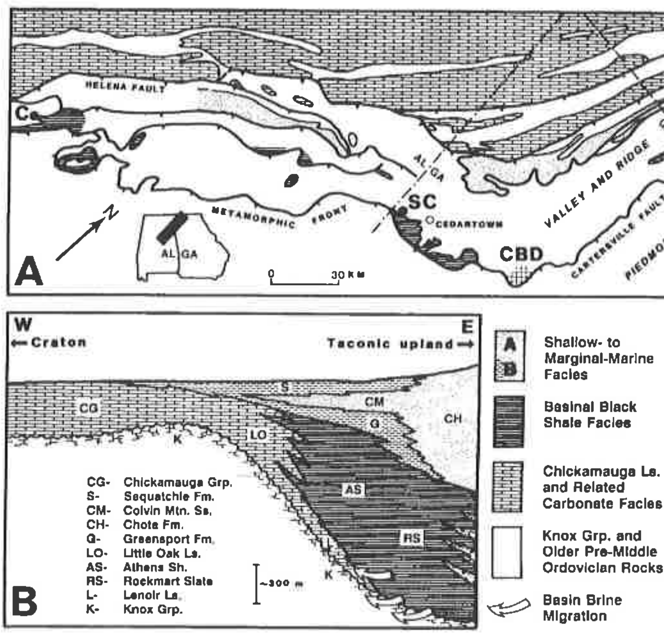
SOUTHEASTERN GEOLOGY



VOL. 33, NO. 4 OCTOBER 1993

SOUTHEASTERN GEOLOGY

PUBLISHED

at

DUKE UNIVERSITY

Editor in Chief:

Duncan Heron

This journal publishes the results of original research on all phases of geology, geophysics, geochemistry and environmental geology as related to the Southeast. Send manuscripts to **DUNCAN HERON, DUKE UNIVERSITY, DEPARTMENT OF GEOLOGY, BOX 90233, DURHAM, NORTH CAROLINA 27708**. Phone 919-684-5321, Fax 919-684-5833, E Mail heron@rogue.geo.Duke.edu. Please observe the following:

1) Type the manuscript with double space lines and submit in duplicate.
2) Cite references and prepare bibliographic lists in accordance with the method found within the pages of this journal.
3) Submit line drawings and complex tables reduced to final publication size (no bigger than 8 x 5 3/8 inches).
4) Make certain that all photographs are sharp, clear, and of good contrast.
5) Stratigraphic terminology should abide by the North American Stratigraphic Code (American Association Petroleum Geologists Bulletin, v. 67, p. 841-875).

Subscriptions to *Southeastern Geology* per volume are: individuals - $14.00 (paid by personal check); corporations and libraries - $19.00; foreign $23. Inquires should be sent to: **SOUTHEASTERN GEOLOGY, DUKE UNIVERSITY, DEPARTMENT OF GEOLOGY, BOX 90233, DURHAM, NORTH CAROLINA 27708**. Make checks payable to: *Southeastern Geology.*

SOUTHEASTERN GEOLOGY is a peer review journal.

ISSN 0038-3678

SOUTHEASTERN GEOLOGY

Table of Contents

Volume 33, No. 4

October 1993

SOUTHEASTERN GEOLOGY
V. 33, No. 4, October 1993, p. 161-170

GEOCHEMISTRY OF THE ATHENS SHALE: IMPLICATIONS FOR THE GENESIS OF MISSISSIPPI VALLEY-TYPE DEPOSITS OF THE SOUTHERNMOST APPALACHIANS

JAMES A. SAUNDERS and CHARLES E. SAVRDA

Department of Geology
Auburn University
Auburn, AL 36849-5305

ABSTRACT

Mississippi Valley-Type (MVT) mineralization near the southern terminus of the Appalachians occurs in karstified Cambro-Ordovician Knox Group carbonates that lie beneath the Middle Ordovician Athens Shale. The Athens Shale is the organic-rich deep basinal facies of a foreland depocenter that formed at the onset of Taconic orogenesis and was subsequently deformed during multiple Appalachian orogenic events. Molybdenum, cobalt, and nickel concentrations in the Athens Shale are enriched relative to average shales and are comparable to typical black shales. MVT mineralization in northwest Georgia locally contains >2000 ppm Mo, >500 ppm Ni, and >30 ppm Co, concentrations which are anomalous compared to typical MVT deposits elsewhere. Given the stratigraphic proximity and geochemical similarities of the Athens Shale and the MVT mineralization, along with the presence of hydrothermal pyrite-quartz-calcite-barite-sphalerite veinlets in the Athens, we propose that the Athens Shale was the source of ore-forming solutions. Northwest-directed middle Paleozoic compressional deformation is interpreted to have focused Athens-derived fluids into the paleoaquifer at the top of the Knox Group, resulting in the formation of MVT mineralization in Georgia.

INTRODUCTION

Mississippi Valley-Type (MVT) Zn-Pb-Ba deposits in the southern Appalachians include epigenetic veins and breccia fillings in Cambrian-Ordovician platform carbonates. The concept that MVT deposits formed from metal-rich sedimentary brines that migrated out of basins is now widely accepted (e.g., Hanor, 1979; Sverjensky, 1984; Oliver, 1986). However, a number of the finer points regarding the origin of these deposits are more controversial. For example, until recently, the timing of the formation of Appalachian MVT deposits was not tightly constrained, complicating the determination of the age and location of basins that produced the ore-forming solutions. Oliver (1986) proposed that movement of fluids attending large-scale thrusting during the late Paleozoic could explain the origin and distribution of Appalachian hydrocarbon accumulations and MVT deposits. Evidence of extensive fluid flow during the late Paleozoic is based largely on paleomagnetic data (McCabe and Elmore, 1989; Lu and others, 1990), K-Ar and Ar-Ar ages of authigenic potassium feldspars (Hearn and Sutter, 1985; Hearn and others, 1987), and isotopic and fluid inclusion data (Schedl and others, 1992). However, recent geochronology on Appalachian MVT's (Kesler and others, 1988; Kesler and van der Pluijm, 1990; Nakai and others, 1990, Nakai and others, 1993) indicate that they predate the late Paleozoic Alleghenian orogenic event, which is consistent with some earlier interpretations based on geologic and textural arguments (e.g., Hoagland, 1976; Tieman and Kopp, 1983). Rb-Sr geochronology on sphalerite from the Coy mine (East Tennessee) define a Devonian isochron age of 377±29 Ma (Nakai and others, 1990), significantly older than the Alleghenian orogenesis (~330-250 Ma). Furthermore, recent Pb-isotope data from late

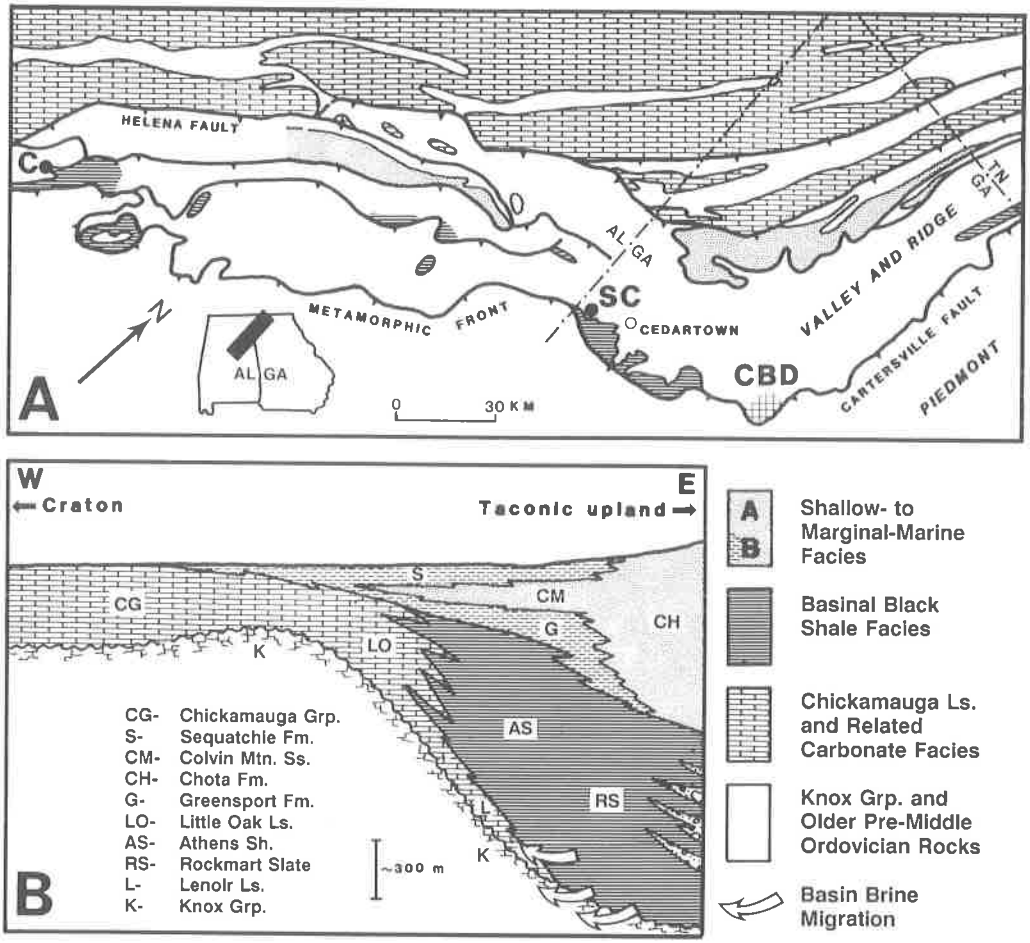


Figure 1. **A-** Map distribution of Middle Ordovician Athens basin facies and stratigraphically subjacent Knox Group carbonates and older pre-Middle Ordovician rocks in the Valley and Ridge province, northeast Alabama and northwest Georgia (modified from Carter and Chowns, 1986). C=Blue Circle quarry near Calera, Alabama; SC= Knox-hosted Shiloh Church Mo-Zn-Pb deposit; and CBD= Cartersville barite district. **B-** Reconstruction of Middle Ordovician stratigraphy and facies relationships of the Athens basin, based on Hoagland (1976), Benson (1986b) and Carter and Chowns (1986). Arrows indicate likely migration pathways of basin brines into karstified Knox aquifer during post-Middle Ordovician foreland compressional deformation.

Paleozoic authigenic K-feldspar overgrowths in east Tennessee indicate that the late Paleozoic fluid event was *not* synchronous with the formation of MVT deposits of the southern Appalachians (Aleinikoff and others, 1993)

The establishment of the mid-Paleozoic age of southern Appalachian MVT deposits makes it possible to evaluate potential source basins for the ore-forming solutions within the context of the developmental history of the southernmost Appalachians. In addition, the unique geochemistry of the southernmost MVT mineralization makes the search for possible source rocks less ambiguous. In this paper we present some initial results from ongoing studies on the Shiloh Church MVT Mo-Zn-Pb deposit in Polk County, Georgia and the geochemistry of the Ordovician Athens Shale, which we believe to be the metal source for this unique MVT deposit. If our interpretations are correct, then this would be an example where it is possible to identify both the source basin and specific formation that provided metals for a particular MVT occurrence.

GENERAL FEATURES OF MVT MINERALIZATION IN GEORGIA

Mississippi Valley-type mineral occurrences in northwest Georgia are hosted by Cambrian-Ordovician carbonates of the Valley and Ridge province (Figure 1). Barite deposits of the Cartersville district (Figure 1) have been the only significant MVT-style mineralization commercially exploited in Georgia. Residual barite is currently mined from clay-rich saprolite derived from highly-weathered Cambrian carbonates (Reade and others, 1980). However, primary epigenetic barite mineralization occurs as veins and as breccia cements in Cambrian dolostone (Rome-Shady Formations, undifferentiated) and is considered to be the source of the residual barite (Kesler, 1950; Rife, 1971; Reade and others, 1980). Both primary barite and residual nodules locally contain abundant quartz and minor sphalerite, pyrite, galena, tennantite, enargite, luzonite, and chalcopyrite (Kesler, 1950). Barite commonly exhibits deformation textures, leading Kesler (1950) to conclude that it pre-dated the youngest deformation in the district.

Exploratory drilling by mining companies in Polk County, Georgia, has encountered minor Zn-Pb-Ba mineralization and led to the discovery of the Shiloh Church Mo-Zn-Pb deposit approximately 10 km southwest of Cedartown (Figure 1). Mineralized zones occur as pyritic sulfide pods hosted by karstified and brecciated carbonates of the Ordovician Knox Group. Ore zones, which are up to 12m wide, have an average grade of approximately 0.33% MoS_2, which is present as either molybdenite or more likely, the amorphous MoS_2 phase, jordisite (Foss and others, 1983). Dolomite, calcite and minor sphalerite, galena, quartz, and bitumen occur as the matrix minerals in the pyrite-rich zones. Combined Pb+Zn values associated with the molybdenum-rich zones are typically <1 wt%, and locally contain >2000 ppm Mo, >500 ppm Ni, >30 ppm Co, and 3400 ppm As. Gossans at the surface contain traces of Mo, Co and Ni and were probably mined as "brown" iron deposits in the 1800's (e.g., Haseltine, 1924).

The geochemical signature (Mo-Zn-Co-Ni-As) of the Shiloh Church mineralization is similar to that of black shales (e.g., Vine and Tourtelot, 1970; Coveney and others, 1987; Coveney and Glascock, 1989; Huyck, 1990), suggesting that a black shale may have been the source of metals. If this interpretation is correct, then the Ordovician Athens Shale is the only black shale of the proper age, significant thickness, paleogeographic distribution, and tectonic setting to have been a metal source for MVT mineralization at Shiloh Church and possibly elsewhere in the Georgia-Alabama Valley and Ridge Province.

GEOLOGY AND GEOCHEMISTRY OF THE ATHENS SHALE

Tectono-stratigraphic Framework

Ordovician sediments in the southern and central Appalachians record the abrupt transition from a regime characterized by shallow-platform carbonate deposition on a passive craton margin to one characterized by rapid subsidence and clastic infilling of foredeep depocenters west of an emerging Taconic orogenic highland (Rodgers, 1970; Read, 1980; Shanmugan and Walker, 1980; Shanmugan and Lash, 1982; Benson, 1986a). The Athens Shale in Alabama and its low-grade metamorphic equivalent in Georgia, the Rockmart Slate, represent the deep basinal facies of a Middle Ordovician sequence that reflects the development and progressive filling of a southern foreland depocenter (Benson, 1986a; Carter and Chowns, 1986), herein referred to as the Athens basin.

Reconstructions of the stratigraphy and facies relations of the Athens basin have been summarized by Benson (1986a,b) and Carter and Chowns (1986) (Figure 1). The basin sequence rests unconformably upon karstified passive-margin shelf carbonates of the Late Cambrian and Early Ordovician Knox Group. In the western part of the basin, towards the craton, the Athens Shale is relatively thin, is separated from the Knox carbonates by shallow- to deep-ramp and basin-margin carbonates (Lenoir Limestone), and grades laterally into similar ramp and basin-margin facies of the Little Oak Limestone. Further west, subaerial exposure and shallow platform conditions were maintained and are represented by the karstified Knox surface and the overlying carbonates of the Chickamauga Group. To the southeast, the Athens Shale/Rockmart Slate package thickens and interfingers with turbiditic coarse clastics that were ultimately sourced from the Taconic highland. In the same direction, the subjacent Lenoir Limestone thins and eventually pinches out, resulting in the juxtaposition of the Rockmart Slate and the Knox carbonates. The Athens-Rockmart argillites and, eventually, the Chickamauga platform carbonates, were progressively overlain by a west- and southwestwardly prograding shallow-marine to terrestrial clastic wedge [Greensport and Colvin Mountain Formations (Alabama), Chota Formation (Georgia), and the Sequatchie Formation] as the basin filled and Taconic orogenic emergence continued in the east-northeast. However, the Athens is now locally unconformably overlain by Devonian and Mississippian strata, reflecting post-Middle Ordovician deformation and uplift associated with ongoing Taconic and/or subsequent Acadian orogenesis.

Depositional Setting and Lithologic Character

The Athens Shale and Rockmart Slate, like their lithostratigraphic counterparts deposited in more northerly foredeep depocenters [e.g., Liberty Hall and Paperville Formations of Virginia (Read, 1980) and the Blockhouse and Sevier Formations in Tennessee (Walker and others, 1983)], reflect oxygen-deficient basinal conditions (Benson, 1986b). Strata are dominated by very dark gray to black, graptolitic, pyritic, laminated, calcareous shales that lack both bioturbation and macro invertebrate benthic body fossils, with the exception of rare scattered brachipods. Although the minor fine-grained carbonate component of the Athens/ Rockmart black shale facies was likely derived via suspension from the platform to the west, fine-grained siliciclastics in the Athens basin were derived from uplands to the northeast and east (Jones and Dennison, 1970). On the basis of stratal thicknesses, Carter and Chowns (1986) estimate basin depths during Athens/ Rockmart deposition to have been 200 to 700 m, which are considerably shallower than paleobathymetric estimates made for more northerly black shale depocenters.

