPPING COASTAL PLAIN SEDIMENTS OF NORTH CAROLINA

The name "Cape Fear Formation" was first used by Stephenson (1907) for the "unfossiliferous" basal Coastal Plain sediments that crop out mainly along the Cape Fear, Neuse, and Tar Rivers, and Contentnea Creek in North Carolina. Stephenson first considered the unit to be Jurassic(?) in age, unconformably overlain by sediments that he assigned to the Cretaceous System.

Later, Stephenson (1912) grouped the Cape Fear Formation with some of the sediments of the Sandhills of the western North Carolina Coastal Plain, and, on the basis of their stratigraphic position and physical similarity, referred to these units as the southward extension of the Patuxent Formation (Lower Cretaceous) of Virginia. However, Dorf (1952, p. 2184) referred to the North Carolina Patuxent beds of Stephenson (1912) simply as "Lower Cretaceous?," because the absence of plant fossils from the Carolina sediments precludes a demonstrable biostratigraphic correlation between the two units.

Meanwhile, a second nomenclatural scheme was evolving based on supposed physical similarities between the North Carolina units and the Tuscaloosa Formation (Upper Cretaceous) of the eastern Gulf Coastal Plain. Cooke (1926) reintroduced the name "Cape Fear Formation" for the North Carolina sediments that Stephenson (1912) had previously referred to the Patuxent Formation, but Cooke considered the rocks to be of Late Cretaceous Age, correlating them with the Tuscaloosa Formation of Georgia and Alabama. Later, Cooke (1936) abandoned the name "Cape Fear", and placed the basal Cretaceous sediments of North Carolina, South Carolina, and part of Georgia in the Tuscaloosa Formation. Designation of these basal Coastal Plain sediments as "Tuscaloosa" was perpetuated by Stephenson and others (1942), Spangler and Peterson (1950), Spangler (1950), LeGrand and Brown (1955), Siple and others (1956), Stuckey and others (1958), and Conley (1962).

Heron (1958) resurrected the name "Cape Fear Formation" for a second time. He suggested that the so-called Tuscaloosa Formation in North Carolina consists of two genetically unrelated sedimentary units. Heron applied the name Cape Fear Formation to the lower unit of presumed Early Cretaceous Age, and the name Middendorf Formation of Late Cretaceous Age to the upper unit. This stratigraphic terminology has since been used by Heron (1960), Heron and Wheeler (1959, 1964), Heron and others (1968), and Swift and Heron (1969).

STRATIGRAPHIC AND LITHOLOGIC CHARACTERISTICS OF THE CAPE FEAR FORMATION

The Cape Fear Formation is underlain by crystalline rocks of the Piedmont physiographic province. At the lectotype locality (locality 5 of Heron and Wheeler, 1964), approximately 10 m of section is exposed, and Swift and Heron (1969) estimate the total thickness of the formation as 15 to 70 m in the vicinity of the Cape Fear River. The Cape Fear Formation is unconformably overlain by the Black Creek Formation in the Cape Fear River Valley, but farther west in Moore County, North Carolina, the Cape Fear is apparently unconformably overlain by the Middendorf Formation (Conley, 1962).

As discussed by Heron and others (1968) and Swift and Heron (1969), the Cape Fear Formation consists of a series of distinctively graded sand-to-mud cycles. Each cycle consists of a basal muddy sand overlain by a bed of sandy mud. In the ideal cycle, the basal sand bed is deposited on an erosional surface; the sand bed is generally 1 to 2 m thick, and the basal 3 to 30 cm tends to be gravelly. Cross-stratification is common toward the top of the sand bed. The contact between the basal sand and upper mud bed is gradational, but the top of the mud bed is an erosional surface, succeeded by the next sand bed. Interruptions in the form of conglomerates within the sand bed or incomplete mud beds are common in the ideal cycle. Minor amounts of lignite and amber have been found in the formation.

Primarily on the basis of clay mineralogy, the continuity of many of the beds over long distances, and the absence of mud cracks and root casts, Heron and others (1968) interpreted the Cape Fear Formation as having been deposited in an estuarine or lagoonal environment; periodic flushing of saline waters by river floods accounted for the cyclicity in sedimentation.

POLLEN ASSEMBLAGES FROM THE CAPE FEAR FORMATION

Sample Location and General Palynologic Characteristics

Sample locality R1433 (Figure 2) is exposed along the Cape Fear River, approximately 0. 8 km upstream from mile board 101 (locality 7 of Heron and Wheeler, 1964). Three to 5 m of the Cape Fear Formation is exposed at the base of the section. At this locality, Heron and Wheeler (1964) recognized three lithologic units within the Cape Fear: a basal, massive gray micaceous sand-silt-clay unit overlain by a massive, darker, sandy, clayey siltstone, which in turn is overlain by a medium-gray mudstone. Unconformably overlying the Cape Fear is the Black Creek Formation which consists of coarse clean sand and dark-gray clay. Two palynological samples were taken here; one from near the top of the Cape Fear, the second from the Black Creek Formation, 0. 7 m above its contact with the Cape Fear (Figure 2).

The Cape Fear sample (R1433A) contained a pollen assemblage having a very low yield, low diversity, and moderate preservation. Neither dinoflagellate cysts nor acritarchs were observed in this sample. No carbonized organic matter was observed, and much of the residue consisted of amorphous organic matter and some woody tissue and cuticle.

The Black Creek sample from this locality (R1433B) contained a rich, diverse, and well preserved pollen assemblage that can be correlated with the highest units of the Magothy Formation of the northern New Jersey Coastal Plain (i. e. , the "Morgan" and "Cliffwood beds"). Rare dinoflagellate cysts and acritarchs were also observed in the Black Creek sample, suggesting some marine influence during deposition of the unit.

A second Cape Fear sample was obtained from an exposure of the Cape Fear Formation along the south bank of Little River on the Fort Bragg Military Reservation, immediately beneath Morrison bridge. At this locality, as much as 1. 5 to 2 m of the formation is exposed, consisting of pale-greenish-gray (weathering to yellow), massive clayey and silty sand. Locally, irregular pods of black clay and carbonized wood can be found in the sand. Sample R1436 came from one of the clay pods approximately 0. 5 m above the base of the section (i. e. , above river level). The yield of palynomorphs in sample R1436 was also low; the assemblage was only slightly more diverse and better preserved than was that of sample R1433A. In addition to the biostratigraphically significant pollen types discussed below, this assemblage

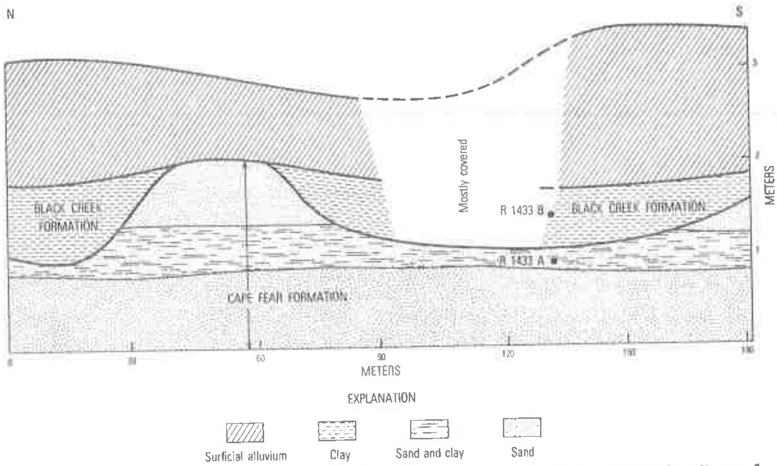


Figure 2. Cross-section of the Cape Fear and Black Creek Formations at locality R1433 (modified from Heron and Wheeler, 1964). Note the location of sample A (Cape Fear Formation) and sample B (Black Creek Formation).

contained several specimens of Araucariacites australis Cookson 1947; Aequitriradies spinulosus (Cookson & Dettman) Dettman 1963; Inaperturopollenites spp.; Cicatricosisporites sp. (with several thin rounded striae on each hemisphere that merge incompletely at the radial corners to produce a fenestrate or foveolate pad or shield covering the corner), several smooth unidentified trilete spores, and fungal spores. Carbonized organic matter was very abundant in this sample, and woody tissue accounted for most of the residue. Individual bordered pits were common throughout the sample.

Biostratigraphically Significant Pollen Types from the Cape Fear Formation

Four pollen species were recovered from either or both of the Cape Fear samples that suggest a biostratigraphic equivalence with the Eutaw Formation of Alabama and western Georgia and with either the South Amboy Fire Clay Member of the Raritan Formation and/or the Magothy Formation of New Jersey. The species are illustrated in Figures 4, 8, 12, and 16-17, where conspecific forms from the Eutaw, Black Creek, and Magothy Formations are also illustrated. Formal description of these species will be made when their geographic and geologic distribution is more completely known. Until that time, the species will be referred to by the designations applied to them in the U. S. Geological Survey (U. S. G. S.) Palynology laboratory, Reston, Virginia. A short diagnosis of each species is included in the figure explanations.

In addition to the observation of Complexiopollis sp. D in samples from the Eutaw, Black Creek, Middendorf, and Magothy Formations and from the South Amboy Fire Clay Member of the Raritan Formation reported on here (Figures 15-19) the species has also been reported by Groot and others (1961) (as Plicapollis serta) from units mapped as lower Tuscaloosa at Murdocksville, North Carolina (Middendorf of Heron, 1958), at Cheraw, South Carolina, and near Macon, Georgia; by Doyle (1969, figs. 4c, d) from the South Amboy Fire Clay (as Complexiopollis sp.); by Doyle and Robbins (1977, pl. 7, figs. 23-25) from the "post-Woodbridge" horizons of the New Jersey Raritan Formation (as Complexiopollis sp. C); and by Leopold and Pakiser (1964, pl. 8, figs. 40-46) from the McShan and Eutaw Formations of western Alabama (as Sporopollis pseudosporites Pflug 1953). Leopold and Pakiser (1964, pl. 5, figs. 34-38) also reported Sporopollis pseudosporites from the Coker Formation of western Alabama, which, together with the Gordo Formation, comprises the Tuscaloosa Group that underlies the McShan and Eutaw Formations. However, the same cores that were used by Leopold and Pakiser were examined by one of us (RAC), and no specimens referrable to this species were observed. Therefore, the occurrence of Complexiopollis sp. D in the Tuscaloosa Group of Alabama is not confirmed.

Complexiopollis sp. B (Figures 7-10) has been recorded from the South Amboy Fire Clay of New Jersey by Doyle (1969, fig. 4a, as Complexiopollis sp.), and from the "post-Woodbridge" horizons of the New Jersey Raritan Formation by Doyle and Robbins (1977, pl. 7, fig. 21, as Complexiopollis sp. B). The species has also been observed by the senior author in the outcropping Black Creek and Middendorf Formations of North Carolina (along the Cape Fear River and near Murdocksville, respectively), and the Eutaw Formation of western Alabama and eastern Georgia.

? Porocolpopollenites sp. A (Figures 3-6) has previously been reported from the South Amboy Fire Clay Member of the Raritan Formation (Doyle, 1969, fig. 4 g; Doyle and Robbins, 1977, pl. 7, figs. 19, 20), and from the McShan and Eutaw Formations of western Alabama (Leopold and Pakiser, 1964, pl. 8, figs. 53-58). Christopher (unpub. data) has demonstrated that ? Porocolpopollenites sp. A occurs in the outcropping Middendorf Formation of North Carolina, and in the subsurface Black Creek Formation of South Carolina. In addition, he has observe this species in all but the uppermost part of the "Cliffwood" beds of the Magothy Formation of New Jersey.