The Rockmart Slate in the vicinity of the Shiloh Church MVT deposit lies closer to the metamorphic front of the southern Appalachians and, hence, has experienced a greater degree of post-Taconic deformation and metamorphism than the Athens Shale exposed in Alabama. In order to reduce the potential for a strong tectonic and/or metamorphic overprint, we opted to focus our study on a relatively less deformed and unweathered Athens Shale section exposed in the Blue Circle quarry near Calera in Shelby County, central Alabama (Figure 1). The Athens Shale at this locality, previously described by Benson and Stock (1986), is in excess of 76 m thick, overlies the Lenoir Limestone, and is unconformably overlain by 0.5 m of weathered Devonian Frog Mountain Sandstone, which, in turn, is overlain by Mississippian Fort Payne Chert. Benson and Stock (1986) divided the Athens here into two informal units. The lower calcareous unit (~15 m) is characterized by decimeter-scale interbedding of dark gray, argillaceous lime mudstones and graptolitic, dark gray to black, laminated, calcareous shales with rare brachiopods and trilobites. The upper unit exceeds 60 m in thickness and is comprised exclusively of dark gray to black, finely laminated, graptolitic shales with variable, albeit relatively lower,

Table 1. Whole Rock Chemical Analyses of the Athens Shale

Major elements (%)	**AS-1**	**AS-5**	**AS-10**	**AS-15**	**AS-18**
SiO_2	56.17	64.75	48.58	49.93	42.11
Al_2O_3	12.74	14.18	9.95	11.67	9.36
Fe_2O_3	4.59	5.22	4.04	4.54	3.59
MnO	0.03	0.02	0.03	0.03	0.03
CaO	7.93	1.26	16.01	12.49	17.78
MgO	2.60	2.43	2.23	3.20	3.97
Na_2O	0.45	0.35	0.32	0.40	0.32
K_2O	3.45	3.87	2.62	2.94	2.60
TiO_2	0.52	0.60	0.41	0.49	0.40
P_2O_5	0.09	0.11	0.09	0.10	0.11
Loss on ig.	11.2	7.0	15.5	14.0	19.5
$C_{org.}$	1.19	1.30	1.12	1.09	1.15
$S_{tot.}$	1.30	1.66	1.31	1.38	1.25
Minor elements (ppm)					
Ba	562	565	582	515	445
Cu	18	27	19	26	16
Zn	261	203	89	103	108
Mo	11	16	12	14	10
Ni	54	91	57	66	48
Co	31	32	29	20	18
V	187	153	132	152	165

carbonate contents. Fine to medium silt-sized clastic grains, which comprise 5 to 10% of all samples, are scattered throughout the matrix, but locally are concentrated along discrete laminae. Although dominated by quartz, the silt fraction also includes subordinate amounts of mica, feldspar, and volcanic (and possibly carbonate) lithic fragments. Pyrite framboids 5 to 10 μm in diameter are disseminated throughout the matrix, as are somewhat larger (~20 μm) euhedral to irregular grains of dolomite.

Locally, narrow (<3 cm) hydrothermal veinlets composed of pyrite, calcite, barite, quartz, and minor sphalerite crosscut bedding orthogonally. One pyrite-rich veinlet contains about 10% barite, 700 ppm Zn, and 25 ppm Mo. Locally, pyrite stringers up to 5 cm long emanate from the veinlets and are parallel to bedding. The pyrite in the veinlets and stringers occur as euhedra up to 1 mm in diameter. Quartz generally encrusts pyrite and coarser crystals point toward the center of the vein, which is filled in turn by barite or calcite. Numerous small (<~3μm) fluid inclusions are present in both quartz and barite but all lack vapor bubbles at room temperature, suggesting a low temperature of formation (<~100°C?), although their small size may have precluded nucleation of a vapor phase (Roedder, 1984).

Geochemistry

Nineteen black shale samples representative of the entire Athens sequence were collected for petrographic and geochemical analysis. Samples were cut into thin slabs oriented perpendicular to bedding and then pulverized. Sample splits were used to determine inorganic carbon by HCL digestion, organic carbon and total sulfur by LECO combustion techniques, and trace metals by an aqua regia digestion followed by inductively coupled plasma emission spectroscopy (ICP). In addition, 5 samples were chosen for whole-rock chemical analysis, which included fusion of the samples, total digestion by nitric acid, and ICP analysis.

Of the nineteen samples, two were collected in the lower unit and have relatively high carbonate contents (60 and 38 wt.% $CaCO_3$ equivalent). Carbonate contents in the samples from the upper unit range from 4 to 31%, with an average of 21%. Whole-rock geochemical analyses of the Athens Shale (Table 1) reflect the variability in the relative proportions of the major minerals (carbonates, clays, and quartz). Organic carbon and total sulfur range from 0.73% to 1.54% (mean=1.15%) and 0.77% to 3.84% (mean=1.45%), respectively. Molybdenum content of the Athens Shale ranges from 6-16 ppm, with a mean of 12 ppm. The average Mo content of the Athens Shale is six times higher than that of the average shale and slightly higher than the average black shale (Table 2). Co, Ni, V, and Ba are also slightly higher in the Athens Shale than the average black shale, but they are within the limits of uncertainty of the analytical and sampling procedures for the compilation of the average black shale of Vine and Tourtelot (1970). However, the Athens Shale contains significantly less Mo and other trace metals than metalliferous black shales (e.g., the Chattanooga Shale) as defined by Huyck and others (1990).

DISCUSSION

The hypothesis that MVT ore-forming solutions evolved from basin brines is based largely on the proximity of MVT deposits to sedimentary basins and the similarity between the composition of solutions trapped by fluid inclusions in MVT ore minerals and present-day metal-rich oil field brines (e.g., Hanor, 1979). Numerous mechanisms have been cited to explain the movement of these brines from hotter, deeper central portions of basins out to the basin margins where MVT deposits typically occur (Noble, 1963; Jackson and Beales, 1967; Cathles and Smith, 1983; Garven and Freeze, 1984; Bethke, 1986; and Deloule and Turcotte, 1989). Oliver (1986) has considered some of the aspects of fluid movement related to Appalachian orogenic events, and emphasizes the role that orogenesis plays on both sediment deposition and subsequent fluid migration during basin deformation. With regards to the southern Appalachians, recent studies (Kesler and others, 1988; Nakai and others, 1990,1993) indicate that the Middle Ordovician Sevier basin was the fluid and metal source for the MVT deposits of the East Tennessee district, and a pre-Alleghenian orogenic event deformed the basin and initiated the formation of ore-forming fluids. We extend this interpretation further to the southeast to northwest Georgia, and suggest that the basal unit in this basin, the Athens Shale, was the source of metals and possibly the fluids responsible for producing MVT mineralization there.

Because of their enrichments in certain trace metals, black shales have been proposed as possible sources for MVT ore-forming solutions and metals in the past (e.g., Hanor, 1979; Coveney, 1979; Long and Angino, 1982; Sverjensky, 1984). The Rockmart Slate crops out within a hundred meters of the Shiloh Church deposit, providing further support that it was a likely metal source. In contrast, the barite

Table 2. Comparison of Trace Metal Contents of Shales

Minor elements (ppm)	Athens Shale (n=19)*	Average Shale[1]	Average Black Shale[2]	Chattanooga Shale[3] (n=9)
Mo	12	2	10	45
Co	15	20	10	ND
As	14	10	ND	36
Cu	39	50	70	127
Zn	117	90	<300	238
Pb	22	20	20	ND
Ni	63	80	50	ND
Ba	530	600	300	ND
U	5	3	2	51
V	160	130	150	ND
Major elements (%)				
S	1.45	0.25	ND	3.7
$C_{org.}$	1.15	ND	3.2	12.2

*n=5 for Ba, V

Data sources: 1. Krauskopf (1979); 2. Vine and Tourtelot (1970); 3. Leventhal and others (1983). Abbreviation: ND= not determined.

deposits of the Cartersville district, which are hosted by Cambrian carbonates, are stratigraphically farther removed from the Athens/ Rockmart, but the Late Paleozoic Cartersville thrust fault (Figure 1) obscures geologic relationships to the east of the district.

That a mineral deposit would reflect the geochemical signature of its source rocks would appear to be highly fortuitous given that it would require: 1) a fluid to dissolve and transport all of the constituents in the source rock to the approximate same extent as their abundance in the rock; 2) that the metal-bearing fluids did not undergo appreciable chemical evolution during migration; and 3) a precipitation mechanism capable of depositing all of the metals. If all of these criteria have to be met, how is it possible that a MVT deposit enriched in Mo, Zn, Pb, Ni, Co, and As could form assuming that a black shale was the source of the metals? We suggest that perhaps the answer lies in the close proximity of the Shiloh Church deposit to its source rocks that are enriched in these elements. It is interesting to note that Mo, Co, and Ni occur in minor concentrations in MVT deposits of the Viburnum trend of southeast Missouri (Hagni, 1983), but we are not aware of any other MVT deposits where molybdenum is a principal ore constituent. The general lack of significant amounts of Mo in MVT deposits may be a testament to the ease with which it can be removed from ore-forming solutions along their migration paths.

Present day Na-Ca-Cl oil field brines from central Mississippi commonly contain high levels (>100 ppm) of zinc and lead (Carpenter and others, 1974; Kharaka and others, 1987) and have been assumed to be analogs of MVT ore-forming solutions (Sverjensky, 1984). However, these brines contain only a maximum of 0.05 ppm Mo (Saunders and Swann, 1990), suggesting that a solution of similar composition would not be capable of producing a Mo-rich MVT deposit. Geochemical modeling by Kharaka and others (1987) showed that high levels of Zn and Pb are present in low-sulfide Mississippi oil field brines due to metal complexing by chloride. In contrast, molybdenum does not form chloride complexes, and is typically present at low levels in hydrothermal solutions as the oxyacid H_2MoO_4 (Wood and others, 1987). Giordano (1990) has suggested that organic complexing might be important in raising the solubility of Mo (and Ni, Cu, and V) in sedimentary environments, but experimental data on these complexes are lacking. The proximity of the Shiloh Church deposit to the organic-rich Athens Shale suggests that Mo-organic complexes may have played a role in the genesis of the deposit.

CONCLUSIONS

The Athens Shale is the basal unit in an Ordovician basin that was deformed during the Middle Paleozoic Taconic and/or Acadian orogenies and the Late Paleozoic Alleghenian orogeny. Compressional deformation, probably during middle Paleozoic foreland thrusting and folding, triggered fluid movement out of the Athens Shale into the karstified unconformity at the top of the Knox Group (e.g., Figure 1). The Knox unconformity could have served as a regional aquifer for fluids moving out of the basin, as proposed by Hoagland (1976) for east Tennessee, and a local trap for precipitation of MVT deposits. Based on the close stratigraphic proximity of the Knox-hosted Shiloh Church molybdenum deposit to the Athens Shale (Rockmart Slate), which has typical black shale abundances of Mo, Co, Ni, V, Zn, Pb, and Ba, we propose that the Athens Shale was the source of these metals in this MVT molybdenum deposit. The lack of appreciable molybdenum in other MVT deposits may be the result of the lack of black shales in the source basins, early precipitation of molybdenum during brine transport, or both.

ACKNOWLEDGEMENTS

We would like to thank John Allan of North American Exploration for providing unpublished data on the Shiloh Church deposit and samples for study. Discussions with Tim Chowns and Mark Steltenpohl added greatly to our understanding of the stratigraphic and tectonic setting of the Athens Shale.

REFERENCES CITED

Aleinikoff, J.N., Walter, M., Kunk, M.J., Hearn, P.P., 1993, Do ages of authigenic K-feldspar date the formation of Mississippi Valley-type Pb-Zn deposits, central and southeastern United States?: Pb isotopic evidence: Geology, v. 21, p. 73-76.

Benson, D.J., 1986a, Stratigraphic setting of the Middle Ordovician of Alabama: *in* Benson, D.J., and Stock, C.W., eds., Depositional history of the Middle Ordovician of the Alabama Appalachians, Alabama Geological Society, 23rd Annual Field Trip Guidebook, p. 3-14.

Benson, D.J., 1986b, Depositional setting and history of the Middle Ordovician of the Alabama Appalachians: *in* Benson, D.J., and Stock, C.W., eds., Depositional history of the Middle Ordovician of the Alabama Appalachians, Alabama Geological Society, 23rd Annual Field Trip Guidebook, p. 15-31.

Benson, D.J., and Stock, C.W., 1986, Field Trip Stops: *in* Benson, D.J., and Stock, C.W., eds., Depositional history of the Middle Ordovician of the Alabama Appalachians, Alabama Geological Society, 23rd Annual Field Trip Guidebook, p. 97-119.

Bethke, C.M., 1986, Hydrologic constraints on the genesis of the Upper Mississippi Valley mineral district from Illinois basin brines: Economic

Geology, v. 81, p. 233-249.

Carpenter, A.B., Trout, M.L., and Pickett, E.E., 1974, Preliminary report on the origin and chemical evolution of lead- and zinc-rich oil field brines in central Mississippi: Economic Geology, v. 69, p. 1191-1206.

Carter, B.D., and Chowns, T.M., 1986, Stratigraphic and environmental relationships of Middle and Upper Ordovician rocks in northwestern Georgia and northeastern Alabama: *in* Benson, D.J., and Stock, C.W., eds., Depositional history of the Middle Ordovician of the Alabama Appalachians, Alabama Geological Society, 23rd Annual Field Trip Guidebook, p. 33-50.

Cathles, L.M., and Smith, A.T., 1983, Thermal constraints on the formation of Mississippi Valley-type lead-zinc deposits and their implications for episodic basin dewatering and deposit genesis: Economic Geology, v. 78, p. 983-1002.

Coveney, R.M., Jr., 1979, Sphalerite concentrations in mid-continent Pennsylvanian black shales of Missouri and Kansas: Economic Geology, v. 74, p. 131-140.

Coveney, R.M., Jr., Leventhal, J.S., Glascock, M.D., and Hatch, J.R., 1987, Origins of metals and organic matter in the Mecca Quarry shale member and stratigraphically equivalent beds across the Midwest: Economic Geology, v. 82, p. 915-933.

Coveney, R.M., Jr., and Glascock, M.D., 1989, A review of origins of metal-rich Pennsylvanian black shales, central U.S.A., with an inferred role for basinal brines: Applied Geochemistry, v. 2, p. 543-561.

Deloule, E., and Turcotte, D.L., 1989, The flow of hot brines in cracks and the formation of ore deposits: Economic Geology, v. 84, p. 2217-2225.

Foss, D.W., Gatten, O.J., and Young, R.S., 1983, Shiloh Church molybdenum deposit, Polk County, Georgia: Society of Mining Engineers of AIME preprint 83-86, 10p.

Garven, G., and Freeze, A., 1984, Theoretical analysis of the role of groundwater flow in the genesis of stratabound ore deposits 2: quantitative results: American Journal of Science, v. 284, p. 1125-1174.

Giordano, T.H., 1990, Organic ligands and metal-organic complexing in ore fluids of sedimentary origin, *in* Grauch, R.I., and Huyck, H.L.O., eds., Metalliferous black shales and related ore deposits--proceedings, 1989 United States working group meeting, International Geological Correlation Program Project 254: U.S. Geological Survey Circular 1058, p. 31-41.

Hagni, R.D., 1983, Minor elements in Mississippi Valley-type ore deposits, *in* Shanks, W.C., ed., Cameron volume on unconventional mineral deposits: New York, Society of Mining Engineers of AIME, p. 71-87.

Hanor, J.S., 1979, The sedimentary genesis of hydrothermal fluids, *in* Barnes, H.L., ed., Geochemistry of Hydrothermal Ore Deposits: New York, John Wiley & Sons, p. 137-172.

Haseltine, R.H., 1924, Iron ore deposits of Georgia: Georgia Geologic Survey Bulletin 41, 222p.

Hearn, P.P., and Sutter, J.F., 1985, Authigenic potassium feldspar in Cambrian carbonates: evidence of Alleghenian brine migration: Science, v. 228, p. 1529-1531.

Hearn, P.P., and Sutter, J.F., and Belkin, H.E., 1987, Evidence for Late-Paleozoic brine migration in Cambrian carbonate rocks in the central and southern Appalachians: implications for Mississippi Valley-type sulfide mineralization: Geochimica et Cosmochimica Acta, v. 51, p. 1323-1334.

Hoagland, A.D., 1976, Appalachian zinc-lead deposits, *in* Wolf, K.H., ed., Handbook of Stratabound and Stratiform Ore Deposits, Volume 6: Amsterdam, Elsevier, p. 495-534.

Huyck, H.L.O., 1990, When is a metalliferous black shale not a black shale?, *in* Grauch, R.I., and Huyck, H.L.O., eds., Metalliferous black shales and related ore deposits--proceedings, 1989 United States working group meeting, International Geological Correlation Program Project 254: U.S. Geological Survey Circular 1058, p. 42-56.

Jackson, S.A., and Beales, F.W., 1967, An aspect of sedimentary basin evolution: The concentration of Mississippi Valley-type ores during late stages of diagenesis: Canadian Petroleum Geology Bulletin, v. 15, p. 383-433.

Jones, M.L., and Dennison, J.M., 1970, Oriented fossils as paleocurrent indicators in Paleozoic lutites of southern Appalachians: Journal of Sedimentary Petrology, v. 40, p. 642-649.

Kesler, T.L., 1950, Geology and mineral resources of the Cartersville district, Georgia: U.S. Geological Survey Professional Paper 224, 97p.

Kesler, S.E., Jones, L.M., and Ruiz, J., 1988, Strontium isotopic geochemistry of Mississippi Valley-type deposits, East Tennessee: Implications for age and source of mineralizing brines: Geological Society of America Bulletin, v. 100, p.

1300-1307.

Kesler, S.E., and van der Pluijm, B.A., 1990, Timing of Mississippi Valley-type mineralization: Relation to Appalachian orogenic events: Geology, v. 18, p. 1115-1118.

Kharaka, Y.K., Maest, A.S., Carothers, W.W., Law, L.M., Lamothe, P.J., and Fries, T.L., 1987, Geochemistry of metal-rich brines from the central Mississippi Salt Dome basin, U.S.A.: Applied Geochemistry, v. 2, 534-561.

Krauskopf, K.B., 1979, Introduction to geochemistry: New York, McGraw-Hill Book Company, p. 544-545.

Leventhal, J.S., Briggs, P.H., and Baker, J.W., 1983, Geochemistry of the Chattanooga Shale, DeKalb County, Tennessee: Southeastern Geology, v. 24, p. 101-116.

Long, D.T., and Angino, E.E., 1982, The mobilization of selected trace metals from shales by aqueous solutions: Effects of temperature and ionic strength: Economic Geology, v. 77, p. 646-652.

Lu, G., Marshak, S., and Kent, D.V., 1990, Characteristics of magnetic carriers responsible for Late Paleozoic remagnetization in carbonate strata of the mid-continent, U.S.A.: Earth and Planetary Science Letters, v. 99, p. 351-361.

McCabe, C., and Elmore, R.D., 1989, The occurrence and origin of Late Paleozoic remagnetization in the sedimentary rocks of North America: Reviews in Geophysics, v. 27, p. 471-494.

Nakai, S., Halliday, A.N., Kesler, S.E., and Jones, H.D., 1990, Rb-Sr dating of sphalerites from Tennessee and the genesis of Mississippi Valley type ore deposits: Nature, v. 346, p. 354-357.

Nakai, S., Halliday, A.N., Kesler, S.E., and Jones, H.D., Kyle, J.R., and Lane, T.E., 1993, Rb-Sr dating of sphalerites from Mississippi Valley-type (MVT) ore deposits: Geochimica et Cosmochimica Acta, v. 57, p. 417-427.

Noble, E.A., 1963, Formation of ore deposits by water of compaction: Economic Geology, v. 58, p. 1145-1156.

Oliver, J., 1986, Fluids expelled tectonically from orogenic belts: Their role in hydrocarbon migration and other geologic phenomena: Geology, v. 14, p. 99-102.

Read, J.F., 1980, Carbonate ramp-to-basin transition and foreland basin evolution, Middle Ordovician, Virginia Appalachians: American Association of Petroleum Geologists Bulletin, v. 64, p. 1575-1612.

Reade, E.H., Power, W.R., Keeners, M.J., and White, D.H., 1980, Barite deposits in the Cartersville district, Bartow County, Georgia: Excursions in Southeast Geology (GSA-1980) Field trip guidebook, trip 19, p. 379-384.

Rife, D.L., 1971, Barite fluid inclusion geothermometry, Cartersville mining district, Georgia: Economic Geology, v. 66, p. 1164-1167.

Roedder, E., 1984, Fluid inclusions: Mineralogical Society of America Reviews in Mineralogy, v. 12, 644 p.

Rodgers, J., 1970, The Tectonics of the Appalachians: New York, John-Wiley and Sons, 271 p.

Saunders, J.A., and Swann, C.T., 1990, Trace-metal content of Mississippi oil field brines: Journal of Geochemical Exploration, v. 37, p. 171-183.

Schedl, A., McCabe, C., Montanez, I.P., Fullagar, P.D., Valley, J.W., 1992, Alleghenian regional diagenesis: a response to the migration of modified metamorphic fluids derived from beneath the Blue Ridge-Piedmont thrust sheet: Journal of Geology, v. 100, p. 339-352.

Shanmugam, G., and Lash, G.G., 1982, Analogous tectonic evolution of the Ordovician foredeeps, southern and central Appalachians: Geology, v. 10, p. 562-566.

Shanmugam, G., and Walker, K.R., 1980, Sedimentation, subsidence, and evolution of a foredeep basin in the Middle Ordovician, southern Appalachians: American Journal of Science, v. 280, p. 479-496.

Sverjensky, D.A., 1984, Oil field brines as ore-forming solutions: Economic Geology, v. 79, p. 23-37.

Tieman, D.J., and Kopp, O.C., 1983, Deformation fabrics in barite and fluorite of the Del Rio district: Their implications for the relative timing of mineralization and deformation: Economic Geology, v. 78, p. 91-104.

Vine, J.D., and Tourtelot, E.B., 1970, Geochemistry of black shales--a summary report: Economic Geology, v. 65, p. 253-272.

Walker, K.R., Shanmugam, G., and Ruppel, S.C., 1983, A model for carbonate to terrigenous clastic sequences: Geological Society of America Bulletin, v. 94, p. 700-712.

Wood, S.A., Crerar, D. A., and Borcsik, M.P., 1987, solubility of the assemblage pyrite, pyrrhotite, magnetite, sphalerite, galena, gold, stibnite, bismuthinite, argentite, molybdenite in H_2O-NaCl-CO_2 solutions from 200° to 350°C: Economic Geology, v. 82, p. 1864-1887.

SOUTHEASTERN GEOLOGY
V. 33, No. 4, October 1993, p.171-193

REVISED AGE AND CORRELATION OF THE UPPER TRIASSIC CHATHAM GROUP (DEEP RIVER BASIN, NEWARK SUPERGROUP), NORTH CAROLINA

PHILLIP HUBER
Department of Geological Sciences
Ohio University
Athens, OH 45701-2979

SPENCER G. LUCAS
New Mexico Museum of Natural History
1801 Mountain Rd.
Albuquerque, NM 87104

ADRIAN P. HUNT
Department of Geology
University of Colorado at Denver
Denver, CO 80217-3364

ABSTRACT

The Chatham Group of east-central North Carolina is exposed in a series of connected synrift basins collectively termed the Deep River basin. While a considerable flora and fauna clearly establish a Late Triassic age for these strata, age assignments at the stage level have ranged from early Carnian to Rhaetian. Recent studies combining palynomorphs, megafossil plants, fossil fishes, tetrapods and vertebrate ichnotaxa have reportedly developed a refined biostratigraphy for the Chatham Group, and recognize an early Carnian through early Norian? age for these strata. We are able to correlate tetrapods from the Chatham Group with similar faunas from the lower Chinle Group of the western United States and the Gettysburg-Newark basin of Pennsylvania and New Jersey. The oldest Chatham Group tetrapod fauna includes the aetosaur *Longosuchus meadei* and the dicynodont *Placerias hesternus*, which co-occur in the basal Chinle Group. This fauna is from the middle Pekin Formation and belongs to the *Paleorhinus* Biochron of late Carnian (early Tuvalian) age. The phytosaur *Rutiodon* is common to the Cumnock Formation, formations of the lower Chinle Group, and the middle New Oxford, upper Stockton and possibly Lockatong Formations of the Gettysburg-Newark basin. Other taxa that occur in this interval include large metoposaurids (*Buettneria*), the aetosaur *Desmatosuchus*, and a diversity of ornithischian dinosaurs, all of which occur in two or more of the regions under consideration. These strata belong to the *Rutiodon* Biochron and are of Late Carnian (late Tuvalian) age. The aetosaur *Stegomus* has been found in the unnamed Sanford Formation equivalent in the Durham sub-basin, the lower Passaic Formation in the Newark Basin and the middle New Haven Arkose of the Hartford basin. These formations are in part, correlatives and are assigned to the *Stegomus* Faunachron. This review of Chatham Group tetrapods permits geographic extension of the Chinle Group provincial tetrapod biochronology to basins of the Newark Supergroup and facilitates direct stratigraphic correlations between these regions. The revised age determinations further suggest that palynostratigraphic correlations of the Newark Supergroup basins are not as accurate as previously believed.

INTRODUCTION

Strata of the Deep River basin complex in North Carolina comprise the Chatham Group and belong to the early Mesozoic Newark

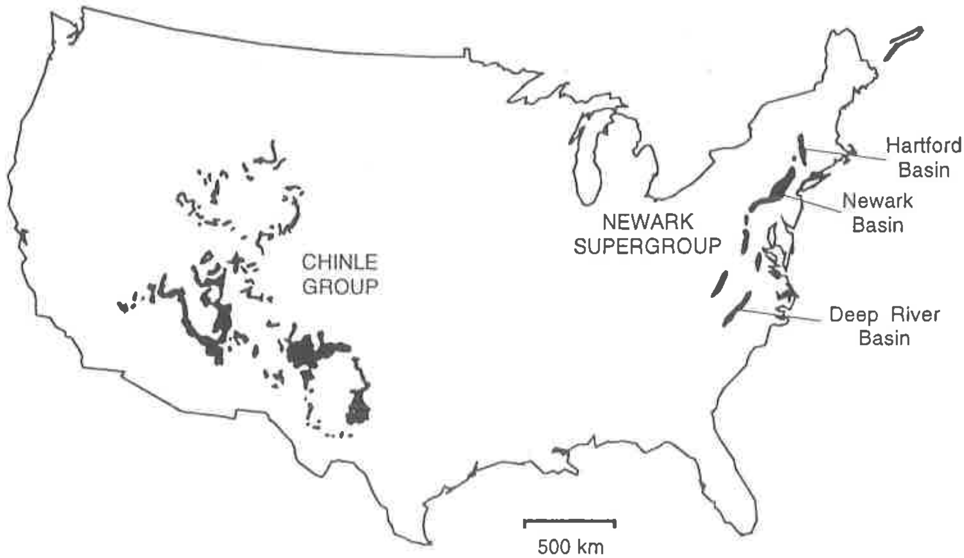


Figure 1. Location of Upper Triassic Newark Supergroup and Chinle Group outcrop areas discussed in text.

Supergroup (Froelich and Olsen, 1984), which fills an extensive series of exposed basins extending from South Carolina to Nova Scotia (Figure 1). Newark Supergroup basins formed in response to crustal extension and initial rifting of the supercontinent Pangaea during the Middle and Upper Triassic and Early Jurassic (Manspeizer, 1988). The Deep River basin complex includes three connected sub-basins which are, from north to south, the Durham, Sanford and Wadesboro sub-basins. Strata in the sub-basins dip gently eastward toward the Jonesboro fault zone, and the generalized basin structure is that of a half graben. Throughout the basin complex, Chatham Group sediments are intruded by numerous diabase dikes and sills that correspond in age to similar tholeiitic intrusives and basalts found in Newark Supergroup basins to the north.

Of the various Triassic formations exposed in the southeastern United States, only those of the Chatham Group have produced a diverse fauna containing age-diagnostic tetrapods. Some horizons in other basins also contain terrestrial vertebrate faunas (e.g. Sues and Olsen, 1991; Huber and others, in press a), but these share few taxa with contemporaneous assemblages of other regions and are not very useful for correlation. A review of Chatham Group paleontology and fossil distribution indicates that much of the basin fill sequence has been misdated. Previous age determinations purporting an early and middle Carnian age, based on palynostratigraphy (Cornet, 1977; Cornet and Olsen, 1985; Robbins, 1985; Traverse, 1986; 1987; Robbins and others, 1988), are not consistent with age assignments suggested by the vertebrate faunas. The distribution of osseous tetrapod fossils indicates that the middle and upper Pekin Formation are late Carnian (early Tuvalian), the Cumnock Formation is also late Carnian (late Tuvalian), and that the Sanford Formation and its' lithostratigraphic correlatives are early Norian. The revised age assignments are based on correlation with the lower Chinle Group of the western United States and the Gettysburg-Newark basin of Pennsylvania and New Jersey, as these stratigraphic sequences possess a similar succession of vertebrate faunas.

Abbreviations

The following abbreviations are used in the text: AMNH, American Museum of Natural History, New York; MCZ, Museum of Comparative Zoology, Cambridge; NCSU, North Carolina State University, Raleigh;

USNM, United States National Museum, Washington, D. C.; YPM, Yale Peabody Museum, New Haven.

Previous Studies

The work of Emmons (1856) first established a Late Triassic age for Deep River basin strata, though he also regarded the "lower red sandstone and conglomerate [Pekin Formation]" to include rocks of Permian age. Redfield (1856) determined (incorrectly) that specimens of fossil fishes collected by Emmons compared favorably with those from the Hartford basin and assigned all Deep River basin strata to the Triassic Newark Group. Ward (1900) and Woodworth (1902) supported this age on the basis of the megaflora, while Fontaine (1883) suggested the flora resembled Rhaetian assemblages from western Europe, which he believed to belong to basal Lias. Colbert and Gregory (1957) attempted to correlate nonmarine Triassic strata throughout North America by comparing their vertebrate faunas and assigned the Chatham Group to the Late Triassic. Colbert (1965) reviewed phytosaur-based correlations of Late Triassic formations in the United States, and concluded that Newark Supergroup formations that contained *Rutiodon* (Cumnock, New Oxford, upper Stockton and Lockatong Formations) were broadly homotaxial and of early Late Triassic (Carnian) age.

Recent biostratigraphic studies of the flora and fauna began with Baird and Patterson (1967) and Patterson (1969), who noted that a vertebrate assemblage from the middle Pekin Formation compared favorably with the lower Chinle-Dockum faunas. Preliminary studies of palynomorphs (Schultz and Hope, 1973; Dunay and Fisher, 1974) and re-evaluation of the megaflora (Delevoryas and Hope, 1971; 1973; Hope and Patterson, 1969; 1970) strengthened this generalized correlation. Cornet (1977) undertook the first comprehensive palynostratigraphic analysis of Newark Supergroup synrift basins and concluded that most strata within the various North Carolina basins were middle and late Carnian (Julian-Tuvalian). Further studies by Robbins (1985), Traverse (1986; 1987) and Robbins and others (1988) expanded Cornets' interbasinal correlations to include strata of early Carnian (Cordevolian) age. Other biostratigraphic data were provided by fossil fishes (Schaeffer and McDonald, 1978; Olsen and others, 1982), and based on all available evidence, Smoot and others (1988), Luttrell (1989) and Olsen and others (1989) correlated the Deep River basin with other basins of the Newark Supergroup. Cornet and Olsen (1985) and Olsen and Sues (1986) extended these correlations to the Western Interior and recognized a middle and late Carnian age for the Chinle-Dockum (Chinle Group of Lucas, in press). Hunt and Lucas (1991a) revised the taxonomy of Pekin Formation aetosaurs, which include taxa common to late Carnian formations of the Chinle Group. Huber and others (1991; in press a; b) reviewed Newark Supergroup biochronology and argued that on the basis of shared tetrapod distribution between the Newark Supergroup, Chinle Group and Germanic Keuper, most "middle" Carnian formations of the Newark should be assigned to the late Carnian.

Stratigraphic Summary

Emmons (1856) defined the basic stratigraphy of the Deep River basin to consist of lower and upper coarse clastics which are separated by a medial interval of coal-bearing shale. He coined the term Chatham Group for these strata. Later, Campbell and Kimball (1923) mapped the basin complex and formally named the major lithostratigraphic units (ascending) Pekin, Cumnock and Sanford Formations. Reinemund (1955) provided a detailed study of the Sanford sub-basin which focused on coal resources of the Cumnock Formation. Bain and Harvey (1977), Parker (1979) and Hoffman and Gallegher (1989) developed a detailed stratigraphy for the Durham sub-basin, but defined the basin fill sequence in terms of mappable facies associations without applying formal lithostratigraphic nomenclature. Olsen and others (1990) have recently summarized the stratigraphy and structure of the Chatham Group. These authors follow earlier workers by defining the Deep River basin to possess a large-scale, tripartite stratigraphy beginning with basal, coarse-grained fluvial facies overlain by medial alluvial plain and lacustrine sandstone and mudrock, followed by an upper facies association of alluvial plain clastics, minor carbonates, and sub-adjacent to the basin margins, alluvial fan conglomerates. Lithostratigraphy and age relationships of strata in the three sub-basins are shown in Figure 2.

The lower strata of the sub-basins consists of thick (500+ m) redbed sequences dominated by pebble conglomerate, sandstone, and siltstone. These sediments were deposited by alluvial fan and fluvial systems building off the faulted basin margins. Basal strata in the Wadesboro and Sanford sub-basins are called Pekin Formation (Campbell and Kimball, 1923) and informally termed "Lithofacies Association I" in the Durham sub-basin. In the Sanford sub-basin, overlying strata are dominated by organic-rich siltstone, sandstone and coal beds of the Cumnock Formation that represent deposition in laterally extensive lakes (Reinemund, 1955; Olsen and others, 1989). Homotaxial strata in the Wadesboro sub-basin include cyclic sandstone-siltstone sequences that represent deposition in shallow lacustrine and playa environments. These strata form part of Campell and Kimballs' type section of the Pekin Formation (cf. Olsen and others, 1990). Shallow lacustrine and playa facies interbedded with fluvial sediments comprise Hoffman and Gallegher's "Lithofacies Association II" in the Durham sub-basin (Olsen, 1977; Olsen and others, 1989). The upper strata in the three sub-basins are poorly exposed, but appears to be composed of fluvial redbeds and adjacent to the Jonesboro fault zone, alluvial fan conglomerate (Reinemund, 1955; Olsen and others, 1989; 1990). These strata in the Wadesboro sub-basin have not been formally named, are called Sanford Formation in the Sanford sub-basin and informally-termed "Lithofacies Association III" in the Durham sub-basin.

Data from palynostratigraphy (Cornet, 1977), fossil fish distribution (Olsen and others, 1982) and lithostratigraphy (Reinemund, 1955; Bain and Harvey, 1977; Olsen and others, 1990), indicates that the fossiliferous, medial lacustrine intervals of the three sub-basins were not the product of synchronous deposition. In the most recent synthesis of Deep River basin geology, Olsen and others (1990, fig. 9-5) concluded that deposition of "Lithofacies Association II" began after Cumnock deposition had ceased, and these authors recognize homotaxis of the former unit with the lower Sanford Formation. This conclusion is supported by tetrapod biochronology of the medial lacustrine intervals of the three sub-basins, which indicates correlation of the Cumnock Formation and the type section of the Pekin Formation. The tetrapods further indicate that the middle Pekin Formation is late Carnian (early Tuvalian), the Cumnock and type Pekin Formations are late Carnian (late Tuvalian), and that the Sanford Formation and "Lithofacies Association II" are early Norian (Figure 2).