Leopold and Pakiser (1964, pl. 8, figs. 18-20) illustrated a species from the Eutaw Formation of western Alabama, which is similar to New Genus D sp.D illustrated here (Figures 11-14), and Groot and others (1961) recorded its presence (as Vacuopollis sp. 2) from units mapped as Tuscaloosa near Macon, Georgia. New Genus D sp. D

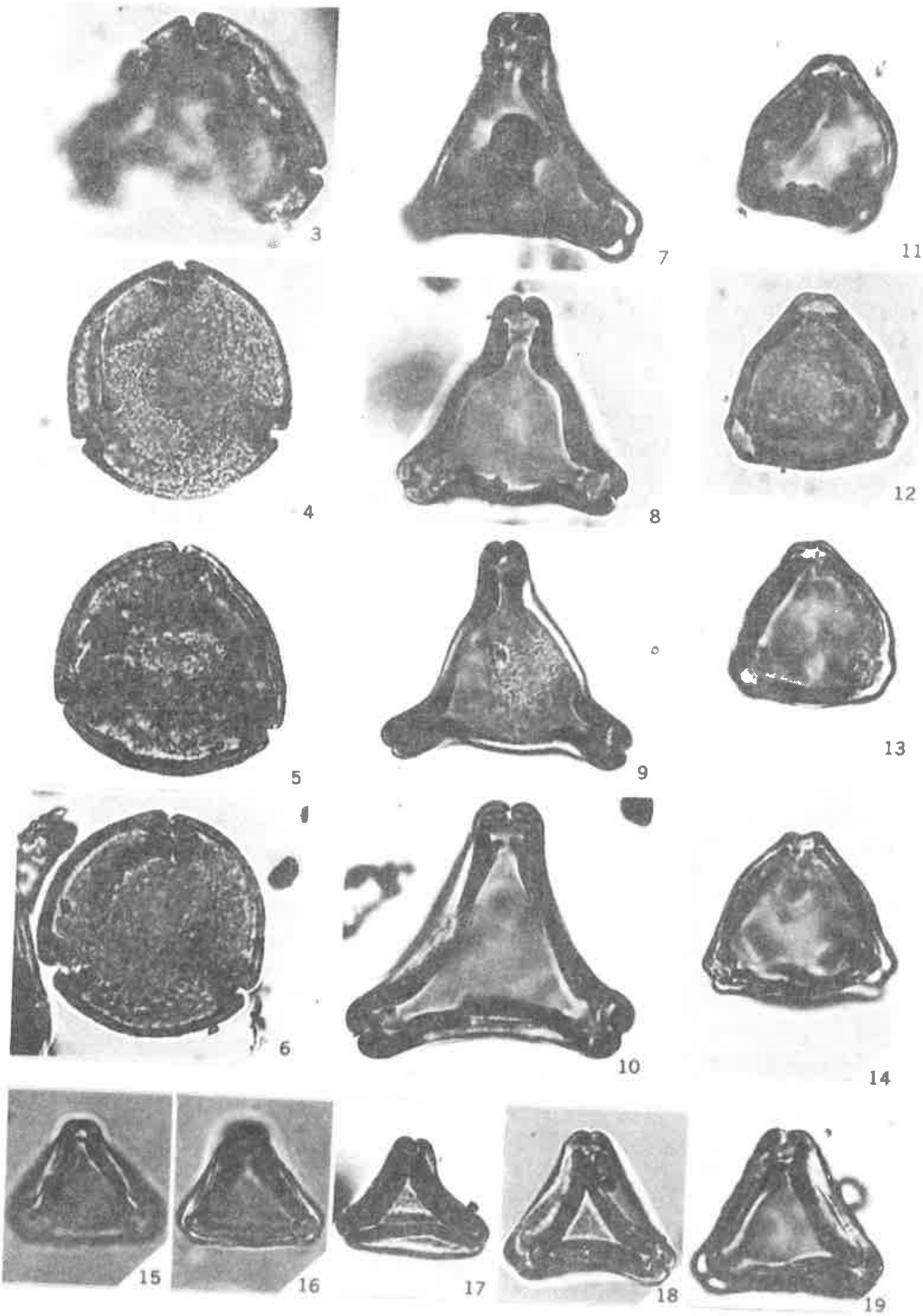
Figures 3-6. ? Porocolpopollenites sp. A from the Cape Fear Formation (Figure 3), Magothy Formation (Figure 4), Eutaw Formation (Figure 5 and Black Creek Formation (Figure 6). The species can be recognized by its oblate shape, circular to subcircular amb, scabrate surface, and brevicolporate apertures.

Figures 7-10. Complexiopollis sp. B from the Cape Fear Formation (Figure 7), Magothy Formation (Figure 8), Eutaw Formation (Figure 9), and Black Creek Formation (Figure 10). This species is characterized by its thick robust annuli and prominent arci that are either restricted to the germinal region or that extend from germinal to germinal.

Figures 11-14. New Genus D sp. D from the Cape Fear Formation (Figure 11), Magothy Formation (Figure 12), Eutaw Formation (Figure 13), and Black Creek Formation (Figure 14). Genus D (working code, U. S. G. S. Palynology laboratory, Reston, Va.) has a convexly triangular amb, small, poorly developed annuli, and a narrow interloculum or "Schictfuge. " In addition, New Genus D sp. D possesses arci that connect the germinals and parallel the sides of the grain.

Figures 15-19. Complexiopollis sp. D from the Cape Fear (Figures 15, 16), Magothy (Figure 17), Eutaw (Figure 18), and Black Creek Formations (Figure 19). The species is characterized by folding of the exine in the interapertual areas that tends to parallel the sides of the grains. On most specimens, the folded edge of the exine produces a straight, wavy, or irregular structure that superficially resembles arci. Some specimens appear to possess thickened equatorial margins. In this respect, and in respect to its small size compared with other species of the genus, Complexiopollis sp. D. is similar to Complexiopollis abditus Tschudy 1973, but the latter species possesses true arci that are set in from the margins of the grain.

is a common element throughout the South Amboy Fire Clay Member of the Raritan and the entire Magothy Formation of New Jersey, the Middendorf and lower part of the Black Creek of North Carolina, lower part of the Black Creek of South Carolina, and the Eutaw of the Chattahoochee River.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STRATIGRAPHIC IMPLICATIONS OF THE CORRELATIONS PROPOSED HEREIN FOR THE CAPE FEAR FORMATION

Stratigraphic palynological studies by Doyle (1969), Sirkin (1974), Doyle and Robbins (1977), and Christopher (1977a, b) have resulted in the recognition of two major biostratigraphic zones within the Raritan and Magothy Formations of New Jersey that can be extended to other parts of the Coastal Plain.

The basal zone, referred to as zone IV, is characterized by an assemblage in which Complexiopollis and Atlantopollis are the only Normapolles and/or triporate genera that occur (see Doyle and Robbins, 1977; Christopher, 1977b for additional characteristics of zone IV). The age of zone IV is considered to range from middle to late Cenomanian, and possibly to early Turonian. The zone occurs in the Woodbridge Clay and Syreville Sand Members of the Raritan Formation of New Jersey (Doyle and Robbins, 1977; Christopher, 1977b), in the supposed Cape Fear Formation from the subsurface of South Carolina (Hazel and others, 1978), and in the Coker Formation (basal formation of the Tuscaloosa Group) of western Georgia and Alabama (Christopher, in press). Scant and poorly preserved marine invertebrates at the stratotype of the zone in New Jersey suggest a biostratigraphic equivalence with the Woodbine Formation of Texas (Richards, 1943; Stephenson, 1954).

The biostratigraphic unit overlying zone IV is informally referred to as zone V and contains a considerably more diverse and abundant Normapolles and triporate element, which Christopher (1977a) suggested points to an hiatus between zone IV below and zone V above. Christopher (1977b) recognized three subzones within zone V based on changes in species composition and changes in the relative abundance of selected pollen genera. These subzones, arranged in ascending stratigraphic order, are referred to as subzones V-A, V-B, and V-C. Subzone V-A occurs in the South Amboy Fire Clay Member of the Raritan Formation and the lower part of the Old Bridge Sand Member of the Magothy Formation; subzone V-B occurs in the upper part of the Old Bridge and the Amboy Stoneware Clay Members of the Magothy; subzone V-C occurs in the "Morgan" and "Cliffwood" beds of the Magothy. Christopher (1977a, b) considered zone V to range from middle Turonian to late Santonian in age. Any time gaps within this zone are of considerably less duration than the hiatus between zones IV and V (which represents the Lower? Turonian).

Of the biostratigraphically significant pollen recovered from the Cape Fear Formation, Complexiopollis sp. B, C. sp. D, and New Genus D sp. D occur throughout zone V but are absent in zone IV. ?Porocolpopollenites is also absent from zone IV and is present throughout all but the uppermost part of zone V. At the stratotype, all four species are more common in subzone V-A than in subzone V-B or V-C, but biostratigraphic placement of the outcropping Cape Fear Formation of North Carolina cannot be made more precisely than to assign the unit to zone V. However, the presence of Aequitriradites spinulosus in sample R1436 suggests placement low in zone V, as the species has not been recorded from higher horizons.

The fact that the lower part of the Black Creek and the Middendorf can also be placed in zone V suggests that perhaps the Cape Fear Formation is not so far removed in time from either of these two units and that the depositional history of the Cape Fear might be related to that of the overlying units.

On the basis of the evidence presented here, it appears that the Cape Fear Formation of North Carolina has a greater paleontological affinity with both the Eutaw and Magothy Formations than it does with the Tuscaloosa Group (which underlies the Eutaw Formation in eastern Alabama and western Georgia) or the lower part of the Raritan Formation (which underlies the Magothy Formation in New Jersey). These findings support those of Groot and others (1961), who considered the units mapped as Tuscaloosa in the Carolinas and eastern Georgia to be closer in age to the Magothy Formation of New Jersey than to either the Tuscaloosa Group of Alabama or to the lower part of the Raritan Formation of New Jersey. These results suggest that biostratigraphic equivalents of the Tuscaloosa Group of the eastern Gulf Coastal Plain are missing in the outcrop section along the axis of the Cape Fear arch, and as Heron (1958), Sohl (1976), and others have suggested, the term "Tuscaloosa" should be abandoned in this area, and the names "Cape Fear" and "Middendorf" be applied to the basal outcropping units. This conclusion is supported by the preliminary results of a palynologic study of so-called Tuscaloosa sediments along the Tar River, North Carolina. Pollen assemblages from the Tar River contain a variety of species belonging to the genera Plicapollis, Pseudoplicapollis, Trudopollis, Complexiopollis, and Minorpollis. These forms are typically associated with subzones V-B and V-C, the stratotypes for which are in the Magothy Formation of New Jersey. However, palynological examination of core material from 1) Halifax County, North Carolina (supplied by P. M. Brown, U. S. G. S., Raleigh, North Carolina), north of the arch in the Albemarle embayment (Figure 20); and 2) Dorchester County, South Carolina, south of the arch in the Okefenokee embayment (see Hazel and others, 1978), indicates that biostratigraphic equivalents of the Tuscaloosa Group (i. e., pollen zone IV) do occur in the embayments that flank the arch. If additional studies substantiate our claim that Tuscaloosa-lower part of the Raritan equivalents are absent over the axis of the arch but are present in the embayments on its flanks, then these units were either removed by an episode of post-Tuscaloosa, pre-Black Creek erosion, or the Cape Fear arch was a positive feature during this time interval (middle to late Cenomanian). As such, it would have served to separate the Albemarle embayment to the north from the Okefenokee embayment to the south.

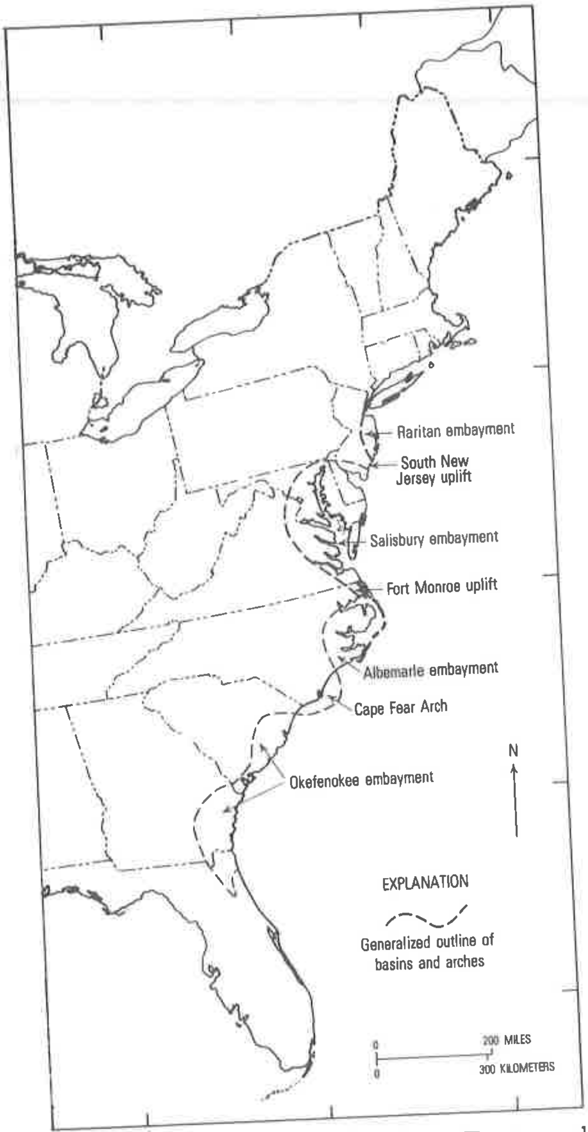

Figure 20. Location of the Cape Fear arch, the Albemarle embayment, the Okefenokee embayment, and other major tectonic features of the basement beneath the Atlantic Coastal Plain (from Owens and Sohl, 1969).