FOSSIL DISTRIBUTION

Koob (1961) first described Chatham Group palynomorphs (Cumnock Formation), though this work was not published. Schultz and Hope (1973) recovered a diverse palynoflora from the Pekin Formation, but did not conclude an age more refined than Late Triassic for their samples (middle Pekin Formation, Sanford sub-basin). Cornet (1977) described well-preserved pollen from several horizons of the Pekin Formation in the Wadesboro and Sanford sub-basins and the Cumnock Formation. He concluded that most Deep River basin strata were of "middle" Carnian age. Robbins (1985), Traverse (1986) and Robbins and others (1988) described or illustrated additional palynomorphs from the Chatham Group and considered the Pekin Formation to be early and "middle" Carnian, and the Cumnock Forma-

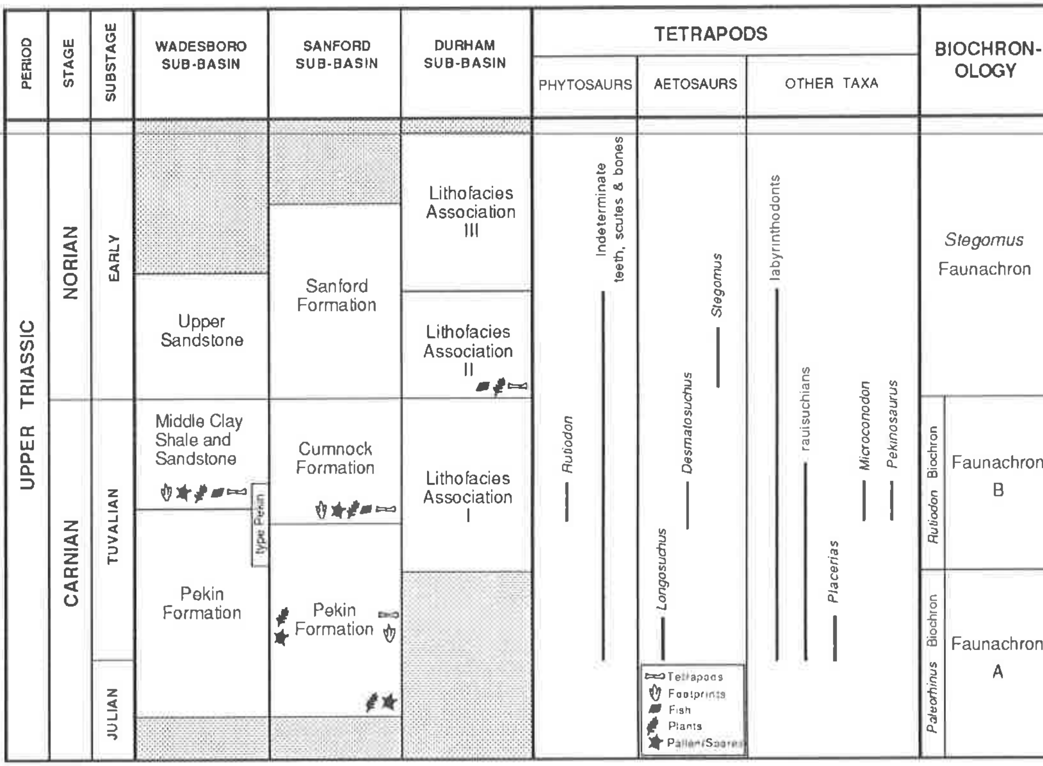


Figure 2. Lithostratigraphy, fossil distribution and age-relationships of the Chatham Group.

tion as "late middle" Carnian. These age estimates have been often cited (Gore, 1986; Olsen and others, 1989; Textoris and Gore, in press) and form in part, the basis for recently advocated intersupergroup correlations of the Newark synrift basins (Smoot and others, 1988; Luttrell, 1989).

Megafossil plants are particularly abundant in the middle Pekin and Cumnock Formations and have been described by Emmons (1856), Fontaine (1883), Ward (1900), Hope and Patterson (1969; 1970), Delevoryas and Hope (1971; 1973) and Gensel (1986). The flora is dominated by cycadophytes (*Zamites*) and equisetalians (*Neocalamites*), with less abundant, though numerous species of pteridophytes and conifers (Gensel, 1986). Ash (1980) placed the Pekin flora within his "Zone of *Eogingkgoites*", which also includes basal Chinle Group formations, and suggested that the two floras are contemporaneous and of "middle" Carnian age. Litwin and others (1991) later revised the age of this floral zone to early late Carnian (early Tuvalian) and placed the overlying "Zone of *Dinophyton*", represented in the Cumnock Formation, in the Late Carnian. These age determinations are supported by the tetrapod-based correlations discussed below.

Several comprehensive faunal lists of Deep River basin vertebrates have been published (Olsen, 1988; Olsen and others, 1989; 1990), but many of the specimens have not been described or illustrated. Important early discoveries were described by Emmons (1856; 1857; 1860), but most of the specimens were either lost or destroyed during the Civil War. Existing material from Emmon's collection is mostly from the Cumnock Formation and includes lower jaws of the therapsids *Microconodon* and *Dromotherium* (Emmons, 1857), a partial skull of the phytosaur *Rutiodon carolinensis* (Emmons, 1860; Lucas and Hunt, 1989), the skull roof of the temnospondyl *Dictyocephalus elegans* (Leidy, 1856), and teeth of the archosaur "*Zamatus*", which Cope (1871) thought belonged to a theropod dinosaur. Other Cumnock vertebrates were collected around the turn of the century and include the several, near-complete skulls of *Rutiodon* described by McGregor (1906) and Colbert (1947). Recent additions to the Deep River Basin fauna mentioned by Schaeffer and McDonald (1978) and Olsen and others (1982; 1989; 1990) include fossil fishes (cf. *Turseodus*, cf. *Cionichthys*, *Synorichthys*, *Semionotus*, cf. *Diplurus* aff. *newarki*) from several lithostratigraphic units of the basin complex, "?rauisuchian" teeth and the ichnogenera *Brachychirotherium*, *Apatopus* and *Coelurosaurichnus* from the middle Pekin Formation, "*Rutiodon*" from the Cumnock Formation, "*Metoposaurus*" and "*Rutiodon*" from "Lithofacies Association II", footprints similar to *Gwyneddichnium*, teeth and dermal scutes referred to "*Rutiodon*", and ornithischian dinosaur teeth from outcrops of the type Pekin Formation.

While this collective faunal list appears extensive, a considerable number of the tetrapod specimens are generically indeterminate. Among these is the type and only specimen of *Dictyocephalus elegans* Leidy 1856 (AMNH 5661), represented by a fragmentary skull roof which incurred substantial pre-burial and post-collection damage. Previous workers (Colbert and Imbrie, 1956; Gregory, 1980) have recognized the specimen as indeterminate, but Davidow-Henry (1989) considered *Dictyocephalus* to be a senior synonym of the genus *Anachisma* Branson 1905 (Popo Agie Formation, Wyoming) because the collective specimen sample lacks a well-defined otic notch and tabular horns. Davidow-Henry (1989) further applied the name *Dictyocephalus* to other specimens of small metoposaurids from the Petrified Forest Formation (Arizona), Redonda Formation (New Mexico) and Bull Canyon Member ("Cooper Member"; Chatterjee, 1986; Small, 1989) of the Dockum Formation (West Texas). The most recent revision of the Metoposauridae (Hunt, in press) follows earlier workers by restricting the name *Dictyocephalus* to the holotype and considers the taxon to be a nomen dubium because the specimen does not possess characters that are even diagnostic for the family Metoposauridae.

Other metoposaurid material called "*Metoposaurus*" by Olsen and others (1990) was collected from the Wadesboro and Durham sub-basins, but consists of isolated centra and skull armor fragments (P. Olsen, pers. comm., 1992), which are not diagnostic at the level of family. Gore (1986, p. 62) stated that both "*Eupelor elegans*" and *Dictyocephalus* were collected from the Cumnock Formation, but the only labyrinthodont material from this unit is the holotype of *Dictyocephalus* earlier mentioned. Other oft-cited Chatham group fossils that are indeterminate are isolated teeth of *Zamotus* Cope 1871 (Rauisuchia indet.) and all isolated teeth, skull fragments and postcrania of phytosaurs called *Rutiodon* (e.g. Gore, 1986; Olsen, 1988). Additional material was identified as "*Zamotus*" by Patterson (1969, Figs. 30, 31), but all the specimens in his Figure 31 actually pertain to the dicynodont *Placerias hesternus*. Furthermore, *Zamotus* is a form genus for laterally compressed and serrated teeth which lack recurvature. This tooth morphology is usually ascribed to rauisuchians, although it is also present in less-derived archosaurs. It is not clear which specimens in Patterson's (1969) Figure 30 are referred to "*Zamotus*", but none of the illustrated elements can be identified as Rauisuchia. Finally, it would be inappropriate to apply the name *Zamotus* to non-dentition cranial and postcranial elements because this taxon as originally defined by Cope (1871) is demonstrably a form genus for teeth.

Osseous tetrapod fossils which may be identified to genus pertain to the dicynodont *Placerias hesternus*, the aetosaurs *Longosuchus meadei*, *Desmatosuchus* cf. *haploceras* and *Stegomus* sp., the ornithischian dinosaur *Pekinosaurus olseni* and the advanced cynodont therapsid *Microconodon*. The articulated partial skull of *Rutiodon* (USNM 214513) figured by Emmons (1860) and additional skulls (e.g. AMNH 1) described by McGregor (1906) and Colbert (1947), all from the Cumnock Formation, are the only phytosaur specimens from

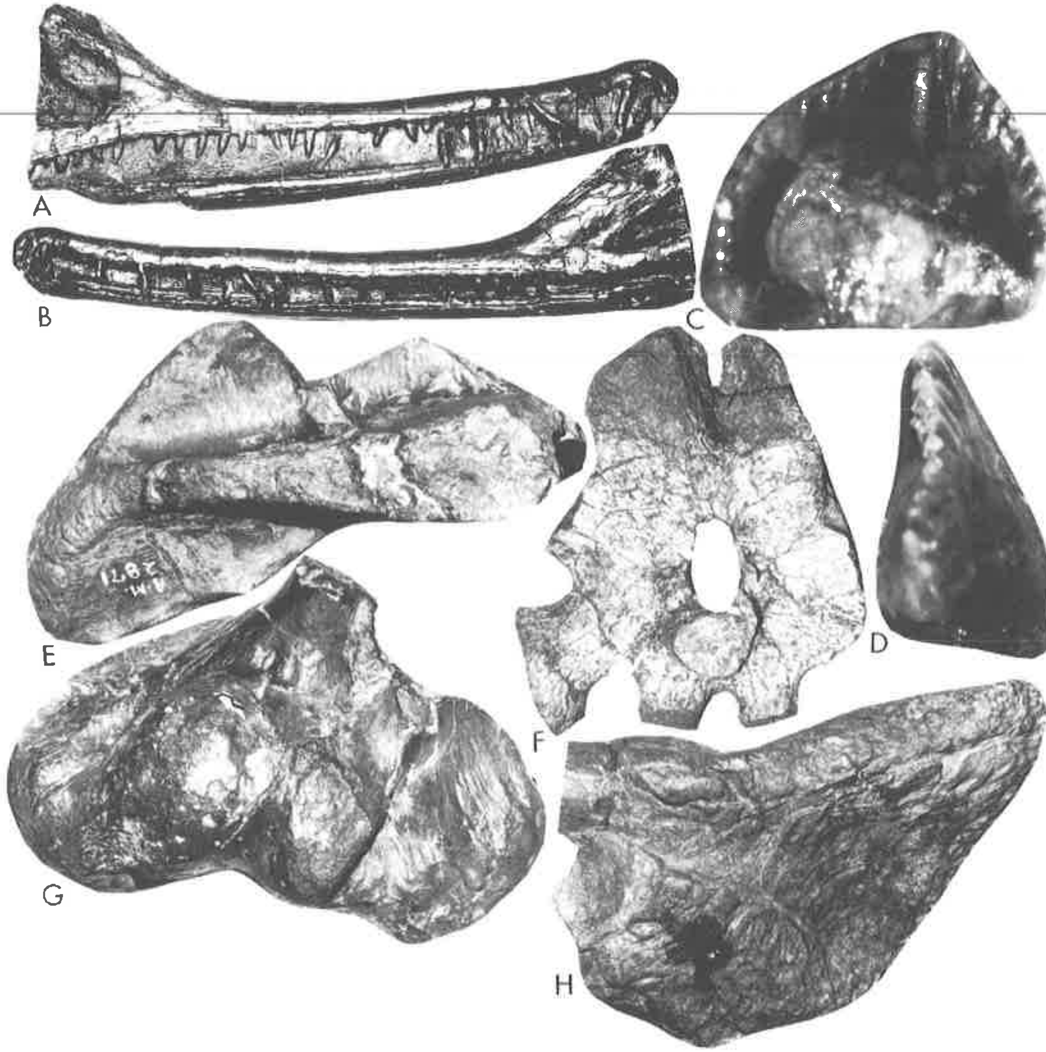


Figure 3. Chatham Group tetrapods. A-B, *Rutiodon carolinensis*, USNM 214513 (neotype), lateral views of near-complete shull. C-D, *Pekinosaurus olseni*, YPM 8545 (holotype), teeth in lingual view (C) and mesial view (D) ×7.48. E-H, Placerias hesternus. (E), AMNH 2871, incomplete squamosal, ×0.38, (F), AMNH 2885, portion of occupit, ×0.22, (G), AMNH 2880, distal end of right humerus (cast), ×0.38, (H), AMNH 2881, left maxilla, ×0.38.

the Deep River basin which may be confidently identified as *Rutiodon carolinensis* (Figure 3 A, B). The majority of phytosaur occurrences listed by Olsen (1988) and Olsen and others (1989) from not only the Deep River basin, but also the entire Newark Supergroup, cannot be identified as *Rutiodon* and we regard them as Phytosauria Indet. This comment is made because isolated teeth, jaw fragments and dermal scutes of phytosaurs are not identifiable to genus, and because *Rutiodon* has a short temporal range. Other Newark Supergroup phytosaur specimens that belong to *Rutiodon* are two near-complete skulls from the middle New Oxford Formation of the Gettysburg basin (Baird, 1986), the type specimen of *Rutiodon manhattensis* von Huene 1913 (upper Stockton Formation, Newark basin) and possibly the

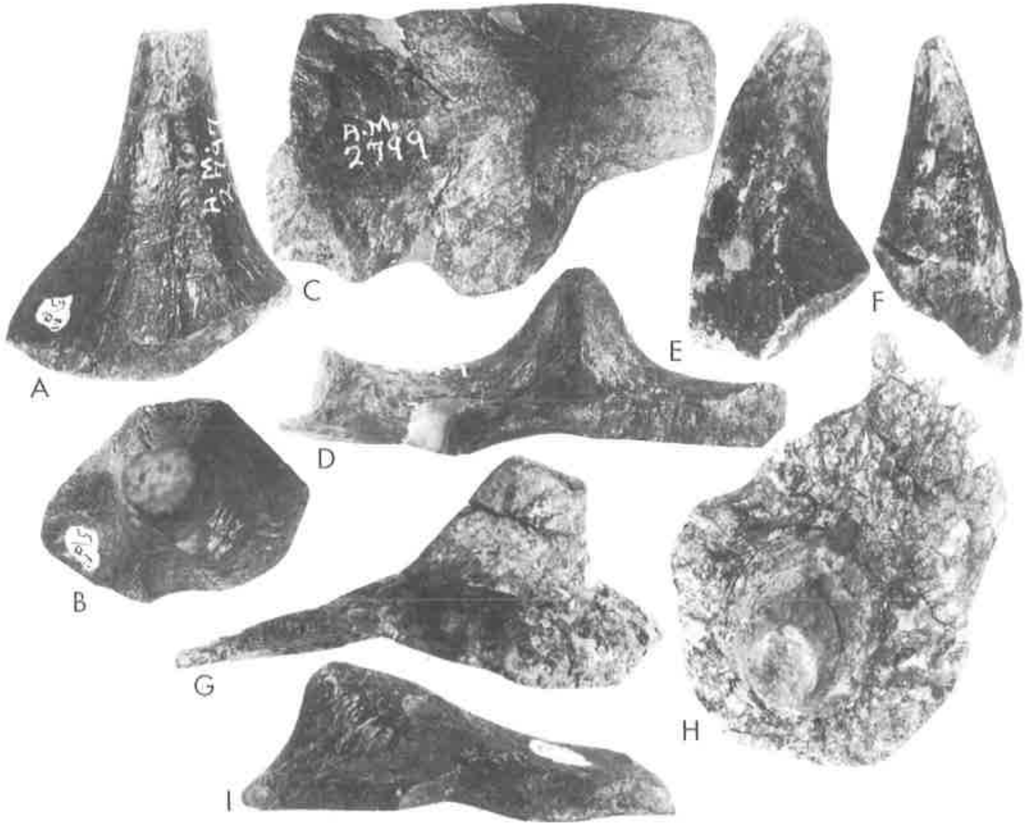


Figure 4. Pekin Formation aetosaurs. A-B, *Longosuchus meadei*, AMNH 2797, lateral spike in lateral view (A) and distal view (B), ×0.68. C-D, *Longosuchus meadei*, AMNH 2799, paramedian scute in dorsal view (C) and anterior view (D), ×0.68. E-F, *Longosuchus meadei*, MCZ 2751, lateral spikes in lateral view, ×0.68. G-H, *Desmatosuchus* sp., AMNH 24141, lateral spike in ventral view (G) and lateral view (H), ×0.82. I, *Longsuchus meadei*, AMNH 2796, paramedian scute in lateral view, × 1.02.

fragmentary skull (AMNH 5500) described by Colbert (1965) from the Lockatong Formation (Newark basin).

Aetosaurs have a wide distribution in Upper Triassic nonmarine strata across Pangaea, and like phytosaurs, most genera have a restricted stratigraphic range. Dermal armor from the Deep River basin had previously been identified as *Desmatosuchus* (type Pekin Formation exposures), "*Typothorax*" (middle Pekin Formation) and *Stegomus* ("Lithofacies Association II") by Baird and Patterson (1967) and Parker (1966). Hunt and Lucas (1991a) revised most of these specimens and confirmed the presence of *Desmatosuchus* in the Wadesboro sub-basin (Figure 4 G, H), while referring the "Typothorax" specimens to a new genus, *Longosuchus meadei* (Figure 4 A-F, I), which is also known from formations of the basal Chinle Group. The *Stegomus* specimen described by Parker (1966) consists of articulated dermal scutes that comprise a portion of the proximal tail (Figure 5). The scutes lack the well-defined, pitted sculpture characteristic of most other aetosaur genera (e.g. *Aetosaurus*, *Stagonolepis*) and instead are relatively smooth and possess faint, radial striae. This scute morphology has been consistently referred to *Stegomus* (Marsh, 1896; Jepsen, 1948; Baird, 1986). However, at least two small, undescribed aetosaurs from the early Norian Bull Canyon Formation of the Chinle Group have

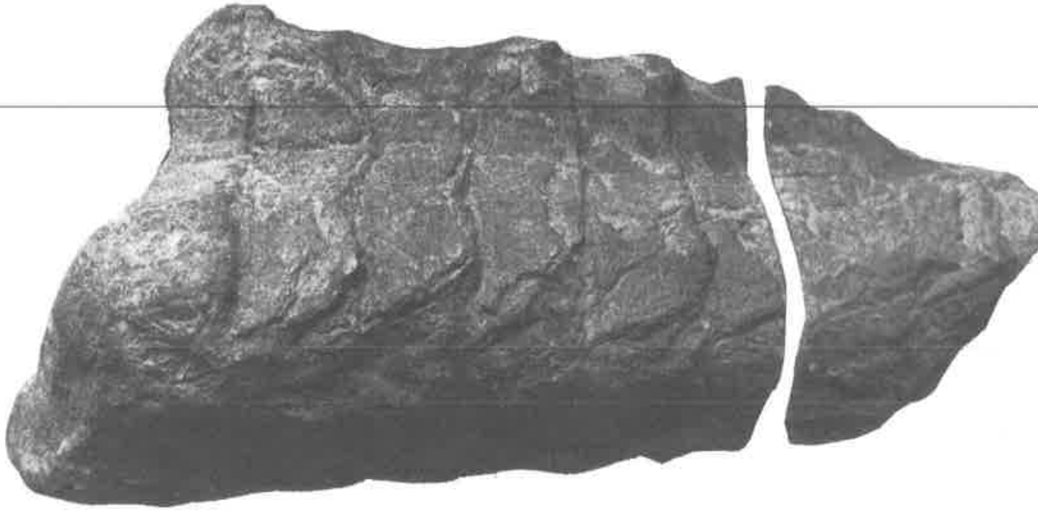


Figure 5. *Stegomus* cf. *arcuatus*, NCSU uncatalogued, natural cast of portion of proximal tail. The space between the two matrix blocks represents cuts made by a rock saw. Photograph courtesy of P.E. Olsen.

scutes which possess this morphology (Hunt, 1992), but differ from *Stegomus* in the overall proportions of the paramedian scutes and the trace of the sinuous suture between the paramedian and lateral scutes. Pending revision of the genus, we consider *Stegomus* to be a valid taxon, and at least the specimens described by Jepsen (1948), Parker (1966) and Baird (1986) to be congeneric.

Other Chatham Group tetrapods of biochronologic significance are fragmentary skulls and postcrania of the dicynodont *Placerias hesternus* from the middle Pekin Formation (Sanford sub-basin) and ornthischian dinosaur teeth from the type Pekin Formation (Wadesboro sub-basin). *Placerias* (Figure 3 E-H) was originally described from the lower Chinle Group, primarily from a large concentration of individuals in the Blue Mesa Member of the Petrified Forest Formation near St. Johns, Arizona (Camp and Welles, 1956; Murray and Long, 1989). Other dicynodont material from the Chinle Group was described by Williston (1904) and has been recently collected from the Santa Rosa Formation in eastern New Mexico (Lucas and Hunt, in review). At least some of Williston's (1904) specimens (Popo Agie Formation, Wyoming) pertain to *Placerias* and it is significant to note that all occurrences of this taxon are in strata dated as late Carnian. Ornithischian dinosaur teeth (Figure 3 C, D) from exposures of the type Pekin Formation have been described as a new taxon, *Pekinosaurus olseni* (Hunt and Lucas, in press). Their occurrence supports an age no older than late Carnian for the type Pekin because ornithischian dinosaurs from other realms of the Pangaean Triassic are found in strata considered to be no older than early Tuvalian (Lucas and others, 1992).

Tetrapod footprints occur in the Sanford sub-basin (middle Pekin and Cumnock formations) and in the type Pekin Formation in the Wadesboro sub-basin (Olsen and others, 1989; 1990). Represented ichnotaxa include *Apatopus*, cf. *Coelurosaurichnus* (middle Pekin Formation), *Brachychirotherium* (middle Pekin and Cumnock Formations) and *Gwyneddichnium* (type Pekin). Olsen and others (1989) considered the collective assemblage to represent the oldest known tetrapod ichnofauna from the Newark Supergroup, which in part is correct. However, the revised correlation of Newark Supergroup basins based on the taxonomy and distribution of osseous tetrapod remains (Huber and others, 1991; in press a, b)

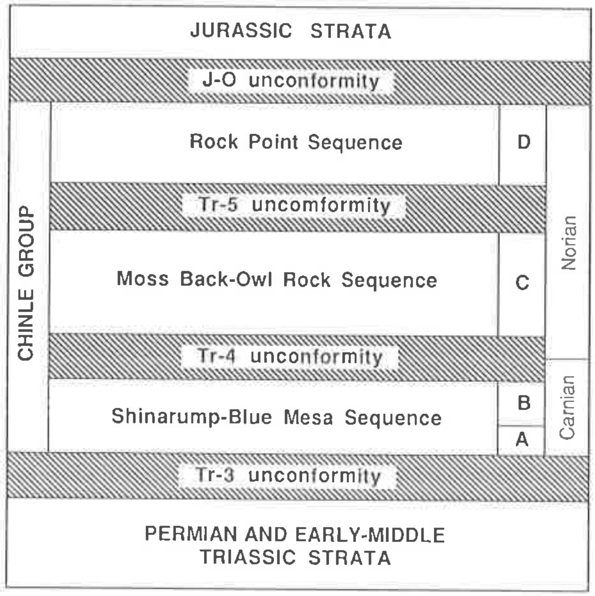


Figure 6. Major genetic lithostratigraphic sequences of the Upper Triassic Chinle Group (after Lucas, in press).

indicate that the combined assemblage from the type Pekin and Cumnock is contemporaneous with the more diverse footprint faunas from the upper Stockton and Lockatong Formations in the Newark basin (Olsen and Flynn, 1988; Olsen and others, 1989). Also note that the ichnotaxa *Apatopus* and *Coelurosaurichnus* have been identified from undoubted Norian formations of the Chinle Group (Conrad and others, 1987), and that *Coelurosaurichnus* is documented from Rhaetian formations of western Europe (Haubold, 1986). Likewise, the range of *Apatopus* extends very high in the Passaic Formation (Newark basin), to a horizon 10 m below the Orange Mountain basalt that approximates the Triassic-Jurassic boundary as defined by palynostratigrphy (Olsen and Baird, 1986). Thus, although the Chatham Group footprint fauna appears to be superficially distinct from other Newark Supergroup Triassic assemblages, these differences are not due to relative stratigraphic age.

Correlation

Vertebrate assemblages from the Chatham Group contain tetrapod taxa that have a shared distribution in other Newark Supergroup basins and formations of the Chinle Group in the southwestern United States. The Chinle Group is defined by three unconformity-bound stratigraphic sequences, and each sequence is comprised of a distinct association of depositional facies that can be traced over an area of 2.3 million square kilometers (Figures 1, 6). Each of the Chinle Group sequences contain distinctive fossil assemblages (Figure 7) and the age of these strata are constrained by sequence stratigraphic correlation with ammonite-dated, marine formations in western Nevada (Lucas, 1991), magnetostratigraphy (Bazard and Butler, 1990; Molina-Garza and others, 1991; Lucas and others, in press), a well-defined succession of Late Triassic provincial tetrapod faunas (Ballew, 1989; Lucas, 1990; in press, Hunt and Lucas, 1991 a, c, d; Hunt, in press), the cosmopolitan distribution of the phytosaur taxon *Paleorhinus* in late Carnian (early Tuvalian) formations across Pangaea (Hunt and Lucas, 1991 b), and of lesser biochronologic value, the stratigraphic distribution of palynomorphs (Litwin, 1986; Litwin and others, 1991), megafossil plants (Ash, 1980; 1987), nonmarine, calcareous microfossils (Kietzke, 1989) and fossil fishes (Huber and others, in press a). Of particular importance in substantiating intergroup correlations of the Chinle are the distribution of metoposaurid, phytosaur and aetosaur genera, and the apparent restriction and first or last appearance of rhynchosaurs, dinosaurs and dicynodonts to specific stratigraphic intervals.

Lucas (1991; in press) defined the lower Chinle Group to comprise the Shinarump-Blue Mesa Sequence. These strata begin with conglomeratic sandstone (e.g. Shinarump Formation, Camp Springs Member of the Dockum Formation) and pedoturbated siltstone ("Temple Mountain Member" of Litwin and others, 1991) that rest unconformably on rocks of Permian or Lower and Middle Triassic age. These sandstones are interpreted to represent infilling of incised topography in response to rising base level in the early late Carnian (early Tuvalian). Overlying bentonitic siltstone and sandstone were deposited in low energy fluvial systems (Santa Rosa Formation), paludal and shallow-lacustrine environments (e.g. Monitor Butte Formation), and large lacustrine basins (Popo Agie Formation). The basal conglomeratic sandstone (Shinarump Formation, Camp Springs Member of the Dockum Formation) and lower siltstone (Salitral Formation, lower Tecovas Member of the Dockum Formation) contain a distinctive tetrapod fauna (Faunachron A; *Paleorhinus* Biochron) that includes the phytosaurs *Paleorhinus* and *Angistorhinus*, the aetosaurs *Longosuchus, Desmatosuchus* and *Stagonolepis*, large metoposaurids (*Metoposaurus bakeri; Buettneria perfecta*), the first appearance of dinosaurs and the only record of rhynchosaurs in the western United States (Lucas, 1990; in press; Hunt, 1990; in press; Hunt and Lucas, 1991 a, b, c; Lucas and others, 1992;). The upper part of the Shinarump-Blue Mesa Sequence contains a different fauna (Faunachron B; *Rutiodon* Biochron) dominated by the phytosaur *Rutiodon* (Ballew, 1989), several aetosaurs (*Desmatosuchus*, *Stagonolepis*, *Paratypothorax;* Long and Ballew, 1985; Hunt and Lucas, 1991 a; 1992) and the metoposaurid *Buettneria*. Both faunas possess rare, small metoposaurids (*Apachesaurus*), often abundant, large dicynodonts (*Placerias, Ischigualastia*), and rare ornithischian dinosaurs (Lucas, 1990; in press; Hunt, in press).

The lower and middle Chinle group are separated by a regional unconformity that coincides with the Carnian-Norian boundary. The middle Chinle Group (early and middle Norian) comprises the Mossback-Owl Rock Sequence and begins with extensive fluvial channel and sheet-deposited sandstone (Sonsela Member of the Petrified Forest Formation, Mossback and Trujillo Formations) that rest unconformably on the Shinarump-Blue Mesa Sequence. Above these sandstones are predominately fluvial siltstone and sandstone (Bull Canyon Formation, Painted Desert Member of the Petrified Forest Formation) deposited over a broad alluvial plain that extended from West Texas to northern Arizona and western Nevada. On the Colorado Plateau, these fluvial deposits are overlain by playa-lacustrine siltstone and pedogenic carbonates of the middle Norian Owl Rock Formation. The fauna of the middle Chinle Group (Faunachron C; *Pseudopalatus-Nicrosaurus* Biochron) includes metoposaurids (*Apachesaurus*), phytosaurs (*Pseudopalatus*), aetosaurs (*Typothorax)* and other taxa, including rare sphenosuchians and ornithischian and coelurosaurian dinosaurs (Long and Ballew, 1985; Lucas and others, 1985; Chatterjee, 1986; Small, 1989; Ballew, 1989; Hunt and Lucas, 1991 d; 1992; Hunt, 1992; in press,). Few elements of the underlying faunas are present, but some genera, particularly certain aetosaurs (*Desmatosuchus*, *Paratypothorax*), cross the Carnian-Norian boundary and range into the lower portion of the middle Chinle Group. Because the Deep River basin apparently does not preserve strata correlative with the upper Chinle Group (late Norian-Rhaetian Rock Point sequence), the latter will not be considered here.

The maximum age of the basal Chinle group is indicated by the cosmopolitan distribution of the phytosaur *Paleorhinus*. This taxon is present near the base of the Shi-

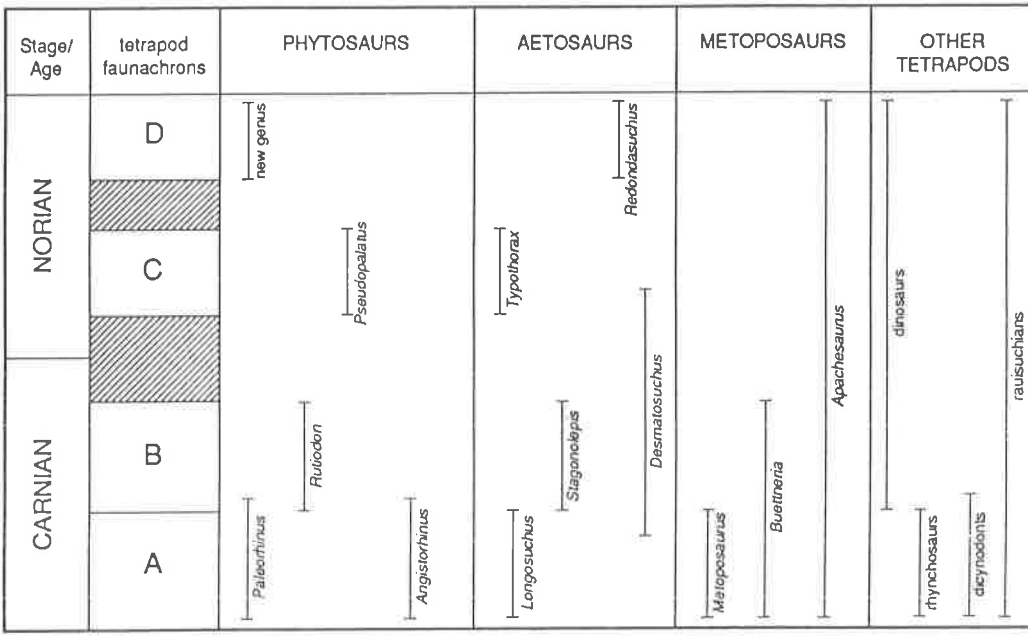


Figure 7. Late Triassic tetrapod biochronology of the Chinle Group (after Lucas, in press).

narump-Blue Mesa Sequence in Arizona, West Texas and Wyoming, and has also been found in the Blasensandstein (Germany), Argana Formation (Morocco) and Maleri and Tiki Formations (India). Of particular importance is the occurrence of *Paleorhinus* in the marine Opponitzer Schichten of Austria, which is dated by a brackish molluscan fauna of late Carnian (early Tuvalian) age, and provides a key tiepoint for correlating late Carnian marine and nonmarine biochronologies across Pangaea (Hunt and Lucas, 1991, b). In the Chinle Group, *Paleorhinus* is range-concurrent with the aetosaur *Longosuchus*. The dicynodont *Placerias hesternus* occurs in *Paleorhinus*-bearing strata of Wyoming (cf. Williston, 1904) and Arizona (Murray and Long, 1989), and extends into overlying strata that contains Faunachron B (*Rutiodon* Biochron). The Moroccan dicynodont *Moghreberia* is very similar to *Placerias*, and the two forms are probably congeneric (Lucas, 1990; Cox, 1991). *Moghreberia* is associated with the *Paleorhinus* fauna of the Argana Formation, and thus lends credence to the late Carnian age assignment for the lower Chinle Group, Argana Formation and middle Pekin Formation. The co-occurrence of *Placerias* and *Longosuchus* in the middle Pekin indicates an age no older than early late Carnian (early Tuvalian), and correlation with Chinle Group Faunachron A (Figure 8).

The Cumnock Formation overlies the Pekin Formation in the Sanford sub-basin and is in part, homotaxial with Campell and Kimball's (1923) type section of the Pekin Formation in the Wadesboro sub-basin. The combined Cumnock and type Pekin fauna includes the type species of the phytosaur *Rutiodon* (*R. carolinensis*), the aetosaur *Desmatosuchus* and ornithischian dinosaur teeth, indicating correlation with Faunachron B of the Chinle Group. Together with the middle New Oxford, upper Stockton and Lockatong Formations of the Gettysburg-Newark basin and formations containing Faunachron B tetrapods in the Chinle Group, the Cumnock Formation defines the *Rutiodon* Biochron of late Carnian (late Tuvalian) age (Huber and others, 1991; in press, a). The lower Sanford Formation in the Sanford sub-basin is the lithostratigraphic correlative of "Lithofacies Association II" in the

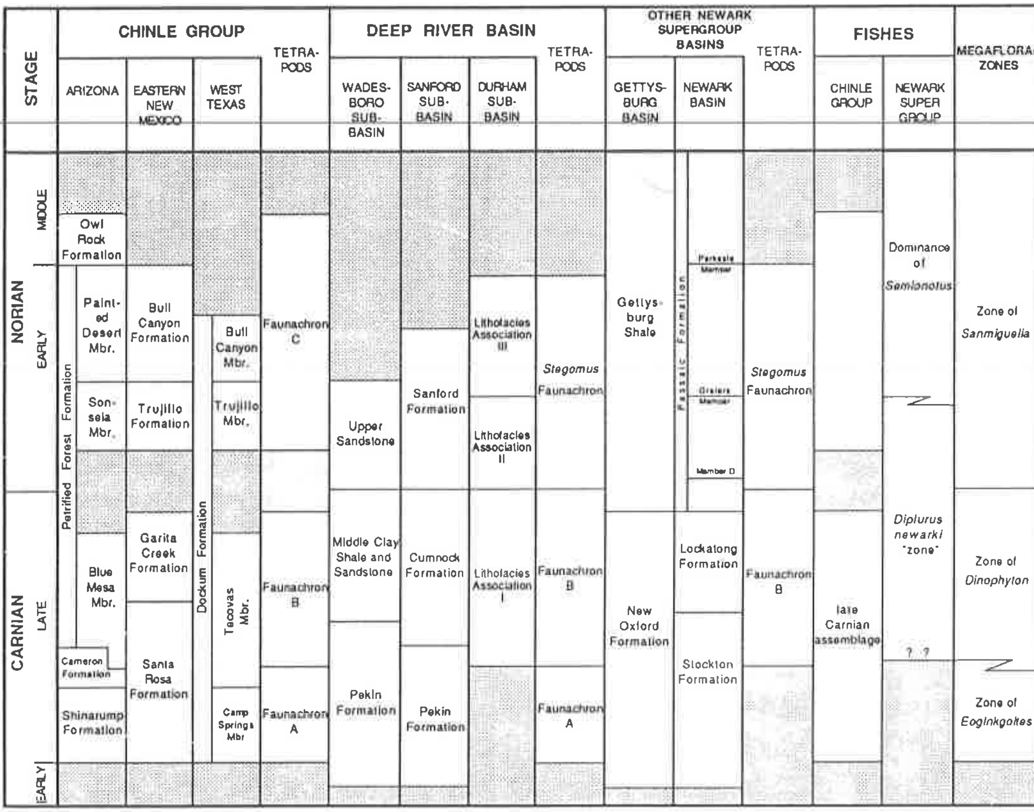


Figure 8. Revised correlation, age and biochronology of the Upper Triassic Chatham Group, strata of the Gettysburg-Newark basin and the Chinle Group. Megafloral zones after Ash (1980;1987); fish data from Olsen and others (1982) and Huber and others (in press b).