If the hypothesis is accepted that the Cape Fear arch was a positive feature during zone IV time, the term "Tuscaloosa" should not be used for the units assigned to pollen zone IV that occur on the northern flank of the arch, as their depositional history would have been different from that of the Tuscaloosa Group of the eastern Gulf Coast. Whether or not the term "Tuscaloosa" can be applied to the zone IV sediments on the southern flank of the arch must await more detailed lithologic and stratigraphic investigations presently being conducted (G. Gohn, written comm., 1977).

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CHROMITE PARAGENESIS IN ALPINE-TYPE ULTRAMAFIC ROCKS OF THE SOUTHERN APPALACHIANS

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ABSTRACT

In the ultramafic rocks of the Southern Appalachians, chromite occurs: (1) as isolated grains disseminated throughout the olivine-rich rock ("disseminated" chromite); and (2) in massive chromitite pods or layers ("massive" chromite). Distribution coefficients based on the partitioning of Fe and Mg between polygonal constant-composition olivine grains and variable composition chromite grains suggest that both forms of chromite at least partially equilibrated with olivine. Massive chromite apparently partially equilibrated with olivine at higher temperatures than did disseminated chromite, either in the mantle or lower crust; the disseminated chromite apparently equilibrated with olivine in the crust, during the Appalachian tectonic event.

INTRODUCTION

Previous petrographic and petrofabric studies on the ultramafic rocks of the Southern Appalachians (Astwood and others, 1972) strongly suggest that extensive recrystallization subsequent to emplacement has occurred. To test this hypothesis, we have attempted to establish if subsolidus chemical equilibration occurred between coexisting chromite and olivine. This information is critical to establish boundary conditions for interpreting the origin, emplacement and tectonic history of the ultramafic rocks.

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We acknowledge financial support for this project through National Science Foundation grant GA-4518. Several people, including our colleagues Arthur Snoke and Edwin Sharp at the University of South Carolina and T. N. Irvine at the Geophysical Laboratory of Washington, read various drafts of this manuscript and offered very helpful suggestions. All conclusions, however, are ours alone and do not necessarily carry the endorsement of any of the above.

GEOLOGIC SETTING

The ultramafic rocks of western North Carolina are typically lenticular masses of dunite or serpentinized dunite elongated parallel to the northeast regional structural trend. The ultramafic rocks are most commonly found in the Ashe Formation in the Blue Ridge anticlinorium (Rankin and others, 1973). R. H. Carpenter (1970) has determined that the rocks of this area belong to the kyanite-staurolite or sillimanite metamorphic zone.

The country rocks enclosing the ultramafic bodies have been subjected to a complex thermal and tectonic history (Kulp and Poldervaart, 1956), including at least two episodes of folding (Neuhauser and Carpenter, 1971; Hatcher, 1972; Butler, 1973) and syntectonic intrusion of granitic rocks (Brobst, 1962). The second folding episode, which generated the predominant meso- and macro-structural elements, resulted in the formation of open to tight northeast trending synforms and antiforms with axial planes dipping steeply to the southeast (Neuhauser and Carpenter, 1971; Hatcher, 1972; Butler, 1973).

PETROGRAPHY

The ultramafic rocks are mainly dunites and partially serpentinized dunites. The least altered rocks contain about 90% forsteritic olivine, 5-8% talc, 1-2% chromite, with variable amounts of serpentine minerals, and minor chlorite, orthopyroxene, clinopyroxene and magnesite. The olivine is equant and polygonal, generally 0.1 to 1.0 mm in diameter with a thin serpentine rind. Polygonal olivine meeting at 120° triple points is generally conceded to have been generated by metamorphic recrystallization. Talc occurs as separate grains replacing earlier orthopyroxene and as stringers along fractures in the ultramafic rocks.

Chromite occurs in basically two different modes: as isolated disseminated grains throughout the rock; and as massive chromite in small, deformed layers and lenses and in large pods. The disseminated grains are euhedral to subhedral and are often surrounded by small amounts of kammererite (chromian chlorite) and exhibit small opaque secondary chromite borders. The "massive" chromite grains are large (up to 2 cm diameter), anhedral, and highly fractured, and are embayed with interstitial olivine and kammererite.

Two samples were taken from the Webster-Addie ultramafic complex. In this rock body layers of chromite, olivine, clinopyroxene and orthopyroxene alternate in a typical magmatic layered texture, as would happen by gravitational settling of the early crystal fraction of a crystallizing basaltic magma. The Webster-Addie body is texturally unique in the Southern Appalachians and samples were analyzed in order to ascertain if element fractionation in this type body is similar to the other, less-differentiated bodies characteristic of the Southern Appalachians.

Based on thin-section and polished section examination it was concluded that the chromite grains exhibited no significant secondary alteration other than the chloritization mentioned above. These observations insured that accurate analyses of chromite could be obtained by analyzing the central parts of grains.

PREVIOUS WORK ON NORTH CAROLINA CHROMITE

There are few major-element analyses of the chromites of western North Carolina. Hunter and others (1942) reported a few analyses of placer and massive chromite from the region, but all showed lower chromium contents than the samples analyzed for this study, probably because of alteration or gangue material. Thayer (1946) reported an analysis of high-aluminum chromian spinel found in troctolite from the Buck Creek area in Clay County. He stated that the high aluminum content was probably genetically related to the high aluminum content of the containing rock. Thayer (1956) also reported an analysis of placer chromite from Democrat, N. C., but stated that the sample may have been contaminated.

Miller (1953) analyzed disseminated chromite from the Webster area but was troubled by secondary chemical alteration of the primary chromite. A short paper by Bentzen (1970) gave chromite analyses from most of the areas sampled in this study, but they were all relatively low in chrome content and high in silica content, implying that gangue material probably was present to some extent. The mode of occurrence of the samples was not identified.

SAMPLING

The sample localities in western North Carolina were extensively described by Hunter (1941) and Hunter and others (1942). From northeast to southwest the areas sampled are as follows: Newdale (ND)

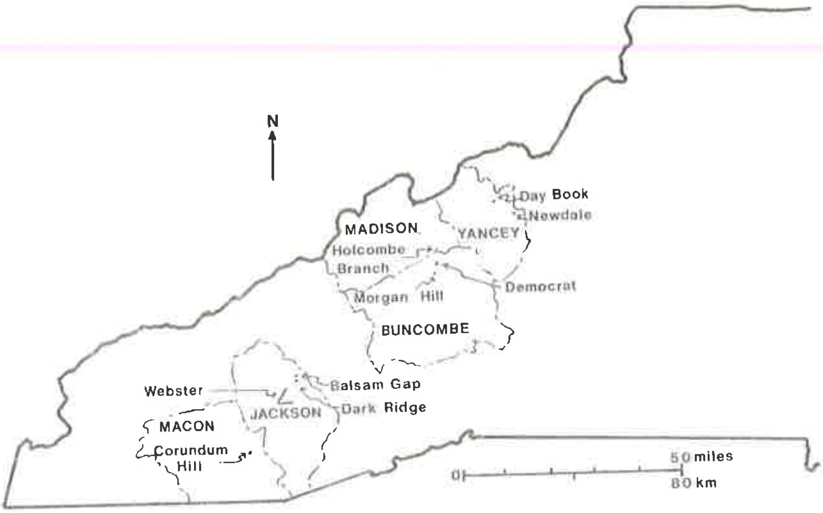

Figure 1. Index map of Western North Carolina, showing location of ultramafic bodies studied.

and Day Book (DB) in Yancey County; Holcombe Branch (HB) in southern Madison County; Democrat (D) and Morgan Hill (MH) in northern Buncombe County; Balsam Gap (BG), Dark Ridge (DR) and Webster (W) in Jackson County; and Corundum Hill (CH) in Macon County (Figure 1). These locations were chosen primarily for the amount and variety of chromite exposed and secondarily for the existence of previously determined compositions of associated olivine.

CHROMITE ANALYSES

The chromite samples were examined in hand specimen, in thin section and in polished section for secondary alteration. Individual grains were hand picked from the crushed rock or stream samples and were imbedded with epoxy cement in small holes drilled in one-inch diameter phenolic resin discs. These disc were then ground, polished and coated with carbon for electron microprobe analysis. Chemical analyses were done on an automated and computer-controlled Materials Analysis Company Model 400 electron microprobe at the Geophysical Laboratory of the Carnegie Institution of Washington. The details of the system are given by Boyd and others (1969) and Finger and Hadidiacos (1972). Three crystal detectors were employed to analyze for four major elements, Fe, Mg, Cr, and Al, and two minor elements, Mn and Ti. The silicon content was determined as a check for silicate contamination and consequently is not reported. Each detector counted for thirty seconds or thirty thousand counts, whichever came first for each element. Threé separate spots were analyzed on each grain and averaged. Homogeneity factors and standard deviations were calculated but are not reported because they were all within acceptable limits. Typical standard deviations for the oxide weight percentage are as follows: FeO (total iron), 0.2; MgO, 0.1; MnO, 0.03; TiO_2, 0.0004; Cr_2O_3, 0.3; Al_2O_3, 0.04 (Dickey, personal communication). Details of the microprobe analysis plan are given in Appendix A.

Oxide weight percentage data were recalculated to moles per four moles of oxygen. In so doing the sum of the cation molar concentrations were normalized to equal three. From charge considerations Fe^{+3} was calculated as the amount necessary to make the total number of moles of trivalent cations equal to two. One mole of Fe^{+2} was combined with one mole of Ti^{+4} and counted as two moles of trivalent cations before Fe^{+3} was calculated. The cation mole fractions were then converted to oxide mole fractions appropriate for graphical representation and thermodynamic treatment.

CHROMITE COMPOSITIONS

The chromite compositions obtained in this study, given in oxide weight percents, cation moles relative to four moles of oxygen, and molar ratios of divalent and trivalent oxides to the theoretical RO and R_2O_3 components respectively, are reported in Appendix B. Where two or more analyses from the same sample of chromite are available, averages are taken to obtain a single graphical point. Again it should be pointed out that all chromite grains are compositionally homogeneous, thus facilitating the plotting of composition. A summary of the data is presented in Figure 2, a triangular compositional diagram in which the apices represent the dominant trivalent constituents, Al, Cr and Fe, and in Figure 3, which is a plot of the mole fraction of Mg (X_{Mg}) against the parameter 100Cr/(Cr+Al). From Figure 3 it can be seen that most of the chromite is quite rich in chromium (>65%). In this diagram, no compositional patterns for disseminated, massive, or Webster chromite emerge.

In Figure 3 it can be seen that compositional patterns emerge for Webster, disseminated, and massive chromite and polygons are drawn around the various chromite types. While some overlap is evident, it is minor and generalizations about compositional variation can be made. Webster chromite, for example, is generally lower in magnesium than either the disseminated or massive chromite, the massive chromite is generally highest in magnesium, and disseminated chromite, while occupying a middle position with respect to magnesium, generally has higher Cr/Al than the other samples.

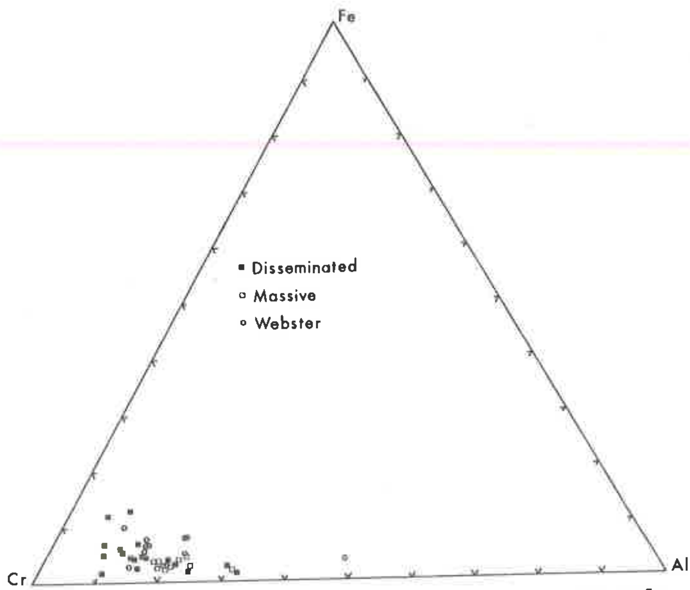

Figure 2. Chromite trivalent constituent triangular compositional diagram.