Durham sub-basin (Olsen and others, 1990). We consider both these units to be early Norian based on the occurrence of the aetosaur *Stegomus* in "Lithofacies Association II". Specimens of *Stegomus* have also been collected from the lower Passaic Formation of the Newark basin and the middle New Haven Arkose in the Hartford basin (Marsh, 1896; Jepsen, 1948; Olsen, 1980; Baird, 1986). These strata comprise the *Stegomus* Faunachron of early and middle? Norian age.

Triassic fish assemblages of the Newark Supergroup and Chinle Group share many genera in common (Huber and others, in press b). These taxa include the paleoniscids *Turseodus* and *Tanaocrossus*, redfieldiids (*Synorichthys*, *Cionichthys*) and the neopterygian *Semionotus*. Furthermore, Newark Supergroup coelacanths (*Diplurus*) and hybodont chondrichthyans (*Lissodus*, *Carinacanthus*) have closely related counterparts in the Chinle Group (*Chinlea* and *Lissodus*, respectively). The two faunas differ by the greater diversity of the Western Interior assemblages, which include xenacanthids, dipnoans, colobodontids and the hyposomatic semionotid *Hemicalypterus*. A previous attempt to define biostratigraphic zonation of Newark Supergroup basins by fish distribution recognized two assemblage 'subzones' of the "*Diplurus newarki* zone" in the Deep River basin (Olsen and others, 1982). The oldest of these was thought to be "late middle" Carnian (Cumnock Formation), while the younger assemblage was identified as late Carnian ("Lithofacies Association II"). Olsen and others (1982) correlated the "*Diplurus newarki* zone" with the Rock Point sequence fish faunas of the Chinle Group, and they considered both faunas to be no younger than late Carnian. When viewed in the context of the revised correlations presented by Huber and others (1991; in press), Late Triassic fishes in general have little biochronologic utility. To illustrate this point, we note that all Chinle Group taxa that are common to the Newark Supergroup occur in strata that span the late Carnian-Rhaetian. The presence of *Turseodus*, cf. *Cionichthys, Synorichthys, Semionotus* and *Diplurus* in "Lithofacies Association II" are not indicative of a late Carnian age, as the former four genera are found throughout the Chinle Group, and at least three of the five listed taxa have also been found in the lower and middle Passaic Formation in the Newark Basin (Schaeffer and McDonald, 1978; Olsen and others, 1982). We consider the distribution of fossil fishes in the Newark Supergroup to be largely constrained by facies control and further believe that fishes do not provide a practical nor precise means by which to correlate Late Triassic nonmarine formations (Huber and others, in press b).

Implications For Palynostratigraphy

The revised age and correlation of the Chatham Group (Figure 8) provides a refined chronostratigraphic context for the palynofloras reported by Schultze and Hope (1973), Dunay and Fisher (1974), Cornet (1977), Robbins (1985), Traverse (1986) and Robbins and others (1988). The oldest palynomorph assemblage was collected from the basal Pekin Formation in the Colon cross structure (sample PK1, Cornet, 1977). This assemblage is dominated by trilete spores (*Aratisporites* spp.), which led Traverse (1986) to conclude an early Carnian (Cordevolian) age for the basal Pekin. A stratigraphically-higher microflora (sample PK2) from the same locality as the middle Pekin vertebrate assemblage (Schultz and Hope, 1973) was dated by Cornet (1977) and Traverse (1986) as "middle" Carnian (Julian), while the Cumnock Formation contains taxa considered by these authors, Robbins (1985) and Robbins and others (1988) to be restricted to the "middle" Carnian. Additional palynomorphs were described from the type Pekin Formation (Wadesboro sub-basin) by Cornet (1977) and dated as "late-middle" or "early-late" Carnian. Cornet (1977) combined these assemblages with those he described from the Richmond-Taylorsville Basin to establish the Chatham-Richmond-Taylorsville palynostratigraphic "zone". Cornet correlated this "zone" with the Giskeuper of Europe, which he considered to be "middle" Carnian.

We disagree with these age-assignments primarily because they are contrary to those indicated by the tetrapod faunas, and also because we believe that at current levels of resolution, the viability of palynostratigraphic-based schemes to subdivide Middle and Late Triassic time has not been demonstrated. The oldest Chatham Group tetrapods (middle Pekin Formation) include indeterminate phytosaur remains and an aetosaur. Both phytosaurs and aetosaurs have not been recovered from pre-late Carnian strata in other basins of Pangaea, and we assign the middle Pekin Formation to the late Carnian and correlate it with the *Paleorhinus* biochron (Hunt and Lucas, 1991b). The co-occurrence of Chinle Group Faunachron A taxa (*Longosuchus*, *Placerias*) in the Pekin Formation indicates correlation with the basal Chinle Group and further restricts the middle Pekin microflora and fauna to the early Tuvalian substage of the late Carnian.

The Cumnock Formation contains the most diverse of the Chatham Group tetrapod assemblages. The biochronologic significance of Cumnock labyrinthodont, rauisuchian and therapsid fossils are not clear because the collective materials are either taxonomically-indeterminate (e.g. *Dictyocephalus*, *Zamotus*) or have a very limited distribution outside of the Deep River Basin (e.g. *Microconodon*). In contrast, the presence of the phytosaur *Rutiodon*, the aetosaur *Desmatosuchus* and an ornithischian dinosaur in the Cumnock Formation and equivalent beds of the type Pekin indicate a late Carnian age. Specifically, *Rutiodon* is considerably more-derived than the stem phytosaur *Paleorhinus*, and as mentioned, aetosaurs and primitive dinosaurs have not been found in strata older than early Tuvalian. In the Chinle Group, tetrapod assemblages that contain *Paleorhinus*, *Longosuchus* and *Placerias* are succeeded by a more derived fauna that includes *Rutiodon* and a greater diversity of aetosaurs (but not *Longosuchus*). The two faunas are found in direct stratigraphic superposition on the Colorado Plateau, in eastern New Mexico and West Texas, and in the Pekin and Cumnock Formations. *Rutiodon* and other tetrapods common to Chinle Group Faunachron B also occur in the middle New Oxford, upper Stockton and possibly Lockatong Formations of the Gettysburg-Newark Basin and in the upper Cow Branch Formation in the Danville-Dan River Basin. We believe this distribution provides ample evidence that all formations containing *Rutiodon* (including the Cumnock) are of late Carnian age and can be correlated with the late Tuvalian substage (Huber and others, 1991; in press a, b).

We compared palynomorph distribution in the Chatham group, Newark basin and Chinle Group to test the viability of palynostratigraphic-based correlations of these strata in light of the age-assignments indicated by vertebrate distributions (Figure 9). The basal Pekin Formation palynoflora (PK 1, Cornet, 1977; Traverse, 1986) is distinct when compared to other Chatham Group microfloras and those from the Gettysburg-Newark Basin and the Chinle Group, and represents one of the few documented, Julian-age palynofloras from North America occurring outside of the lower Tuckahoe Formation of the Richmond Basin (Cornet and Olsen, 1990) and the Arctic Archipelago (Fisher, 1979). The middle Pekin microflora (Schultz and Hope, 1973; sample PK 2 of Cornet, 1977; Traverse, 1986) shares 14 species in common with described lower Chinle Group microfloras (Dunay and Traverse, 1971; Gottsfeld, 1972; Dunay and Fisher, 1974; 1979; Stone, 1978; Fisher and Dunay, 1984; Litwin, 1985; Litwin, 1986; Litwin and others, 1990), while several of the remaining taxa occur in the Norian-basal Hettangian Passaic Formation in the Newark Basin (cf. Folwell and Olsen, in press). Of the 37 taxa reported from the Cumnock Formation by Cornet (1977), Robbins (1985), Traverse (1986) and Robbins and others (1988), 25 of these were encountered by Dunay and Traverse (1971), Dunay and Fisher (1974; 1979), Stone (1978), Fisher and Dunay, 1984; Litwin (1986) and Litwin and others (1991) in lower Chinle Group strata we judge to be the same age as the Cumnock Formation (*Rutiodon* Biochron). Thus, Robbins (1985) and Robbins and others (1988) assertion that the presence of *Cycadopites* "sp. *103* (*C.* "sp. *A*" of Litwin, 1986), *Plicatisaccus badius*, *Colpectopollis ellipsoideus* and *Klauspollenites gouldii* restricts the Cumnock Formation to the "middle" Carnian is unsubstantiated. All these species occur in late Carnian strata of the lower Chinle Group, *Plicatisaccus badius* crosses the Carnian-Norian boundary on the Colorado Plateau and the later two species occur in the late Norian-Rhaetian Rock Point Formation (cf. Litwin, 1986; Litwin and others, 1991).

The purpose of this discussion is to illustrate the most problematic aspect of Newark Supergroup palynostratigraphy; i.e. identifying taxa whose stratigraphic occurrences imply temporal significance. The age assignments for Chatham Group strata made by Robbins (1985), Traverse (1986) and Robbins and others (1988) were based on comparison of palynofloras from other Newark Supergroup basins and general acceptance of Cornet's (1977) initial palynostratigraphic zonation proposal. Comparisons of Newark Supergroup and European Triassic palynofloras were discussed in varying detail by these authors, but the age assignments and/or temporal ranges of individual taxa in the cited European analogs was seldom questioned. Critical to correlating Newark Supergroup with European microfloras is refined age control for the lower Keuper and definition of the Carnian-Norian boundary in the Germanic basin and Britain. Unlike their counterparts in the Alpine and circum-Mediterranean realms, Keuper palynofloras in general,

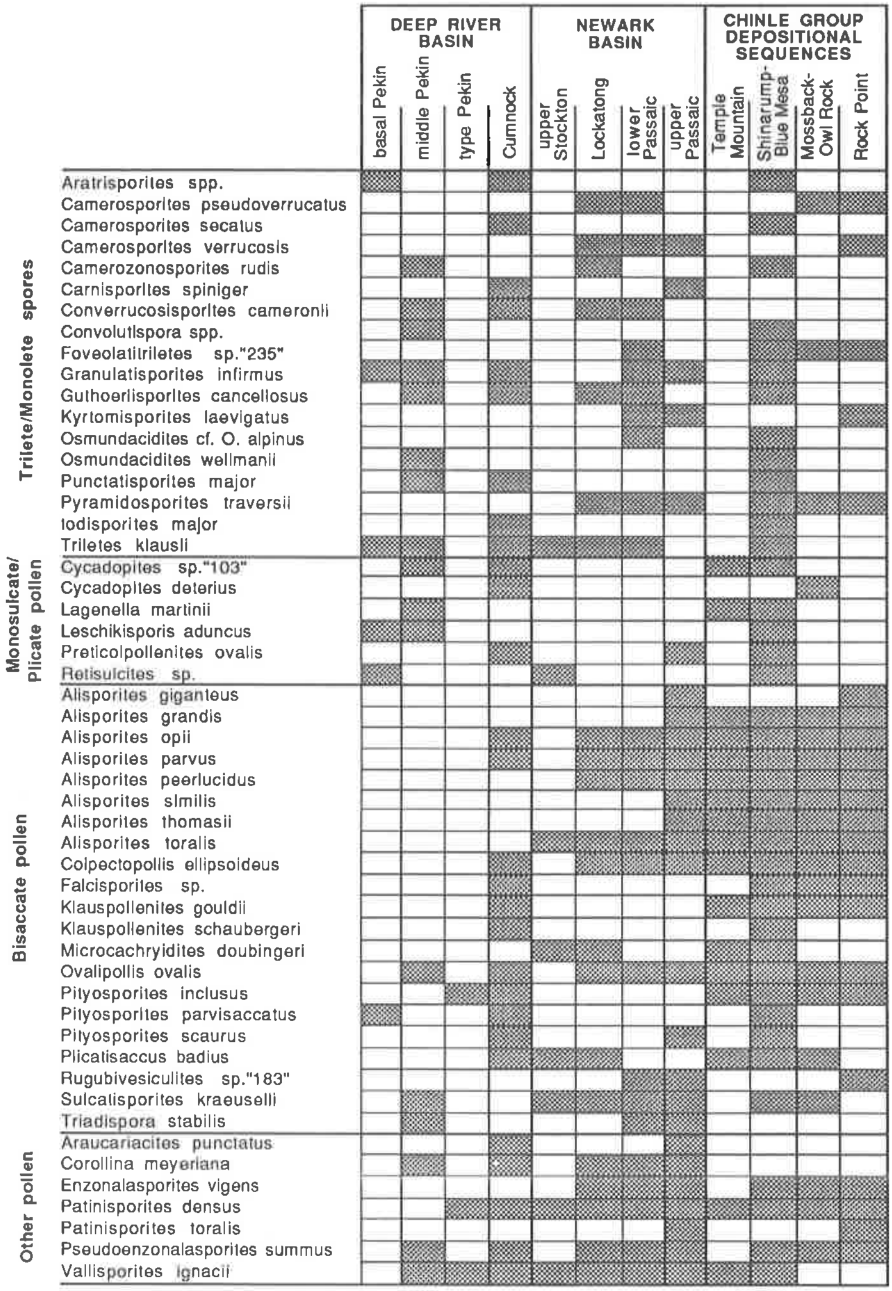


Figure 9. Distribution of palynomorphs common to the Chatham Group, Newark basin and/or Chinle Group (data from Cornet, 1977; Robbins, 1985; Litwin, 1986; Traverse, 1986; Robbins and others, 1988; Litwin and others, 1991; Folwell and Olsen, in press).

lack the independent age control afforded by ammonoids, other molluscs and conodonts.

An example from the lower Keuper which has played a significant role in biasing age-assignments of Newark Supergroup palynofloras is the repeated misinterpretation of the age of the Schilfsandstein and it's supposed British lithostratigraphic correlative, the Arden Sandstone. Gieger and Hopping (1968), Ager (1970), Warrington (1970), Fisher (1972, a, b), Hahn (1984) Benton (1986; 1991) and Mader (1990) have dated these units by various methods (palynostratigraphy, tetrapod biochronology, lithostratigraphy, magnetostratigraphy) as late Carnian (Tuvalian), and most of these authors place the Carnian-Norian boundary at the top of the Schilfsandstein and Arden Sandstone, and regard the overlying Blasensandstein and Bunte Mergel (Germany) and Keuper Marls (Britain) to be Norian. As previously stated, the Blasensandstein contains a *Paleorhinus* fauna and is late Carnian (early Tuvalian) in age. In addition, the Blasensandstein and Schilfsandstein are separated stratigraphically by the Rote Wand and Unterer Bunte Mergel. Thus, the Schilfsandstein and Rote Wand-Unterer Bunte Mergel, which form a genetic sequence onto themselves (Aigner and Bachmann, 1991; 1992), can be no younger than early Carnian and are more properly referred to as middle and late Julian, respectively. On face value, this correlation might prompt re-evaluation of Newark Supergroup palynofloras previously dated as late Carnian, with efforts directed toward expanding definition of the Newark "middle" Carnian. However, this would only compound the problems already created by the overemphasis placed on the reliability of Newark Supergroup palynostratigraphy, and would only serve to widen the disparity between palynological and vertebrate-based chronologies. The point we illustrate by the above example is that palynology is not the panacea for the long-standing problems of regional and intercontinental Triassic correlations, and that more reliable correlations are possible using other biostratigraphic and chronostratigraphic methods.

The revised age and correlation of the Chatham Group also impacts on the reported "middle" Carnian extinction event of Cornet (1977), Cornet and Olsen (1985; 1990) and Olsen and Sues (1986). Cornet's (1977) recognition of a palynofloral turnover within the "middle "Carnian or between the "middle" and late Carnian was in large part based on the assumption that this temporal interval was represented by the Giskeuper-Schilfsandstein transition, which as discussed above, is incorrect. Palynofloras from these lower Keuper units are demonstrably-older (Julian) than the Pekin-Cumnock transition (Tuvalian). Furthermore, Cornet (1977), Cornet and Olsen (1985; 1990), Robbins (1985), Ediger (1986), Traverse (1986; 1987), Olsen and Sues (1986), Robbins and others (1988) and Olsen and others (1989; 1990) have treated the "middle" Carnian as a distinct substage of Late Triassic time without definition, while the traditional ammonoid biochronology on which temporal division of the Triassic is based, recognizes a two-fold subdivision of the Carnian into early (Cordevolian and Julian) and late (Tuvalian) substages. Furthermore, there is considerable disagreement as to the chronostratigraphic utility and definition of the Cordevolian as a distinct substage (Lieberman, 1980) and comparison of recent radiometric timescales for the Triassic (Kerp, 1991) indicates that there is little consensus as to the temporal length of the Carnian. Thus, it is not clear what is meant by the repeated references to the Newark Supergroup "middle" Carnian cited above. We believe that identification of a "middle" Carnian extinction event was based on incorrect assumptions concerning the age of lower Newark Supergroup strata. The only other basin where this biotic turnover (restricted to pollen and spore taxa) has been identified is the Richmond-Taylorsville basin complex of Virginia. Unfortunately, the strata contained by these basins lack age-diagnostic tetrapods, except for indeterminate phytosaur remains (Sues and Olsen, 1990) and the therapsid *Microconodon* (Olsen and others, 1989; Cornet and Olsen, 1990), both of which may indicate a late Carnian age for at least the Turkey Branch Formation (Huber and others, in press a).