EQUILIBRATION TEMPERATURES

One method for estimating the last equilibration temperature for olivine-chromite pairs has been worked out theoretically by Irvine (1965) and calibrated by Jackson (1969). In this model the partitioning of Mg and Fe^{+2} is a function of R_2O_3 and is based on the equation:

$$\ln K_D = \ln\ [(X^{OL}_{Mg}/X^{OL}_{Fe})\ (X^{SP}_{Fe}/X^{SP}_{Mg})] = \ln K + Y^{SP}_{Cr} \ln K_2 + Y^{SP}_{Fe^{+3}} \ln K_3 \quad (1)$$

Where X^{B}_{α} is the mole fraction of element α in phase β, Y^{SP}_{i} is the mole fraction of trivalent constituent $\underline{i}$ in spinel, and K is an equilibrium constant for the given exchange reaction (Irvine, 1965):

K_1: $1/2\ Fe_2SiO_4 + Mg\ Al_2O_4 = 1/2\ Mg_2SiO_4 + FeAl_2O_4$; (2)

K_2: $MgCr_2O_4 + FeAl_2O_4 = MgAl_2O_4 + FeCr_2O_4$ (3)

K_3: $FeAl_2O_4 + MgFe_2O_4 = FeFe_2O_4 + MgAl_2O_4$; (4)

Considerable errors in equilibration temperatures deriving from the Jackson (1969) calibration might be anticipated due to uncertainties in the thermochemical data. In fact Evans and Wright (1972) demonstrated that calculated equilibration temperatures were several hundred degrees too high for some chromite-olivine pairs. In spite of the expected errors in absolute equilibration temperatures, however, the method of Irvine (1965) and Jackson (1969) should be useful for comparing olivine-chromite pairs from different P-T (especially T) environments. This approach has been used by both Medaris (1975)

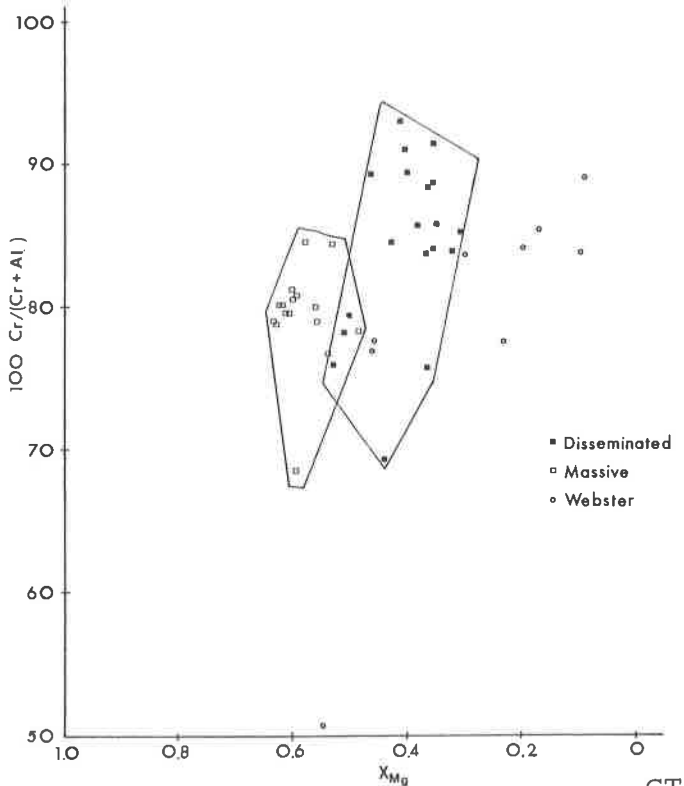

Figure 3. Chromite compositions plotted on X_{Mg}^{CT} vs 100 Cr/(Cr + Al) diagram.

and Evans and Frost (1975) in their analyses of coexisting olivine-chromite pairs and the reader is referred to their papers (as well as those of Irvine and Jackson) for details of the procedure.

According to Medaris (1975, p. 955; see also Irvine, 1955, p. 660)

> "Coexisting pairs of olivine and spinel in equilibrium at a given temperature should define a planar surface, assuming ideal solutions, in a three-dimensional figure with axes corresponding to (ln K_D), Y_{CR}^{SP} and Y_{Fe}^{SP}. An isothermal plane would have an intercept of (ln K_1), and slopes of (ln K_2) and (ln K_3) on the appropriate faces of the figure... "

(see Figure 4 of this diagram).

Because $Y_{Fe}^{SP}+3$ is uniformly low in the chromite grains of this study it is possible to facilitate comparison of olivine-chromite data by plotting data points on a ln K_D - Y_{Cr}^{SP} diagram by projecting each point up to the front face of Figure 4 where $Y_{Fe}^{SP}+3$ = O. Olivine compositions from Carpenter and Phyfer (1975) are used. Following Medaris (1975) and Evans and Frost (1975), we will estimate ln K_3 = 4.

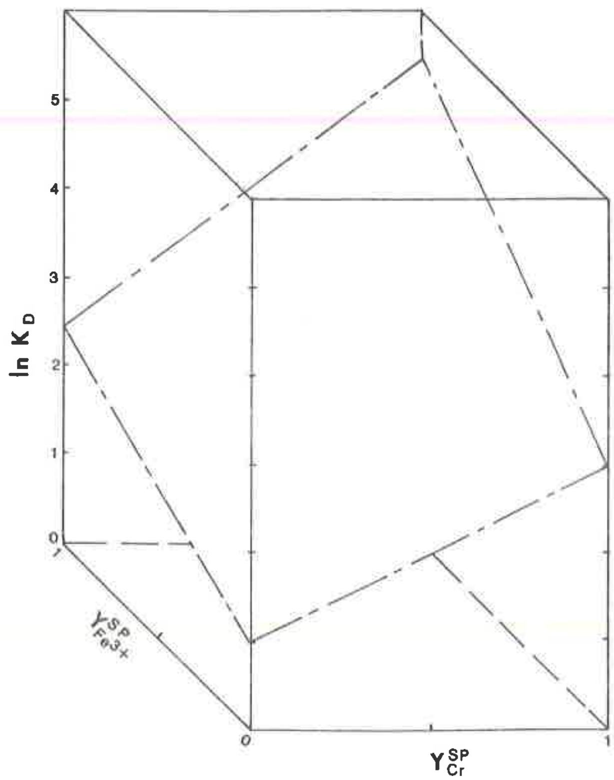

Figure 4. General model for olivine-spinel partitioning showing approximate location of 800° C isothermal plane (after Medaris, 1975).

Medaris (1975) plotted isotherms for 800°C, 1000°C and 1200°C based on the model of Irvine (1965) and Jackson (1969), but he noted that these temperatures are possibly several hundred degrees too high. Evans and Frost (1975), on the other hand, empirically plotted a 1200°C isotherm based on olivine-chromite pairs from Kilauea Iki pumice (Evans and Wright, 1972) and pairs from mid-Atlantic pillow basalt (Thompson, 1973). They also plotted an empirically derived 700°C isotherm based on pairs from enstatite-olivine-chlorite-spinel rocks.

Carpenter and Phyfer (1975) have shown that the X^{OL}_{Mg} for polygonal olivine is quite uniform in the ultramafic rocks of the Southern Appalachians, showing no significant variation from a mean value of $Fo_{92.3}$ either within a given ultramafic rock body or between ultramafic rock bodies. We have chosen to use this mean value of $Fo_{92.3}$ and our variable chromite compositions to obtain ln K_D values.

Analyses of olivine intimately associated with massive chromite were not performed. Later, when it became apparent that these analyses would be important in developing the story contained herewith, an unsuccessful attempt was made to re-collect. Due to very extensive quarrying at several of the massive-chromite bearing ultramafic bodies no samples were found. In an attempt to see the extent to which the olivine compositions effected the ln K_D values, we calculated the olivine composition, which, with average massive chromite MgO/RO and FeO/RO values, would yield ln K_D values similar to those for disseminated chromite. We used the mean value for X_{Fe}^{SP}/X_{Mg}^{SP} of 0.77 for all massive chromite samples. It was found that the olivine composition of Fo_{96}, considerably higher than any olivine reported in this or even the paper of Loney and others (1971) for the Burro Mountain, California ultramafic body, would be necessary to generate ln K_D = 2.88, the mean value for olivine-disseminated chromite analyses.

Therefore we have calculated ln K_D values for disseminated and massive chromite using the mean polygonal olivine composition of $Fo_{92.3}$.

We have plotted our data in Figure 5 on a diagram similar to that used by both Medaris (1975) and Evans and Frost (1975). We have superimposed both Medaris' (1975) 800°C and 1200°C isotherms and Evans' and Frost's (1975) 700°C and 1200°C isotherms. Our data for massive and disseminated chromite generally fall between Medaris' 700°C and 1200°C isotherms, whereas data from Webster are scattered across the diagram and show no particular pattern. Although some overlap exists, massive and disseminated chromite seem to occupy different areas in Figure 5. In general disseminated chromite grains have a higher ln K_D than massive chromite grains (mean K_D for disseminated chromite = 2.88; mean K_D for massive chromite = 2.18). If we use Evans' and Frost's isotherms, the disseminated chromite equilibrated with polygonal olivine at temperatures of approximately 700°C to 900°C. Massive chromite grains, when compositionally compared to polygonal olivine, yield higher equilibration temperatures, near 1000°C, but still well below melting temperatures of these rocks. Again, it must be stressed that the temperature values given above are only approximate and are based on empirical isotherms. More significant is the observation that the ln K_D values for olivine-disseminated chromite are higher than ln K_D values for massive chromite, and therefore, relatively, the disseminated chromite equilibrated with olivine at lower temperatures than did the massive chromite.

CONCLUSIONS AND SPECULATIONS

We conclude that disseminated chromite in the ultramafic rocks of the Southern Appalachians equilibrated with polygonal olivine at temperatures well below the melting temperature of the rocks. Because disseminated chromite and massive chromite are frequently found in very close proximity in the same ultramafic rock, we must explain

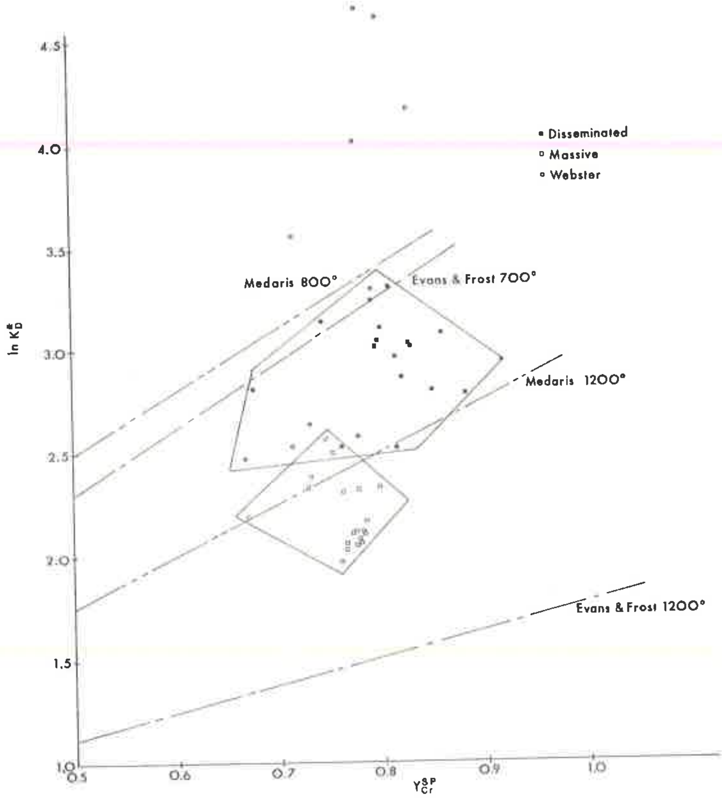

Figure 5. Logarithm of partition coefficient K_D for exchange equilibrium equation - as a function of Y^{SP}_{Fe}. ln K_D has been normalized for $Y^{SP}_{Fe} = 0.05$. Tentative 700°C and 1200°C isotherms after Evans and Frost (1975); 800°C and 1200°C isotherms after Medaris (1975).

the apparent differences in equilibrium temperatures. We hypothesize that all chromite in these rocks was in equilibrium with olivine in the upper mantle, when these rocks were formed at temperatures considerably higher than the 700°-1000°C temperatures estimated in this paper for the most recent equilibration.

We further speculate that the ultramafic rocks were emplaced into the crust such that chromite occurred both as disseminated grains surrounded by olivine and as massive layers in which a single chromite grain was surrounded by other chromite. During the Appalachian tectonic event these rocks were subjected to amphibolite facies temperatures (probably greater than 500°C), during which time olivine recrystallized (texturally and probably compositionally), to the polygonal form now seen. Those chromite grains surrounded by polygonal olivine are thought to have re-equilibrated at these metamorphic temperatures. That chromite which existed as massive layers was not in contact with polygonal olivine and thus did not re-equilibrate. Therefore, the compositional variation we now see in chromite is thought to be a reflection of only a partial re-equilibration under metamorphic conditions.

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RECENT ACCUMULATION OF SEDIMENT IN LAKES IN THE BEAR CREEK WATERSHED IN THE MISSISSIPPI DELTA

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ABSTRACT

Sediment samples were collected from six lakes along Bear Creek in Leflore, Sunflower, and Humphrey Counties in Mississippi to determine rates of sediment accumulation from 1954 to 1977 and 1962 to 1977 by measuring the fallout radionuclides (^{137}Cs) content in the sediment profile. Sediment accumulation rates since 1964 ranged from less than 1 cm/yr to more than 7 cm/yr averaging 3.3 cm/yr. These sediment accumulation rates indicate that surface erosion is occurring within the Bear Creek watershed. These oxbow lakes along Bear Creek are probably remnants of Ohio River and may have been in existance for 3000 years. The present rate of sediment accumulation would fill them in another 100 to 200 years.