In summary, we believe that palynofloras from the Deep River basin do not provide evidence that the bulk stratigraphic thickness of the Chatham Group is of early and "middle" Carnian age. We qualify this statement by noting that most palynomorph taxa found in the

Pekin and Cumnock Formations occur in formations of other regions that are dated as late Carnian through basal Hettangian (Figure 9). Of equal importance, strict comparison of pollen and spore taxa common to the Chatham Group, formations of the Newark basin and the Chinle Group, strata that can be directly correlated by their tetrapod faunas, do not contradict the revised age assignments. In light of the above-discussion, we believe that palynomorphs do not possess the regional time-stratigraphic and distributional consistency that are necessary to define boundary points of stages and substages of Late Triassic time in general, and particularly within the Newark Supergroup.

Late Triassic tetrapods, in contrast, do not suffer from the same limitations imposed by palynostratigraphic analysis. In regions where successive and superimposed faunas are preserved, especially in the Germanic Keuper, Newark Supergroup and Chinle Group, many archosaurian groups demonstrate a "zonal" distribution of genera and are very useful in establishing local and regional biochronologies (e.g. Figure 7). Furthermore, certain taxa, such as *Metoposaurus*, *Buettneria*, *Scaphonyx*, *Paleorhinus*, *Rutiodon*, *Paratypothorax*, *Placerias* and *Ischigualiastia* are found in widely-separated regions across Pangaea in formations of the same age (e.g. Hunt and Lucas, 1991 a, b, c; Hunt, in press; Lucas and Hunt, in review). Distributional data of these and other taxa, based on precise taxonomy, provide an unprecedented level of resolution for correlating nonmarine late Carnian formations. In addition, the fortuitous occurrence of *Paleorhinus* in marine sediments, coupled with this taxon's cosmopolitan distribution (Hunt and Lucas, 1991 b), constrains the age of faunas long considered to be Carnian (e.g. Argana, Maleri, Tiki, basal Chinle) to the early Tuvalian (*Dilleri* zone).

Though the Chatham Group fauna is comparatively small, the stratigraphic succession of it's tetrapod assemblages provides a pivotal tie-point for correlating faunas of the Newark Supergroup and Chinle Group. Vertebrate assemblages from the Chatham Group, formations of the Gettysburg-Newark basin and lower Chinle Group share several taxa in common that allow for geographic extension of the Chinle Group tetrapod biochronology to the Newark Supergroup. The similarities of these assemblages further suggests that Laurasia was characterized by a high degree of faunal homogeneity during the late Carnian. The extent of this homogeneity is underscored when faunas from the above regions are compared with those from the Wolfville (Nova Scotia), Lossiemouth (Scotland), Blasensandstein (Germany) and Argana (Morocco) Formations, and with Gondwanan faunas of the Ischigualasto (Argentina), Maleri and Tiki (India) Formations. These faunas share in common the same or closely-related genera of metoposaurids, rhynchosaurs, aetosaurs, phytosaurs, dinosaurs and dicynodonts that have restricted stratigraphic ranges in the few regions that preserve younger (Norian), superimposed faunas (e.g. Chinle Group, Los Colorados Formation (Argentina), middle and upper Keuper). The fact that most of the younger, Norian faunas share few taxa with each other has been cited as evidence for an extinction event at the Carnian-Norian boundary (e.g. Benton, 1986; 1991), but we interpret this trend to be a function of increasing provincialism over time, probably related to the rifting and initial breakup of Pangaea during the Norian, as these events would impose potential barriers to faunal exchange between the regions mentioned above.

ACKNOWLEDGMENTS

We thank Paul E. Olsen for invaluable discussions and generous access to unpublished data that concern many aspects of Newark Supergroup stratigraphy, paleontology and biochronology. Special thanks go to Nicholas G. McDonald, Paul E. Olsen, Eleanora I. Robbins and Daniel A. Textoris for providing literature on short notice. Ronald Rukstela drafted most figures in the text, and Paul E. Olsen provided the photograph of *Stegomus* (Figure 5). Finally, we acknowledge Pamela J. W. Gore for her critical review of the manuscript and comments which improved it's quality.

REFERENCES

Ager, D. V., 1970, The Triassic System in Britain and its stratigraphical nomenclature: Quarterly Journal of the Geological Society of London, v. 126, p. 3-17.

Aigner, T. and Bachmann, G. H., 1991, Sequence stratigraphic concept of the Germanic Triassic: Albertiana, v. 9, p. 24-25.

Aigner, T. and Bachmann, G. H., 1992, Sequence stratigraphic framework of the German Triassic: Sedimentary Geology, v. 80, p. 115-135.

Ash, S. R., 1980, Upper Triassic floral zones of North America, *in* Dilcher, D. L. and Taylor, T. M., eds., Biostratigraphy of Fossil Plants: Stroudsburg, PA, Dowden, Hutchinson and Ross, Inc., p. 153-170.

Ash, S. R., 1987, The Upper Triassic redbed flora of the Colorado Plateau, western United States: Journal of the Arizona-Neveda Academy of Science, v. 22, p. 95-105.

Bain, G. L. and Harvey, B. W., 1977, Field guide to the geology of the Durham Triassic basin: Carolina Geological Society, 40th Annual Meeting, Geology and Mineral Resources Section, Raleigh, 84 p.

Baird, D., 1986, Some Upper Triassic reptiles, footprints, and an amphibian from New Jersey: The Mososaur, v. 3, p. 125-135.

Baird, D. and Patterson, O. F. III, 1967, Dicynodont-archosaur fauna in the Pekin Formation (Upper Triassic) of North Carolina: Geological Society of America Special Paper, no. 115, p. 11.

Ballew, K. A., 1989, Phylogenetic analysis of Phytosauria from the Late Triassic of the western United States, *in* Lucas, S. G. and Hunt, A. P., eds., Dawn of the Age of Dinosaurs in the American Southwest: Albuquerque, NM, New Mexico Museum of Natural History, p. 309-339.

Bazard, D. R. and Butler, R. F., 1991, Paleomagnetism of the Chinle and Kayenta Formations, New Mexico and Arizona: Journal of Geophysical Research, v. 96, p. 9847-9871.

Benton, M. J., 1986, The Late Triassic tetrapod extinction events, *in* Padian, K., ed., The Beginning of the Age of Dinosaurs: Faunal Change Across the Triassic-Jurassic Boundary: Cambridge, Cambridge University Press, p. 303-320.

Benton, M. J., 1991, What really happened in the Late Triassic?: Historical Biology, v. 5, p. 263-278.

Camp, C. L. and Welles, S. P., 1956, Triassic dicynodont reptiles, Part I. The North American genus *Placerias*: Memoirs of the University of California, v. 13, p. 255-304.

Campbell, M. R. and Kimball, K. W., 1923, The Deep River coal field of North Carolina: North Carolina Geological and Economical Survey Bulletin, no. 35, 95 p.

Chatterjee, S., 1986, The Late Triassic Dockum vertebrates: Their stratigraphic and paleobiogeographic significance, *in* Padian, K., ed., The beginning of the Age of Dinosaurs: Faunal Change Across the Triassic-Jurassic Boundary: Cambridge, Cambridge University Press, p. 139-150.

Colbert, E. H., 1947, Studies of the phytosaurs *Machaeroprosopus* and *Rutiodon*: Bulletin of the American Museum of Natural History, v. 88, p. 53-96.

Colbert, E. H., 1965, A phytosaur from North Bergen, New Jersey: American Museum Novitates, no. 2230, 25 p.

Colbert, E. H. and Gregory, J. T., 1957, Correlation of continental Triassic sediments by vertebrate fossils: Geological Society of America Bulletin, v. 68, p. 1456-1467.

Colbert, E. H. and Imbrie, J., 1956, Triassic metoposaurid amphibians: Bulletin of the American Museum of Natural History, v. 110, p. 403-452.

Conrad, K., Lockley, M. G. and Prince, N. K., 1987, Triassic and Jurassic vertebrate-dominated trace fossil assemblages of the Cimarron Valley region: Implications for paleoecology and biostratigraphy: New Mexico Geological Society Guidebook, no. 41, p. 127-138.

Cope, E. D., 1871, Observations on the distribution of certain extinct vertebrates in North Carolina: Proceedings of the American Philosophical Society, v. 12, p. 210-216.

Cornet, B., 1977, The palynostratigraphy and age of the Newark Supergroup: Pennsylvania State University, unpublished Ph.D dissertation, 506 p.

Cornet, B. and Olsen, P. E., 1985, A summary of biostratigraphy of the Newark Supergroup of eastern North America, with comments on Early Mesozoic provinciality, *in* Weber, R., ed., III Congreso Latinoamericano de Paleontologia Mexico, Simposio Sobre Floras del Triassico, su Fitografia y Palecologia Memoria, p. 67-81.

Cornet, B. and Olsen, P. E., 1990, Early to middle

Carnian (Triassic) flora and fauna of the Richmond and Taylorsville basins, Virginia and Maryland, U.S.A.: Virginia Museum of Natural History, Guidebook no. 1, 87 p.

Cox, C. B., 1991, The Pangaea dicynodont *Rechisaurus* and the comparative biostratigraphy of Triassic dicynodont faunas: Palaeontology, v. 34, p. 767-784.

Davidow-Henry, B., 1989, Small metoposaurid amphibians from the Triassic of western North America and their significance, *in* Lucas, S. G. and Hunt, A. P., eds., Dawn of the Age of Dinosaurs in the American Southwest: Albuquerque, NM, New Mexico Museum of Natural History, p. 278-292.

Delevoryas, T. and Hope, R. C., 1971, A new Triassic cycad and its phyletic implications; Postilla, no. 150, 21 p.

Delevoryas, T. and Hope, R. C., 1973, Fertile coniferophyte remains from the Late Triassic Deep River basin, North Carolina: American Journal of Botany, v. 60, p. 810-818.

Dunay, R. E. and Fisher, M. J., 1974, Late Triassic palynofloras of North America and their European correlatives: Review of Paleobotany and Palynology, v. 17, p. 179-186.

Dunay, R. E. and Fisher, M. J., 1979, Palynology of the Dockum Group (Upper Triassic), Texas, U.S.A.: Review of Palaeobotany and palynology, v. 28, p. 61-92.

Emmons, E., 1856, Geologic Report of the Midland Counties of North Carolina: New York, George P. Putnam and Co., 247 p.

Emmons, E., 1857, American Geology, Part VI: Albany, NY, Sprague and Co., 152 p.

Emmons, E., 1860, Manual of Geology, Designed for the Use of Colleges and Academics: New York, A. S. Barnes and Burr, 297 p.

Fisher, M. J., 1972 a, Rhaeto-Liassic palynomorphs from the Barnstone railway cutting, Nottinghamshire: The Mercian Geologist, v. 4, p. 101-106.

Fisher, 1972 b, The Triassic palynofloral succession in England: Geoscience and Man, v. 4, p. 101-109.

Folwell, S. J. and Olsen, P. E., in press, Time-calibration of Triassic-Jurassic microfloral turnover, eastern North America: Tectonophysics.

Fontaine, W. M., 1883, Contributions to the knowledge of the older Mesozoic flora of Virginia; United States Geological Survey Monograph, no. 6, 144 p.

Froelich, A. J. and Olsen, P. E., 1984, Newark Supergroup, a revision of the Newark Group in eastern North America: United States Geological Survey Bulletin 1537A, p. A55-A58.

Geiger, M. E. and Hopping, C. A., 1968, Triassic stratigraphy of the southern North Sea basin: Philosophical Transactions of the Royal Society of London, Series B, v. 254, p. 1-36.

Gensel, P., 1986, Plant fossils of the Upper Triassic Deep River Basin, *in* Textoris, D. A., ed., Society of Economic Paleontologists and Mineralogists Field Guidebook, Third Annual Midyear Meeting, Raleigh, North Carolina, p. 82-86.

Gore, P. J. W., 1986, Depositional framework of a Triassic rift basin: the Durham and Sanford subbasins of the Deep River basin, North Carolina, *in* Textoris, D. A., ed., Society of Economic Paleontologists and Mineralogists Field Guidebook, Third Annual Midyear Meeting, Raleigh, North Carolina, p. 56-66.

Gregory, J. T., 1980, The otic notch of metoposaurid labyrinthodonts, *in* Jacobs, L. L., ed., Aspects of Vertebrate History: Essays in Honor of Edwin Harris Colbert: Flagstaff, AZ, Museum of Northern Arizona press, p. 125-136.

Hahn, G. G., 1984, Palaomagnetische Untersuchungen im Schilfsandstein (Trias, Km2) Westeuropas: Geologische Rundschau, v. 73, p. 499-516.

Haubold, H., 1986, Archosaur footprints at the terristrial Triassic-Jurassic transition, *in* Padian, K., ed., The Beginning of the Age of Dinosaurs: Faunal Change Across the Triassic-Jurassic Boundary, p. 189-201.

Hoffman, C. W. and Gallagher, P. E., 1989, Geology of the Southeast and Southwest Durham 7.5-minute quadrangles, North Carolina: North Carolina Geological Survey Bulletin, no. 92, 34 p.

Hope, R. C., 1970, *Pekinopteris auriculata*, a new plant from the North Carolina Triassic: Journal of Paleontology, v. 44, p. 1137-1139.

Hope, R. C. and Patterson, O. F. III, 1969, Upper Triassic flora from the Deep River Basin: North Carolina Department of Conservation and Development Special Publication, no. 2, 22 p.

Huber, P., Lucas, S. G. and Hunt, A. P., 1991, Late Triassic tetrapod biochronology of the Newark Supergroup, eastern North America: Geological Society of America, Abstracts with Programs, v. 23, p. 47.

Huber, P., Lucas, S. G. and Hunt, A. P., in press a, Vertebrate biochronology of the Newark Supergroup Triassic, *in* Lucas, S. G. and Morales, M., eds., Proceedings of the Symposium on the Nonmarine Triassic: Albuquerque, NM, New Mexico

Museum of Natural History.

Huber, P., Lucas, S. G. and Hunt, A. P., in press b, Late Triassic fish assemblages of the North American Western Interior and their biochronologic significance: Flagstaff, AZ, Museum of Northern Arizona Press.

Hunt, A. P., 1990, The early diversification pattern of dinosaurs in the Late Triassic: Modern Geology, v. 16, p. 43-60.

Hunt, A. P., 1992, Aetosaurs from the Bull Canyon Formation (Late Triassic: Norian) of eastern New Mexico and their biochronologic significance: New Mexico Geology, v. 14, p. 66.

Hunt, A. P., in press, A revision of the Metoposauridea (Amphibia: Temnospondyli) of the Late Triassic: Flagstaff, AZ, Museum of Northern Arizona Press.

Hunt, A. P. and Lucas, S. G., 1991 a, Re-evaluation of "Typothorax *meadei*", a Late Triassic aetosaur from the United States: Palaeontologische Zeitschrift, v. 64, p. 317-328.

Hunt, A. P. and Lucas, S. G., 1991 b, The *Paleorhinus* Biochron and the correlation of the nonmarine Upper Triassic of Pangaea: Palaeontology, v. 34, p. 487-501.

Hunt, A. P. and Lucas, S. G., 1991 c, A new rhynchosaur from the Upper Triassic of West Texas, and the biochronology of Late Triassic rhynchosaurs: Palaeontology, v. 34, p. 927-938.

Hunt, A. P. and Lucas, S. G., 1991 d, *Rioarribisaurus*, a new name for a Late Triassic dinosaur from New Mexico (USA): Palaeontologische Zeitschrift, v. 65, p. 191-198.

Hunt, A. P. and Lucas, S. G., 1992, The first occurrence of *Paratypothorax andressi* (Reptilia, Aetosauria) in the western United States and its biochronologic significance: Palaeontologische Zeitschrift, v. 66, p. 147-157.

Hunt, A. P. and Lucas, S. G., in press, Ornithischian dinosaurs from the Upper Triassic of the United States, *in* Fraser, N. C. and Sues, H. D., eds., In the Shadow of the Dinosaurs, Early Mesozoic Tetrapods: New York, Columbia University Press.

Jepsen, G. L., 1948, A Triassic armored reptile from New Jersey: New Jersey Department of Conservation Miscellaneous Geologic Paper, no. 20, 20 p.

Kerp, H., 1991, Some recently published (numerical) time scales compared: Albertiana, v. 9, p. 19-21.

Kietzke, K. K., 1989, Calcareous microfossils from the Triassic of the southwestern United States, *in* Lucas, S. G. and Hunt, A. P., eds., Dawn of the Age of Dinosaurs in the American Southwest: Albuquerque, NM, New Mexico Museum of Natural History, p. 223-232.

Koob, J. D., 1961, Triassic pollen and spore flora of the Cumnock Formation (Newark Series) of North Carolina: University of Massachusetts, Amherst, unpublished Masters thesis, 47 p.

Leidy, J., 1856, Notices of remains of extinct vertebrate animals discovered by Professor E. Emmons: Proceedings of the Philadelphia Academy of Natural Sciences, v. 8, p. 255-257.

Lieberman, H. M., 1980, The suitability of the Raibl sequence as a stratotype for the Carnian Stage and the Julian Substage of the Triassic: Newsletters On Stratigraphy, v. 9, p. 35-42.

Litwin, R. J., 1986, The palynostratigraphy and age of the Chinle and Moenave Formations, southwestern USA: State College, PA, Pennsylvania State University, unpublished Ph.D dissertation, 265 p.

Litwin, R. J., Traverse, A. and Ash, S. R., 1991, Preliminary palynological zonation of the Chinle Formation, southwestern USA, and its correlation to the Newark Supergroup (eastern USA): Review of Paleobotany and Palynology, v. 68, p. 269-287.

Lucas, S. G., 1990, Toward a vertebrate biochronology of the Triassic: Albertiana, v. 8, p. 36-41.

Lucas, S. G., 1991, Sequence stratigraphic correlation of nonmarine and marine Late Triassic biochronologies: Albertiana, v. 9, p. 11-18.

Lucas, S. G., in press, The Chinle Group: Revised Upper Triassic stratigraphy and chronology on the Colorado Plateau: Flagstaff, AZ, Museum of Northern Arizona Press.