INTRODUCTION

Sediment accumulation in lakes and reservoirs is a major problem around the world (Robinson, 1971). Sediment fills lakes and reservoirs, reducing their water storing capacity, and adversely affecting aquatic ecosystem. Little is known of the rates of sediment accumulation in the lakes in the Mississippi Delta. Most of these lakes are abandoned channel scars and oxbow remnants of the earlier Ohio, Mississippi, and Yazoo Rivers. As such, they are geologically recent but in terms of modern man's settlement of the Mississippi Delta they are old. The purpose of this study was to estimate the recent rates of sediment accumulation in six lakes in the Bear Creek watershed.

Acknowledgments

This paper is a contribution of the USDA Sedimentation Laboratory, Delta States Area, Agricultural Research, U. S. Department of

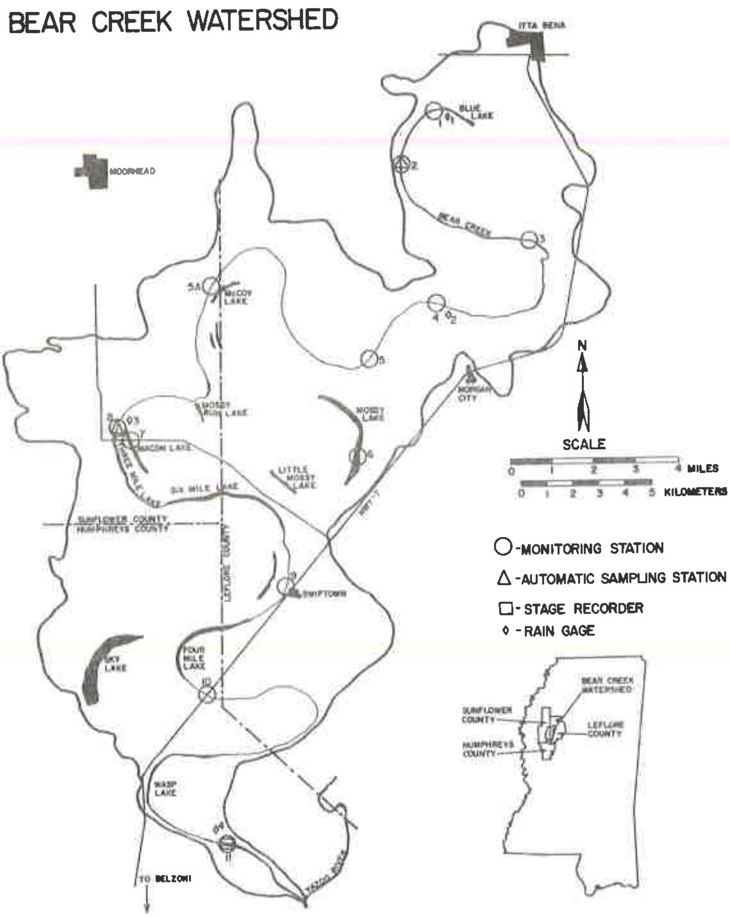

Figure 1. Sketch of Bear Creek Watershed, Leflore, Sunflower, and Humphrey Counties, Mississippi.

Agriculture in cooperation with the Vicksburg District of the U. S. Corps of Engineers, Department of the Army. Steve Scott assisted in field collection and laboratory preparation of the samples. Bob Wilson assisted with the computer reduction of the data and Valerie Herberger, Peggy Hamilton, and Winfred Cook helped prepare the manuscript.

STUDY SITE

Bear Creek is located in Leflore, Sunflower, and Humphrey Counties in Mississippi (Figure 1). The Bear Creek segment of the Mississippi Delta was probably cut by the youngest of the Ohio River scars. The meander size indicates that it was formed by the Ohio River. The diversion of the Ohio River from this course probably occurred prior the diversion of the Mississippi River to the Morehouse lowland route which probably occurred approximately 3000 years ago (Fisk, 1944). Bear Creek watershed is located entirely within the Mississippi Delta lowland. The stream meanders 80 km (50 miles) from its headwater in Blue Lake to the Yazoo River. There is 7.9 meters (26 ft) difference in elevation between the headwaters and the mouth of Bear Creek. Bear Creek lies outside the Mississippi River levee and is affected by Mississippi River flood only as they affect backwater on the Yazoo River.

METHOD AND MATERIALS

Sediment samples were collected from six lakes in the Bear Creek watershed. Two sample sites were selected in Blue, Macon, and Six-mile lakes and three sample sites were selected in Mossy, Three-mile, and Wasp lakes (Figure 1). For the two sample lakes, samples were collected near the upstream and downstream ends of each lake and where three sites were selected, a sample site was also collected in the middle of the lake. At each sampling site eight sediment cores were taken and divided into 10-cm vertical increments and each increment was composited for radiological analysis. The samples were dried at 105°C for 48 hours and ground to pass a 12-mm screen. Radiological analyses were made using a 1024 channel-pulse height analyzer and a NaI (Tl) crystal (Ritchie and McHenry, 1973). The complex gamma-ray spectra were reduced by a least squares technique (Schonfeld, 1966) to give the concentration of ^{137}Cesium in the samples. Previous research (Ritchie, McHenry, and Gill, 1973; Pennington, et al., 1976) has shown that the ^{137}Cs first appeared in the sediment profile in 1954 the year nuclear weapons testing began and that ^{137}Cs concentrations in the sediment profile peaked in 1964 when radioactive fallout was maximum (Figure 2). These radioactive fallout measurements for strontium-90 at Birmingham, Alabama (Figure 2) are the nearest continuous measuring station to the study area. The ratio of ^{137}Cs to ^{90}Sr in fallout is relative constant; a value of 1.6 was used to convert ^{90}Sr to ^{137}Cs. Using the ^{137}Cs distribution pattern in the sample profile, we estimated rates of sediment accumulation since 1964 and where possible sediment accumulation rates since 1954 for each profile.

In calculating sediment accumulation rates, a threshold value of 0.1 pCi/g was taken as positive identification of the first introduction of ^{137}Cs in 1954. The 1964 peak concentration was assumed to occur at the middle of the 10-cm increment in which it was found.

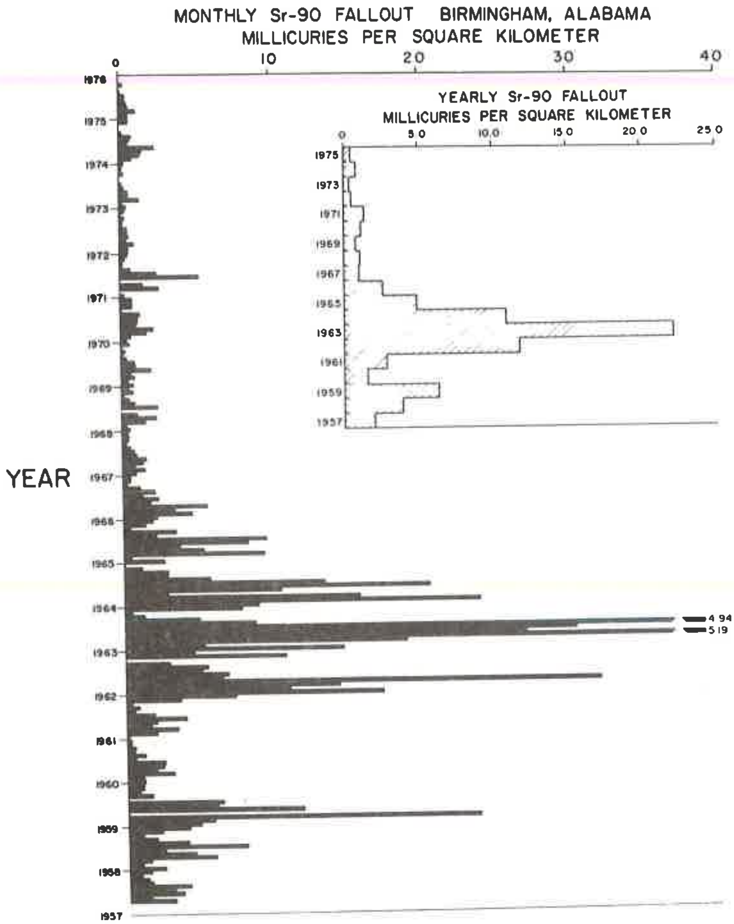

Figure 2. Yearly distribution patterns of fallout strontium-90 at Birmingham, Alabama.

RESULTS AND DISCUSSION

The distribution of ^{137}Cs in the sample sediment profiles (Figures 3-5) showed that calculated sediment accumulation rates (Table 1) since 1964 ranged from 0.4 to over 7.7 cm/yr. At one Wasp Lake site, the peak ^{137}Cs concentration was not found in the upper 100 cm of deposited sediment. Sediment accumulation rates estimated from 1954 to 1964 ranged from 1.5 to 4.5 cm/yr. At five sites, sediment profiles of 80, 90, 100, 100, and 50 cm depth were collected, but these depths did not include the total depth at which ^{137}Cs had been deposited.

The average rate of sediment accumulation since 1964 in the

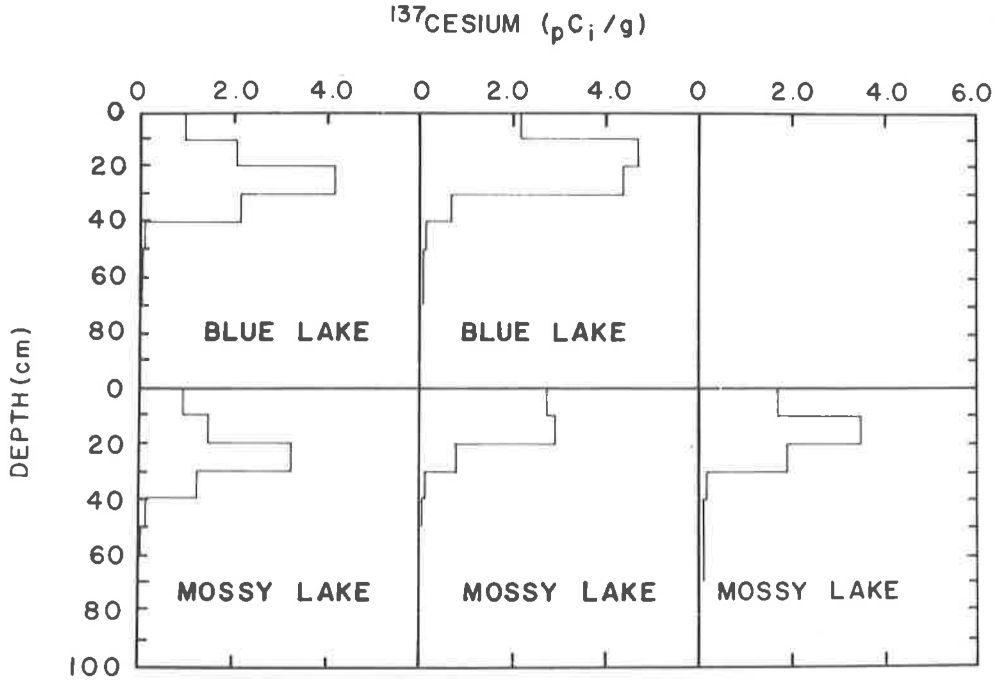

Figure 3. Distribution of ^{137}Cs in sediment profile in Blue and Mossy Lakes in Bear Creek watershed in Mississippi. The left-hand graph is the upstream position in the lake; the right-hand graph is the mid-lake position (if sampled).

six lakes approximated 3. 3 cm/yr. The average rate of sediment accumulation from 1954 to 1964 was about 3. 1 cm/yr. The sediment accumulation rates from 1954 to 1964 and from 1964 to 1977 indicated little change has occurred in the sediment accumulation rates since 1954. The fact these rates have remained nearly constant over these time periods indicates current agricultural practices have not significantly affected the amount of eroded material being deposited in these lakes. However, these sediment accumulation rates do pose a major problem for these lakes since most of them are less than 3 m deep. Assuming a sediment accumulation rate of 3 cm/yr most of these lakes would be filled with sediment within the next 100 years. This is a relatively short period considering that some of these lakes probably have existed since the late Pleistocene age (> 7000 years).

The sediment accumulation rates since 1964 for the lakes along the main channel (Three-mile, Six-mile, and Wasp) are two to three times greater than those for lakes not directly connected with the channel (Mossy and Macon) and Blue Lake, where Bear Creek begins. This indicates that significant amounts of sediment are moving in the channel and are being deposited in the lakes. Field studies of sediment

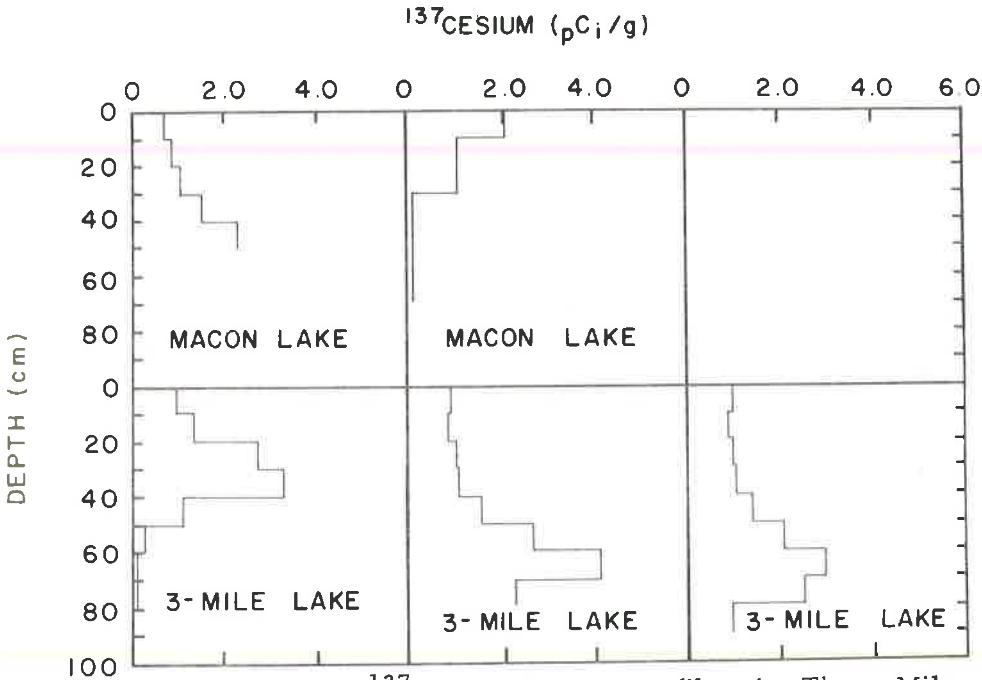

Figure 4. Distribution of ^{137}Cs in sediment profiles in Three-Mile and Macon Lakes in Bear Creek watershed in Mississippi. The left-hand graph is the upstream position in the lake; the right-hand graph is the mid-lake position (if sampled).

load in the stream also show much movement of suspended sediment. Land areas immediately surrounding the lakes are the source of some of the depositing sediment, but more sediment is moving into the stream system from upstream, field and channel erosion.

In the individual lakes, sediment accumulation rates increase somewhat downstream. The 1964 to 1977 rate increased from 1. 9 cm/yr at Blue Lake, the stream origin, to over 5. 6 cm/yr at Wasp Lake near the mouth of the stream. Results were similar for the 1954 to 1964 accumulation rates. Part of the sediment accumulating in Wasp Lake may be due to sediment moving in from the Yazoo River, since during high water periods, water flows from the Yazoo River into Wasp Lake. How much sediment in Wasp Lake is derived from the river and how much is moving in from Bear Creek is difficult to estimate. However, since sediment accumulation rate in Six-mile Lake is estimated to be 3. 9 cm/yr, the 5. 6 cm/yr in Wasp Lake would indicate not more than 1. 7 cm/yr was derived from the river and probably less. This conclusion is supported by the sediment accumulation rate from 1954 to 1964 which showed Six-mile Lake had an averaged 3.5 cm/yr deposited while Wasp Lake had an average of something like 4. 6 cm/yr

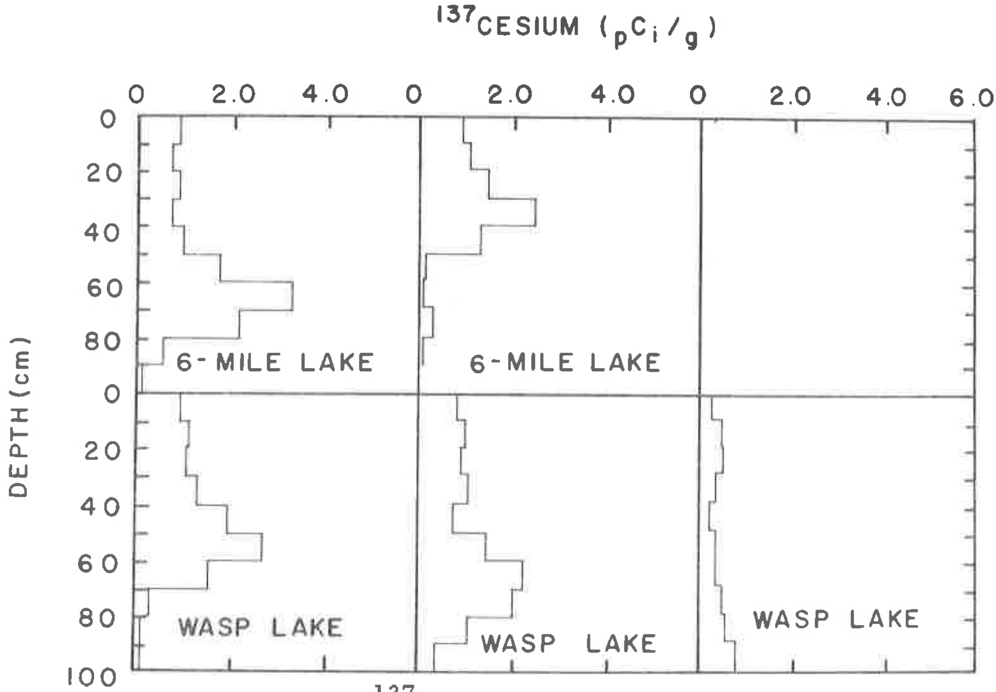

Figure 5. Distribution of ^{137}Cs in sediment profiles in Six-Mile and Wasp Lakes in Bear Creek watershed in Mississippi. The left-hand graph is the upstream position in the lake; the right-hand graph is the mid-lake position (if sampled).

Table 1. Estimated rates of sediment deposition in the lakes of the Bear Creek watershed for the periods 1954-1964 and 1964-1977 based on ^{137}Cs data.

Lake	Location					
	Upper		Middle		Lower	
	1954-1964	1964-1977	1954-1964	1964-1977	1954-1964	1964-1977
	cm/yr					
Blue	2.5	1.9	--	--	3.0	1.9
Three-Mile	2.5	2.7	2.5	5.0	>1.5	5.0
Six-Mile	2.5	5.0	--	--	4.5	2.7
Wasp	2.5	4.2	*	>7.7	3.5	5.0
Mossy	2.5	1.9	2.5	1.2	2.5	1.2
Macon	*	>3.8	--	--	2.5	0.4

*Peak ^{137}Cs concentration at bottom of profile sampled. No rates can be assigned to the 1954-64 period.

deposited. The Yazoo River does cause some sedimentation in the lower end of Wasp Lake but the amounts are probably not significant as compared with the amount moving in from Bear Creek.

CONCLUSIONS

Measured sediment accumulation rates in the lakes in the Bear Creek watershed during the past 23 years averaged from 1.4 to over 5.6 cm/yr. If sediment continues to accumulate at these rates over the next 100 years, most of these lakes will be filled with sediment. Considering that some of these lakes may have been formed as much as 3000 years ago, the present rate of deposition is significant.

The rates of sediment accumulation increase downstream in the Bear Creek system, indicating that more sediment is being transported in the main stream than is coming from the lateral streams feeding directly into the lakes.

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MACROFAUNA AND FOSSIL PRESERVATION IN THE MAGOFFIN MARINE ZONE, PENNSYLVANIAN BREATHITT FORMATION OF EASTERN KENTUCKY

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ABSTRACT

The Magoffin marine zone of the Pennsylvanian Breathitt Formation is well-exposed along the Daniel Boone Parkway near Hazard, Kentucky. A diverse macrofauna occurs in strata interpreted as transgressive and bayfill facies. Mollusca, Brachiopoda, Echinodermata, Bryozoa, Coelenterata, Foraminiferida, and trace fossils are represented by a total of 73 taxa. Aragonitic preservation occurs in many molluscs.

INTRODUCTION

Fossiliferous zones containing marine invertebrate faunas have been recognized within the Pennsylvanian Breathitt Formation of eastern Kentucky for many years (Jillson, 1919; Hudnall, 1927; Morse, 1931; McFarlan, 1943). The intervals occur within deltaic paleoenvironmental settings and, because of their distinctive nature, they have been widely used for correlations and as structural indicators (e. g. Moore, 1944; Prostka, 1965). One such interval is the Magoffin marine zone (Figure 1). As presently conceived (Huddle and Englund, 1966; Horne et al., 1971), the Magoffin zone includes two fossiliferous sequences originally defined by Morse (1931): the Magoffin Beds and the overlying Salt Lick Beds. At some outcrops these two fossiliferous zones are not clearly separable, whereas at others two or more closely spaced marine horizons may occur within the Magoffin outcrop belt (Huddle and Englund, 1966; Horne et al., 1971). Such local changes are to be expected in deltaic settings.

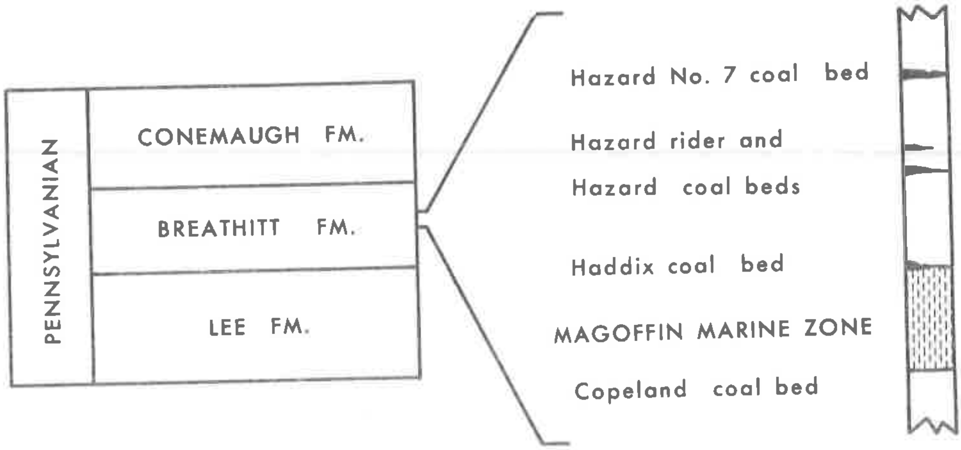

Figure 1. Generalized Pennsylvanian stratigraphy in eastern Kentucky; important Breathitt Formation coal beds in Hazard, Ky. area; position of Magoffin marine zone. Thicknesses of coal beds and intervening strata variable; complete Magoffin sections ranging from 19-23 m.

The descriptive paleontology of the Magoffin invertebrates has not been adequately studied. Morse (1931) described 16 taxa from his type Magoffin Beds and 7 taxa from his type Salt Lick Beds, and some of Morse's collections have been subsequently redescribed (e.g. Newell, 1937). Later Strimple and Knapp (1966) described a new crinoid species from additional Magoffin material. Faunal lists are available for three localities in Martin County, Kentucky (Huddle and Englund, 1966). Other studies usually contain some reference to the general taxonomic content of the Magoffin zone (e. g. Adkison, 1957; Welch, 1958; Bergin, 1962; Patterson and Hosterman, 1962; Adkison and Johnson, 1963).

Fresh exposures of the Magoffin zone, along the Daniel Boone Parkway near Hazard, Kentucky (Appendix), have yielded the most diverse macrofauna yet reported from the Magoffin zone. In this paper we outline the zone's paleoenvironmental setting at Hazard, comment on and list the fauna, and note preservation of original skeletal aragonite among the molluscs. The Parkway sections, 1. 3 to 3. 8 kilometers apart, were measured using a hand level and staff. Fossiliferous portions of the Magoffin zone were sampled vertically, in 50 cm intervals, at three or four stations along each outcrop. Bulk and unweathered samples were extracted from the middle of each interval and ranged between 1/16 and 1/8 m^3 of rock. Specimens of each taxon were transported to the lab for further analysis.

Acknowledgments

We thank M. Gordon, Jr. and E. L. Yochelson for help with initial determinations of the genera, and W. D. Carpenter for hospitality in the Hazard area. This work would have been impossible without the active input and support from J. C. Horne, J. C. Ferm, and the Carolina Coal Group.

GEOLOGIC SETTING

Near Hazard, the Magoffin marine zone ranges from 19-23 m in thickness and includes distinctive coarsening-upward sequences of shale, siltstone, and sandstone; the interval is bounded above and below by delta plain deposits (Horne and Ferm, 1978). In such deltaic settings, coarsening-upward sequences are generally interpreted as recording bay or pro-delta deposition (Brown, 1973; Baganz, 1975; Horne and Ferm, 1978).

Along the Daniel Boone Parkway, detailed mapping (Dennis, 1978) has led to the recognition of three major Magoffin facies based upon lithic criteria (Figure 2). A basal transgressive facies includes relatively thin and widespread strata. Rock types include a coal or carbonaceous shale with an underlying root-penetrated zone, and limestone and shale. Concentrations of marine macrofauna occur in the limestone and at one level within the shale.

A second sequence represents a lobe-building facies and records deposition at the distal portions of either a splay or distributary mouth bar. This is a coarsening-upward sequence, characterized by thinly-bedded and burrowed clastics nearly devoid of macrofauna. Where exposed, the lobe-building facies always gradationally overlies the transgressive phase.

A bayfill facies may also gradationally overlie the transgressive facies; bayfilling took place concurrently with lobe building. This third sequence begins with shales containing banded ironstones. Bayfill strata also coarsen upwards in section, where they are characterized by predominant silty shale with limestone concretions. Siltstones and sandstones cap this sequences, and the facies contains an abundant marine macrofauna throughout its thickness.

Scour and channel deposition have "cut out" parts of the bayfill facies in the Parkway exposures. The cross-bedded sandstones preserved from these events (Figure 2) are devoid of macrofauna. Thus, the fauna listed in Table 1 comes from the transgressive and bayfill facies.

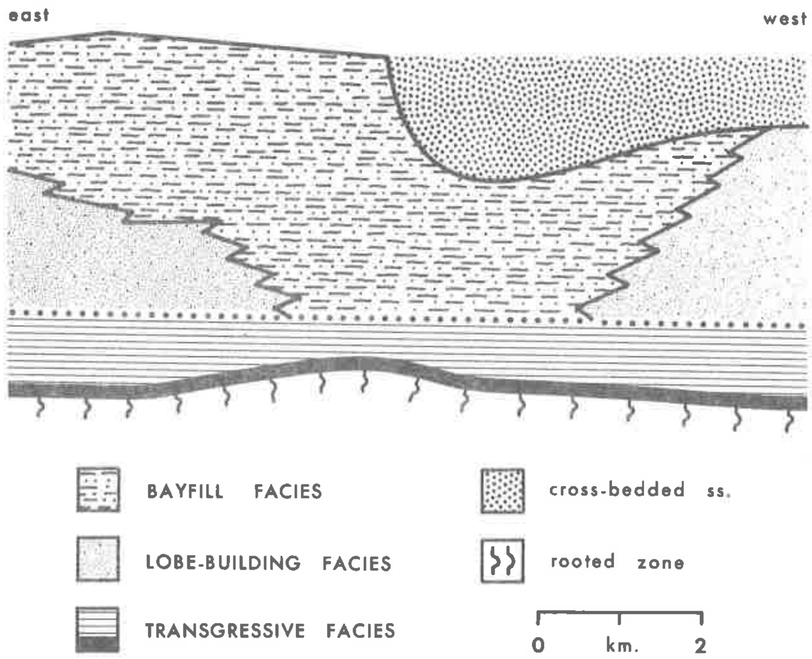

Figure 2. Geometry of major facies recognized in Magoffin marine zone in exposures along Daniel Boone Parkway near Hazard, Ky. Localities and description of lithic types in text. Dotted line represents gradational contact between transgressive and overlying facies. Vertical exaggeration variable;- ca. 1000 x in transgressive facies and 200 x in succeeding facies.

FAUNA

Our collections have considerably enlarged the number of taxa recovered from the Magoffin (see Huddle and Englund, 1966, p. 38-39) and have added 3 phyla to those known from these beds (Table 1). A total of 73 taxa includes representatives of Mollusca, Brachiopoda, Echinodermata, Bryozoa, Coelenterata (Paraconularia), and Foraminiferida, as well as trace fossils. Bivalved molluscs and crinoids dominate in the transgressive facies, whereas brachiopods and bivalves are the most prominent elements in the bayfill sequence.

The Bivalvia and Gastropoda are the most abundant and taxonomically diverse of the molluscs. Common bivalves include Astartella concentrica, Nuculopsis girtyi, Pteronites sp., and Posidoniella sp. Important gastropods in the Hazard area Magoffin are Glabrocingulum cf. G. (G.) grayvillense, Straparollus cf. S. (Amphiscapha) catilloides,

Table 1. Taxa of the Magoffin Marine Zone in Exposures Along the Daniel Boone Parkway Near Hazard, Kentucky.

Foraminiferida
- *Serpulopsis* ? sp.

Coelenterata
- Scyphozoa
  - *Paraconularia* sp.

Bryozoa
- rhabdomesid
- fenestellid
- encrusting form

Brachiopoda
- Inarticulata
  - *Crania modesta* White and St. John, 1867
  - *Lindstroemella patula* (Girty, 1911)
  - *Trigonoglossa nebrascensis* (Meek, 1872)
- Articulata
  - *Anthracospirifer* cf. *A. matheri* (Dunbar and Condra, 1932)
  - *Beecheria* sp. indet.
  - *Cleiothyridina* cf. *C. pecosii* (Marcou, 1858)
  - *Crurithyris planoconvexa* (Shumard, 1858)
  - *Composita* "*subtilita*" (Hall, 1852)
  - *Derbyia crassa* (Meek and Hayden, 1859)
  - *Desmoinesia* cf. *D. missouriensis* (Girty, 1915)
  - *Eolissochonetes* sp.
  - *Hustedia* cf. *H.* "*mormoni*" (Marcou, 1858)
  - *Juresania* ? sp.
  - *Linoproductus* sp.
  - *Plicochonetes* sp.
  - *Pulchratia* sp.
  - *Punctospirifer kentuckyensis* (Shumard, 1855)
  - *Sandia* sp.
  - *Schizophoria* cf. *S. resupinoides* (Cox, 1857)

Mollusca
- Scaphopoda
  - Scaphopod sp. a, ribbed form
  - Scaphopod sp. b, smooth form
- Bivalvia
  - *Acanthopecten* ? sp.
  - *Astartella concentrica* (Conrad, 1842)
  - *Cypricardinia* ? *carbonaria* Meek, 1871
  - *Dunbarella* sp.
  - *Grammatodon* cf. *G. carbonaria* (Cox, 1857)
  - *Lima retifera* Shumard, 1858
  - *Limapecten morsei* Newell, 1937
  - *Nuculopsis girtyi* Shenck, 1934
  - *Nuculopsis* sp.
  - *Paleyoldia glabra* (Beede and Rogers, 1898)

Permophorus? sp.
Phestia bellistriata (Stevens, 1898)
Phestia sp.
Posidonia sp.
Posidoniella sp.
Promytilus cf. *P. vetulus* Newell, 1942
Prothyris sp.
Pteronites sp. indet.
Schizodus sp. a, large form
Schizodus sp. b, small form
Septimyalina sp.
Solemya sp. indet.
Wilkingia terminale (Hall, 1852)
Wilkingia sp.
bivalve, gen. sp. indet.

Gastropoda
Bellerophon cf. *B.* (*Pharkidonotus*) *percarinatus* (Conrad, 1842)
Euphemites cf. *E. nodocarinatus* (Hall, 1858)
Glabrocingulum cf. *G.* (*Glabrocingulum*) *grayvillense* (Norwood and Pratten, 1855)
Ianthinopsis cf. *I. paludinaeformis* (Hall, 1858)
Ianthinopsis cf. *I. regularis* (Cox, 1857)
Knightites (*Cymatospira*) *montfortianus* (Norwood and Pratten, 1855)
Orthonema? sp.
Shansiella carbonaria (Norwood and Pratten, 1855)
Straparollus cf. *S.* (*Amphiscapha*) *catilloides* (Conrad, 1842)
Trepospira (*Trepospira*) *illinoiensis* (Worthen, 1884)
Worthenia tabulata (Conrad, 1835)

Cephalopoda
Gastrioceras occidentale (Miller and Faber, 1892)
Metacoceras? sp.
Mooreoceras cf. *M. normale* (Miller, Dunbar, and Conrad, 1933)
Pseudoparalegoceras compressum (Hyatt, 1891)
Pseudorthoceras knoxense (McChesney, 1860)
nautiloid, gen. sp. indet.

Echinodermata
Echinoidea
echinoid plates and spines
Crinoidea
Diphuicrinus patina Strimple and Knapp, 1966
Paragassizocrinus cf. *P. diculus* Strimple, 1960
crinoid, gen. sp. indet.
crinoid plates

Trace Fossils
Zoophycos
crawling traces
burrow structures

Trepospira (T.) illinoiensis, and Euphemites cf. E. nodocarinatus. Cephalopods include nautiloids, both orthoceraconic and coiled forms, and ammonoids. The ubiquitous Pseudorthoceras knoxense is by far the most abundant cephalopod in these Magoffin sections. Scaphopods are represented by at least two species, though only five specimens were recovered.

The most abundant inarticulate brachiopod is Lindstroemella patula; identification is based upon examination of the interior of a single brachial valve. Externally, L. patula is indistinguishable from at least one other common Pennsylvanian acrotretid species (Sturgeon and Hoare, 1968) and, because only external features are normally observable in our acrotretid specimens, it is possible that other taxa have been grouped with L. patula.

The articulate brachiopods are predominantly Strophomenida and Spiriferida. The productaceans Desmoinesia cf. D. missouriensis and Sandia sp. are abundant in the bayfill sequence and one of the chonetaceans, Eolissochonetes sp., is the most frequently collected Magoffin invertebrate. Common spiriferids include Anthracospirifer cf. A. matheri, Hustedia cf. H. "mormoni", and Composita "subtilita."

Echinoderms and bryozoans are an important part of the Magoffin fauna. At least three crinoid species are presented by numerous solitary plates and rare intact calices and arms. Relatively large, fan-like fenestellids and smaller club-shaped rhabdomesids are the most frequently encountered bryozoans. Encrusting bryozoans occur on several brachiopod valves.

Tolypamminid foraminiferids, probably Serpulopsis, are present on the exterior surfaces of brachiopods and are particularly abundant on the brachial valves of numerous individuals of Eolissochonetes sp. Notably absent from our collection are any representatives of corals and trilobites. Of equal note is the rarity of linguloid brachiopods. We found only two linguloids during extensive sampling; and only one other specimen, from the carbonaceous shales of the transgressive facies, has been brought to our attention (J. C. Horne, pers. comm., 1978).

Biogenic sedimentary structures occur through most of the Magoffin interval and extreme bioturbation helps to delimit the basal bayfill strata. Zoophycos and crawling traces are present throughout the bayfill sequence; Zoophycos is most common in the siltier shales near the top of this facies whereas crawling traces are most abundant in the finer-grained and lower parts of this sequence. Ovoid fecal pellets fill the chambers of one ammonoid specimen.

PRESERVATION OF THE FAUNA

Good to excellent preservation is a common attribute of the Magoffin macrofauna. Most striking in our collections is the occurrence aragonite in a number of mollusc specimens. Chalky, white, aragonitic shell material occurs in examples of all four molluscan classes present. Most specimens are considerably exfoliated, appear "rotten" and are not suitable for detailed analysis. However, *Astartella* and the bellerophontacean gastropods yield evidence of entirely aragonitic shell layers. One coiled cepholopod fragment bears a thin, chalky outer layer and an inner layer of lustrous nacre. Many of the speccimens of *Paleyoldia* show traces of an outer chalky layer and an inner aragonitic layer that is distinctly pale blue in color; this tint most likely denotes original coloration. Calcitic molluscs also are present in our collections (Table 1). Two bivalve specimens, questionably assigned to *Permophorus*, retain portions of an external, opisthodetic ligament.

Molds are a common mode of preservation for both the molluscs and brachiopods only in the coarser-grained clastics and ironstone bands. All in all, preservation helps to make the Magoffin marine zone well-suited for detailed paleoenvironmental analysis (Dennis, 1978).

DISCUSSION

Thus we have noted another interval and area where aragonitic preservation may be found in late Paleozoic invertebrates. Aragonitic preservation has also been found in molluscs from the Kendrick shale, another Breathitt Formation marine zone which begins some 60-70 m beneath the Magoffin strata in eastern Kentucky (see Stehli, 1956; Grandjean *et al.*, 1964; Yochelson *et al.*, 1967). Similar rock types, biotas, and similar paleoenvironmental settings are recorded in the Magoffin and Kendrick intervals (Huddle and Englund, 1966; Baganz, 1975; Horne and Ferm, 1978); syn- and post-depositional environments must promote the preservation of aragonite in these marine bayfill units.

Our observations of preservation were incidental to other work on the Magoffin, and it is obvious that additional studies in these marine bayfill sequences could substantially enlarge our knowledge of skeletal mineralogy, especially in late Paleozoic molluscs. Special care must be taken to use cores or extremely fresh outcrops in this type of work. The Parkway roadcuts were made during 1971-1973 (B. Baganz, pers. comm., 1978) and considerable chalky aragonite, with obliteration of skeletal microarchitecture, typified our Magoffin molluscs collected during the summers of 1975 and 1976. By the spring of 1978, the number of aragonitic molluscs at or near the surface of Parkway exposures was strikingly reduced.

Detailed correlations of the Breathitt Formation, and of its marine zones, are still in flux. Ammonoids, crinoids, and at least one productacean from the Kendrick shale suggest that this lower part of the formation is of Morrowan, probably late Morrowan, age (Furnish and Knapp, 1966; Yochelson *et al.*, 1967). The abundance of *Eolissochonetes* in the Magoffin beds suggests that this younger zone is of

Desmoinesian age (Moore, 1965) which is compatible with the known stratigraphic ranges of the other macrofauna identified. Geographic changes in the Magoffin strata clearly indicate that this zone is not synchronous throughout its outcrop belt.

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APPENDIX: MAGOFFIN LOCALITIES STUDIED

HN-11: Roadcut on south side of Daniel Boone Parkway, 0.2-0.3 km west of bridge crossing the North Fork, Kentucky River; Perry Co., Ky. Base of Magoffin approx. 65 m above road level. 21 m section representing transgressive, lobe-building, and bayfill facies.

KR-3: Roadcut on south side of Daniel Boone Parkway, 1.9-2.0 km west of bridge crossing the North Fork, Kentucky River; Perry Co., Ky. Base of Magoffin between first and second bench above road level (west end of cut). 23 m representing transgressive, lobe-building, and bayfill facies.

KR-7: Roadcut on south side of Daniel Boone Parkway, 2.8-3.0 km west of bridge crossing the North Fork, Kentucky River; Perry Co., Ky. Base of Magoffin not exposed; upper portions occur at road level. 14 m, representing bayfill and portion of lobe-building facies.

HYE-3: Roadcut on south side of Daniel Boone Parkway, 3.3-3.4 km east of point where Perry - Leslie County line crosses Parkway; Perry Co., Ky. Base of Magoffin at road level. 20 m of transgressive and bayfill facies.

HYE-11: Roadcut on north side of Daniel Boone Parkway, 2.1-2.2 km east of point where Perry - Leslie County line crosses Parkway; Perry Co., Ky. Base of Magoffin at road level (east end of cut). 11 m, representing transgressive and bayfill facies. Channel-deposited sandstone overlying bayfill sequence.

HYE-23: Roadcut on south side of Daniel Boone Parkway, 1.7-1.8 km west of point where Perry - Leslie County line crosses Parkway; Leslie Co., Ky. Base of Magoffin at road level (middle of cut). 16 m of transgressive and lobe-building facies; channel-deposited sandstone above lobe-building sequence.

CROSS-STRIKE STRUCTURAL DISCONTINUITIES IN

THRUST BELTS, MOSTLY APPALACHIAN

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ABSTRACT

Cross-strike structural discontinuities (CSDs) are fundamental parts of several thrust belts. CSDs are best recognized as structural lineaments or alignments of disruptions in structural or geomorphic patterns. Detailed field studies show that the discontinuities are complex zones of diverse sizes, represent different degrees of basement involvement, and have various structural, geophysical, chemical, and

Table 1. Preliminary characterizations of CSDs. O = phenomenon is present in, and an outstanding characteristic of, the named CSD. P = present. - = unevaluated. A = absent. * = property by which the CSD was first recognized. MP = McAlevys Fort - Port Matilda CSD, Pennsylvania. TM = Tyrone - Mount Union CSD, Pennsylvania. EB = Everett - Bedford CSD, Pennsylvania. PB = Petersburg CSD, West Virginia. PA = Parsons CSD, West Virginia. 38 = extension of 38th Parallel lineament into West Virginia. LN = Little North Mountain CSDs, Virginia. MO = Modoc CSD, West Virginia. WS = White Sulfur Springs CSD, West Virginia. CV = Covington CSD, West Virginia and Virginia. AN = Anniston CSD, Alabama. KC = Kelly Creek CSD, Alabama. HV = Harpersville CSD, Alabama. IR = Irish CSDs. SC = Southern Chilean CSDs (parautochthonous foreland). Dots before and after numbers in the "Spacing" line indicate that the numbers apply to the columns under which the dots appear.

other characteristics. CSDs are not single surfaces, such as tear faults, joint zones, or outcropping unconformities. They can affect concentration and quality of hydrocarbons, mineralizing fluids, or ground water.

INTRODUCTION

In a symposium in 1976 we described preliminary results and speculations from current field studies of major cross-strike structural discontinuities (CSDs), in allochthonous or parautochthonous parts of three orogens. The CSDs described are map-scale structural lineaments or alignments, at high angles to regional strikes, and best recognized as disruptions in strike-parallel structural or geomorphic patterns (see below and Table 1). This paper summarizes results reported at the 1976 symposium, and subsequent work, (1) to promote exchanges between the authors and other field geologists working on similar structures in widely separated areas, and (2) to attempt a preliminary characterization of a class of structures of growing academic and economic interest. This paper contains some preliminary generalizations, and cites available work on the areas described at the session. Table 1 summarizes the characteristics of 15 CSDs or groups of CSDs. Details have been or will be published separately by individual authors. Work on most of our CSDs is incomplete, so only 13 of the 81 cited references are articles in established journals or series; all are readily available however, if one knows that they exist.

Phenomenon	MP	TM	EB	PB	PA	38	LN	MO	WS	CV	AN	KC	HV	IR	SC
	Named CSD or group of CSDS														
Size (km)															
Length	70+	160+	130	80	55	175+	11-25+	58	48+	36+	270	50	240	200	5-80
Width	1-2	1	1	8	10	80	1-4	5-12	3	-	3-4	1-12	3-4	5	0-1
Depth	-	-	-	5	3	-	-	4	4	-	-	5+	-	-	5
Spacing between CSDs	..20........20-90.			-	-	-	60-90	...15-25......			-	-	-	30	10-15
Structural effects															
Major scale	P*	P	P*	O*	O*	O	O*	O*	O*	O*	O	O*	O	O*	O
Minor scale	P*	P	-	-	P	-	P	A	A	A	P	P	P	-	O
Microscopic	-	-	-	-	-	-	-	-	-	-	-	-	-	-	P
More fracturing	P	P	P	-	P	P	-	-	-	-	-	-	-	-	O*
Dip separation	A	A	A	A	A	-	P	A	A	-	P	P	P	-	P
Strike separation	A	A	P	A	A	P	-	-	-	-	-	P	-	-	O
Faulting reversal	P	P	-	-	A	-	P	-	-	-	A	P	A	-	P
Geophysical effects															
Gravity anomalies	-	P	P	P	P	P	-	P	P	P	P	-	-	P	P
Magnetic anomalies	-	P	P	-	-	-	P	P	-	P	P	-	-	P	-
Seismic activity	A	P	P	-	-	-	-	-	-	-	P	P	P	-	P
Fluids: Effects on...															
Sulfides	-	P	P	-	-	P	-	-	-	-	P	A	P	P	P
Quartz, calcite	P	P	P	-	-	-	-	-	-	-	-	P	-	-	P
Magmatic activity	A	A	A	A	A	P	A	-	-	-	-	A	-	P	P
Thermal waters	A	A	A	-	-	P	A	-	-	-	A	A	A	P	-
Oil and gas yields	-	-	P	P	-	-	A	-	-	-	A	A	A	-	P
Water-well yields	P	-	-	-	P	-	-	-	-	-	P	O	P	-	-
Water-well quality	-	-	-	-	P	-	-	-	-	-	A	-	A	-	-
Other effects															
Sedimentation patterns	-	P	-	-	-	P	-	-	-	-	P	A	P	-	-
Extent into basement	-	-	P	-	A	-	-	-	-	-	P	A	P	P	P
At outcrop scale	-	P	-	P	A	-	P	-	-	-	A	P	A	P	P*
Microscopic	-	-	-	-	-	-	-	-	-	-	-	-	-	-	P
Satellite photos	O	O	O	O	P	O*	P	P	P	O	P*	O	O*	P	P
Air photos	P	A	A	O	P	A	P	P	P	O	P	P	P	-	O
Geomorphic	P	O*	P	O	A	P	P	P	P	O	P	P	P	P	O
Drainage anomalies	P	P	P*	O	P	P	P	P	P	P	P	P	P	P	P
Most likely origin(s)															
Step-up along trend	-	P	-	P	P	P	P	A	P	P	-	O	-	-	-
Tear-fault zone	-	-	-	-	A	P	-	-	-	-	P	O	A	-	P
Active basement	-	-	P	A	A	-	A	-	-	-	P	A	P	P	P
Passive basement	P	-	-	-	-	P	-	-	-	-	P	A	P	-	-
Other	-	-	-	-	-	-	-	P	A	A	-	-	-	-	-

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GENERALIZATIONS

CSDs form an important class of major structures, located in, but in some places extending beyond, petroliferous and metalliferous thrust belts. Individual CSDs are too poorly known for us to decide yet whether all have the same or similar relationships to orogeny or plate tectonics, to figure a typical example, or to formulate an exact definition. Also, the list in Table 1 is geographically biased toward the central and southern Appalachians, and is surely incomplete (see, for example, Dahlstrom, 1970, Price and Lis, 1975, Lis and Price, 1976, and Price and Kluvver, 1974; also oral communications from R. A. Price on the Canadian Rockies and H. P. Laubscher on the Jura, at the Penrose Conference on Relative Chronology of Thrusting in the Southern and Central Appalachians). However, we predict that CSDs are generally characteristic of allochthonous and perhaps parautochthonous fold and thrust belts.

CSDs should not be classed with simple tear faults or joint zones and can be considerably more complex than either. CSDs' differences may be as important as their similarities. CSDs are not single surfaces. Many may include hundreds of cubic kilometers of rock, which are volumes of economic and academic importance in their own rights, with their own physical and chemical properties. Some CSDs are deep, nearly vertical zones, but others are apparently wider than deep. Some but not all CSDs contain unusually fractured rock.

Basement (suballochthon rock and structure) may or may not be involved in any given CSD. Basement involvement can be active, the basement becoming part of the final CSD, or it can be passive, the older inactive basement structures influencing formation of structures in an allochthon passing overhead. Basement involvement might also change from active to passive to nonexistent, craton-ward along a single CSD. Perhaps active and passive basement involvement can be distinguished by comparing the gravity, magnetic, and seismic signatures, which can characterize the rocks over which an allochthon now sits, with early-acquired chemical signatures, which, if identifiable, can be characteristics of the rocks over which the allochthon once passed.

For both academic and economic reasons, it is important to determine what CSDs do to the rock at and within their boundaries, in both penetrative and non-penetrative senses. To what predictable extents do CSDs or their boundaries possess unusual degrees of fracture porosity, fracture permeability, variation in fold style or intensity, hydrothermal mineralization, cementation, or recrystallization?

It is not yet clear how one best prospects for CSDs. Because of their diverse characters and expressions (see Table 1), the process of finding them reliably and of determining or predicting their properties and extents is much more complex than just seeking straight lines or alignments on photographs and maps. Not all CSDs are visible on aerial photographs, and not all are outstandingly visible on satellite imagery (Table 1). On the other hand, the field methods being used to define and characterize several CSDs are probably too time-consuming to be applied to more than a few field areas, on which faster methods can then be tested before their application to new areas.

TERMINOLOGY

Unnecessary confusion has already arisen over terminology, partly because workers are scattered and CSDs are complex, and partly because study of CSDs spans several branches of geology not always in close contact (for instance, detailed field mapping and interpretation of LANDSAT images). Terminology standardization is beyond the scope of this paper but should be done soon, while the number of geologists working on CSDs is still small.

A few CSDs appear to form where a detachment fault cuts upsection to shallower structural levels; this process can occur in two ways: (1) The fault can step up across regional trend, on a reverse fault or fault zone that strikes perpendicular to the transport direction, or (2) it can step up along regional trend, on a strike-slip fault or fault zone that strikes parallel to the transport direction. The first way produces an anticline, and the second forms an end or lower edge of a detached block. For schematic examples, see Harris's (1970, page 171) diagrams. At least 11 terms are in common or local use for one or both of those methods of cutting upsection: ramp, step, riser, longitudinal fault, longitudinal step, step fault, cross fault, tear fault, transverse fault, transverse step, and domain boundary. The words "transverse" and "longitudinal" are surprisingly confusing in the context of CSDs, because a longitudinal step (usually a reverse fault, though the term has been used for a strike-slip fault) has transverse motion. Some writers refer to the structure and the motion on it almost interchangably. Having one term to include both types of steps seems unnecessarily confusing. For cross-strike structural discontinuities whose nature and cause are not clear, the general term "structural lineament" seems applicable and conformable with recommended usage (for example, see Lattman, 1958, and Werner, 1976).

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