Lucas, S. G. and Hunt, A. P., 1989, New genotype designations for the phytosaurs *Mystriosuchus* and *Rutiodon* with a discussion of the taxonomic status of *Mystriosuchus*, *Clepsysaurus* and *Rutiodon*, *in* Lucas, S. G. and Hunt, A. P., eds., Dawn of the Age of Dinosaurs in the American Southwest: Albuquerque, NM, New Mexico Museum of Natural History, p. 340-348.

Lucas, S. G. and Hunt, A. P., in review, A Late Triassic dicynodont from New Mexico and the biochronology of Late Triassic dicynodonts: Journal of Paleontology.

Lucas, S.G., Hunt, A.P. and Long, R.A., 1992, The oldest dinosaurs: Naturwissenschaften, v. 79, p. 171-172.

Luttrell, G.W., 1989, Stratigraphic nomenclature of the Newark Supergroup of eastern North America: United States Geological Survey Bulletin,

no. 1572, 136 p.
McGregor, J.H., 1906, The Phytosauria, with especial reference to *Mystriosuchus* and *Rhytidodon*: American Museum of Natural History Memoir, v. 9, p. 29-101.
Mader, D., 1990, Palaeoecology of the flora in Buntsandstein and Keuper in the Triassic of Middle Europe: New York, Gustav Fischer Verlag, 1582 p.
Manspeizer, W., 1988, Triassic-Jurassic rifting and opening of the Atlantic: An overview, in Manspeizer, W., ed., Triassic- Jurassic Rifting, Continental Breakup, and the Origin of the Atlantic Ocean and Passive Margins: New York, Elsevier, p. 41- 79.
Marsh, O.C., 1896, A new beledont reptile (*Stegomus*) from the Connecticut River sandstone: American Journal of Science, v. 2, p. 59-62.
Molina-Garza, R.S., Geissman, J.W., Van der Voo, R., Lucas, S.G., and Hayden, S.N., 1991, Paleomagnetism of the Moenkopi and Chinle Formations in central New Mexico: Implications for the North American apparent polar wander path and Triassic magnetostratigraphy: Journal of Geophysical Research, v. 96, p. 14239-14262.
Murray, P.A. and Long, R.A., 1989, Geology and paleontology of the Chinle Formation, Petrified Forest National Park and vicinity, Arizona and a discussion of vertebrate fossils of the southwestern Upper Triassic, *in* Lucas, S.G. and Hunt, A.P., Dawn of the Age of Dinosaurs in the American Southwest: Albuquerque, NM, New Mexico Museum of Natural History, p. 29-64.
Olsen, P.E., 1977, Stop 11, Triangle Brick Quarry, *in* Bain, G.L. and Harvey, B.W., Field Guide to the Geology and of the Durham Triassic Basin: Raleigh, NC, Carolina Geological Society Fieldtrip Guidebook, p. 59-60.
Olsen, P.E., 1980, A comparison of the vertebrate assemblages from the Newark and Hartford Basins (early Mesozoic, Newark Supergroup) of eastern North America, *in* Jacobs, L.L., ed., Aspects of Vertebrate History: Essays in Honor of Edwin Harris Colbert: Flagstaff, AZ, Museum of Northern Arizona Press, p. 35- 53.
Olsen, P.E., 1988, Paleontology and paleoecology of the Newark Supergroup (Early Mesozoic, eastern North America), *in* Manspeizer, W., ed., Triassic-Jurassic Rifting, Continental Breakup, and the origin of the Atlantic Ocean and Passive Margins: New York, Elsevier, p. 186-230.
Olsen, P.E., and Baird, D., 1986, The ichnogenus *Atreipus* and its significance for Triassic biostratigraphy, *in* Padian, K., ed., The Beginning of the Age of Dinosaurs: Faunal Change Across the Triassic-Jurassic Boundary: Cambridge, Cambridge University Press, p. 61-87.
Olsen, P.E., Schilsche, R.W. and Gore, P.J.W., 1989, Tectonic, Depositional and Paleoecological History of Early Mesozoic Rift Basins, eastern North America: Washington, D.C., American Geophysical Union, 174 p.
Olsen, P.E. and Flynn, J.J., 1989, Field guide to the vertebrate paleontology of Late Triassic rocks in the southwestern Newark basin (Newark Supergroup, New Jersey and Pennsylvania): The Mososaur, v. 4, p. 1-43.
Olsen, P.E., Froelich, A.J., Daniels, D.L., Smoot, J.P. and Gore, P.J.W., 1990, Rift basins of Early Mesozoic age, *in* Horton, J.W. and Zullo, V.A., eds., The Geology of the Carolinas: Knoxville, TN, University of Tennessee Press, p. 142-170.
Olsen, P.E., McCune, A.R. and Thomson, K.S., 1982, Correlation of the Early Mesozoic Newark Supergroup by vertebrates, principally fishes: American Journal of Science, v. 282, p. 1-44.
Olsen, P.E. and Sues, H.D., 1986, Nature of the Triassic-Jurassic faunal transition, *in* Padian, K., ed., The Beginning of the Age of Dinosaurs: Faunal Change Across the Triassic-Jurassic Boundary: Cambridge, Cambridge University Press, p. 321-351.
Parker, J.M. III, 1966, Triassic reptilian fossil from Wake County, North Carolina: Journal of the Elisha Mitchell Society, v. 82, p. 92.
Parker, J.M., 1979, Geology and mineral resources of Wake County: North Carolina Department of Natural Resources and Community Development, Geological Survey Section Bulletin, no. 86, 122 p.
Patterson, O.F. III, 1969, The depositional environment and paleoecology of the Pekin Formation (Upper Triassic) of the Sanford Triassic basin, North Carolina: North Carolina State University, unpublished Masters thesis, 104 p.
Redfield, W.C., 1856, On the relations of the fossil fishes of the sandstone of Connecticut and other Atlantic states to the Liassic and Oolitic Periods: American Journal of Science, v. 22, p. 357-363.
Reinemund, J.A., 1955, Geology of the Deep River coal field, North Carolina: United States Geological Survey Professional Paper, no. 246, 159 p.
Robbins, E.I., 1985, Palenostratigraphy of coal-bearing sequences in Early Mesozoic basins of the eastern United States, *in* Robinson, G.R. and

Froelich, A.J., eds., Proceedings of the Second United States Geological Survey Workshop on the Early Mesozoic Basins of the Eastern United States: United States Geological Survey Circular, no. 946, p. 27-29.

Robbins, E.I., Wilkes, G.P. and Textoris, D.A., 1988, Coal deposits of the Newark rift system, *in* Manspeizer, W., ed., Triassic-Jurassic Rifting, Continental Breakup, and the Origin of the Atlantic Ocean and Passive Margins: New York, Elsevier p. 649-682.

Schaeffer, B. and McDonald, N.G., 1978, Redfieldiid fishes from the Triassic-Jurassic Newark Supergroup of eastern North America: America Museum of Natural History Bulletin, v. 155, p. 131-173.

Schultz, G. and Hope, R.C., 1973, Late Triassic microflora from the Deep River basin, North Carolina: Paleontographica, B, v. 141, p. 63-88.

Small, B.J., 1989, Aetosaurs from the Upper Triassic Dockum Formation, Post Quarry, West Texas, *in* Lucas, S.G. and Hunt, A.P., eds., Dawn of the Age of Dinosaurs int he American Southwest: Albuquerque, NM, New Mexico Museum of NaturalHistory, p. 301-308.

Smoot, J.P., Froelich, A.J. and Luttrell, G.W., 1988, Uniform symbols for the Newark Supergroup, *in* Froelich, A.J. and Robinson, G.R., eds., Studies of Early Mesozoic Basins of the Eastern United States: United States Geological Survey Bulletin, no. 1776, p. 1-6.

Stone, J.F., 1978, Pollen and Spores, *in* Ash, S.R., ed., Geology, Paleontology and Paleoecology of a Late Triassic Lake, Western New Mexico: Bringham Young University Geological Studies Series, v. 25, p. 45-59.

Sues, H.D. and Olsen, P.E., 1991, Triassic vertebrates of Gondwanan aspect from the Richmond basin of Virginia: Science, v. 249, p. 1020-1023.

Textoris, D.A. and Gore, P.J.W., in press, Upper Triassic organic-rich lake, Sanford sub-basin, North Carolina, USA, *in* Gierlowski-Kordesch, E. and Kelts, K., eds., Global Geological Record of Lake Basins: Cambridge, Cambridge University Press.

Tozur, E.T., 1984, The Trias and its ammonoids: The evolution of a time scale: Geological Survey of Canada Miscellaneous Report, no. 35, 169 p.

Traverse, A., 1986, Palynology of the Deep River basin, North Carolina, *in* Textoris, D.A., ed., Society of Economic Paleontologists and Mineralogists Field Guidebook, Third Annual Midyear Meeting, Raleigh, NC, p. 66-71.

Traverse, A., 1987, Pollen and spores date origin of rift basins from Texas to Nova Scotia as early Late Triassic: Science, v. 236, p. 1469-1472.

Ward, L.F., 1900, Status of the Mesozoic floras of the United States: United States Geological Survey Annual Report, no. 20, part II, p. 213-430.

Warrington, G., 1970, The stratigraphy and palaeontology of the 'Keuper' Series of the central Midlands of England: Quarterly Journal of the Geological Society of London, v. 126, p. 183-223.

Williston, S.W., 1904, Notice of some new reptiles from the Upper Trias of Wyoming: Journal of Geology, v. 12, p. 688-697.

Woodworth, J.B., 1902, The Atlantic Coast Triassic coal field in North Carolina: United States Geological Survey Annual Report, no. 22, part III, p. 25-53.

SOUTHEASTERN GEOLOGY
V. 33, No.4, October 1993, p. 195-218

THE PEN BRANCH FAULT: DOCUMENTATION OF LATE CRETACEOUS-TERTIARY FAULTING IN THE COASTAL PLAIN OF SOUTH CAROLINA

D. S. SNIPES[1], W. C. FALLAW[2], VAN PRICE, Jr.[3], and R. J. CUMBEST[3]

[1]*Department of Earth Sciences*
Clemson University
Clemson, SC 29634-1908

[2]*Department of Geology*
Furman University
Greenville, SC 29613

[3]*Environmental Monitoring Section*
Environmental Protection Department
Westinghouse Savannah River Company
Aiken, SC 29808

ABSTRACT

The Late Cretaceous-Tertiary Pen Branch Fault has been mapped for at least 15 mi (24 km) across the central portion of the Savannah River Site (SRS), South Carolina. Drill cores or geophysical logs from 57 wells located on or in the immediate proximity of the Savannah River Site provide detailed control for structural surface, isopach, and cross section constructions. Twenty-four cores sampled crystalline rocks below the sub-Cretaceous unconformity, 12 cores sampled early Mesozoic conglomerates or red beds and the remainder sampled Late Cretaceous strata. The subsurface geologic control is augmented by seismic reflection data (Chapman and Di Stefano, 1989). Based on these data the average strike of the fault is N 55° E with minor variations, and it closely follows the trend and is coincident with the northwest border of the early Mesozoic Dunbarton extensional basin.

Stratigraphic relationships and seismic studies indicate that the Pen Branch fault is a sub-vertical growth fault with down-to-the-northwest movement sense. Near the center of SRS, the thickness of Upper Cretaceous clastic strata is about 670 ft (203 m) on the down-thrown side, in contrast to 610 ft (185 m) on the up-thrown side. The throw decreases in successively younger beds from 80 - 100 ft (24 - 30 m) at the base of the Late Cretaceous Cape Fear Formation to 30 ft (9 m) at the top of the late Eocene Dry Branch Formation. These relationships yield estimated slip rates from 0 to 1.5 m/m.y. with an average of about 0.4 m/m.y. over the last 85 m.y.

The down-to-the-northwest movement sense for the Pen Branch fault is intriguing in that early Mesozoic deposition of fluvial sequences in the Dunbarton basin indicates that the basin surface must have been lower than the erosional surface of the crystalline terrain to the northwest. However, this paleoerosional surface is presently about 80 - 100 ft (24 - 30 m) below the basin surface at the location of the fault. This relationship has led to speculation that the Pen Branch fault is a Dunbarton basin border fault reactivated in a reverse sense due to compressional stresses or that is an antithetic basin fault.

INTRODUCTION

Stratigraphic sequences in the Coastal Plain of South Carolina and Georgia form a southeastward-thickening wedge of poorly consolidated sediments that onlap the crystalline basement lithologies of the Piedmont. In

Figure 1. Location of the Savannah River Site, S.C.

updip localities Late Cretaceous clastic sequences unconformably overlie terrigenous early Mesozoic clastic basin infill and crystalline basement rocks. The presence of this easily recognized unconformity and the relatively undisturbed nature of the Coastal Plain sequences make Coastal Plain geologic studies potentially very useful in providing time - stratigraphic constraints on post-Late Cretaceous tectonism of the southeast U.S. Atlantic margin. Resolution of the influences of basement structures on primary depositional features and deformation of Coastal Plain sequences can provide insights on both syn- and post-depositional basement movements.

It has long been recognized that a "layer cake" model for Coastal Plain structure and stratigraphy is an oversimplification and that basement-involved faulting is present in several localities (see numerous references in next section). Recent studies of Coastal Plain geology on and in the vicinity of the Savannah River Site, SC have documented features of a fault in Coastal Plain sequences that is closely related to early Mesozoic extensional basement structure. This fault, the Pen Branch Fault, is named for a prominent stream tributary of the Savannah River situated near the fault (Snipes and others, 1992). This paper presents subsurface geologic data which document the location and stratigraphic relationships associated with the Pen Branch Fault.

Geologic Setting

The Savannah River Site encompasses approximately 300 sq mi (780 sq km) in the central Savannah River area and at its nearest point is about 20 mi (30 km) southeast of the Piedmont physiographic province (Fig. 1). Coastal Plain sedimentary sequences which underlie SRS form a wedge of Cretaceous and Cenozoic sediments that thicken from about 700 ft (210 m) in the northwest to about 1400 ft (430 m) at the southeastern boundary. Regional dip is to the southeast and decreases upward from 48 ft/mi (9 m/km) at the base of the section to 15 ft/mi (3 m/km) at the top of middle Eocene beds.

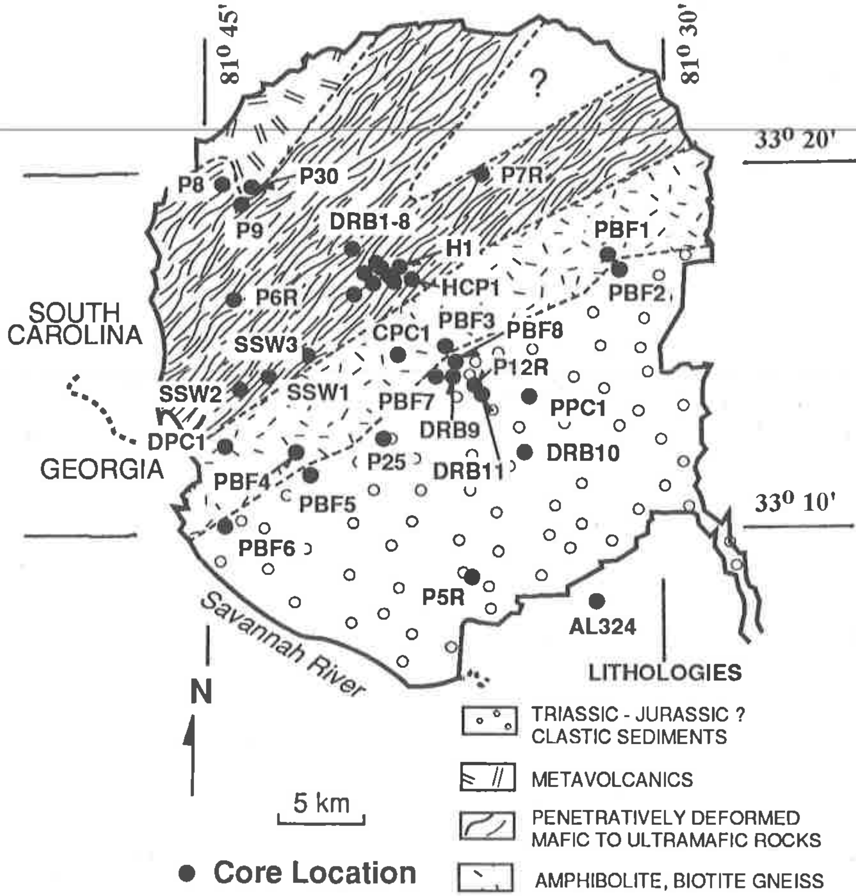


Figure 2. Sub-Cretaceous geologic map of SRS with well locations used as control indicated (adapted from Cumbest and others, 1992).

Basement Crystalline Rocks and Dunbarton Basin

The base of the Coastal Plain section is a sub-Cretaceous unconformity which separates Coastal Plain sedimentary sequences from underlying crystalline rocks and early Mesozoic basin infill. The crystalline rocks occur in northwestern portions of SRS and they consist primarily of greenschist facies metavolcanic rocks and amphibolite facies schists and gneisses (Fig. 2). However, southeastern portions of SRS are underlain by terrigenous clastic sequences associated with the early Mesozoic Dunbarton extensional basin which is the principal structural feature of the basement beneath SRS (Cumbest and others, 1992). The basin sediments are red beds and conglomerates which we believe are Karnian in age based on palynological studies by Traverse (1987) of sediments from other extensional basins in North Carolina and Georgia.

The Dunbarton basin (Siple, 1967; Marine and Siple, 1974; Cumbest and others, 1992) is one of several early Mesozoic depositional basins beneath the Coastal Plain in South Carolina and Georgia (Fig. 3). The relationship of the Dunbarton basin with the Riddleville basin to the southwest and the south Georgia Rift/ Summerville basin to the south is uncertain. Based on aeromagnetic data, Daniels and others (1983) suggested that the Summerville basin is an extension of the south Georgia Rift and that the Dunbarton basin connects to the larger Summerville basin to the southeast. However, drill core sampling of basement to

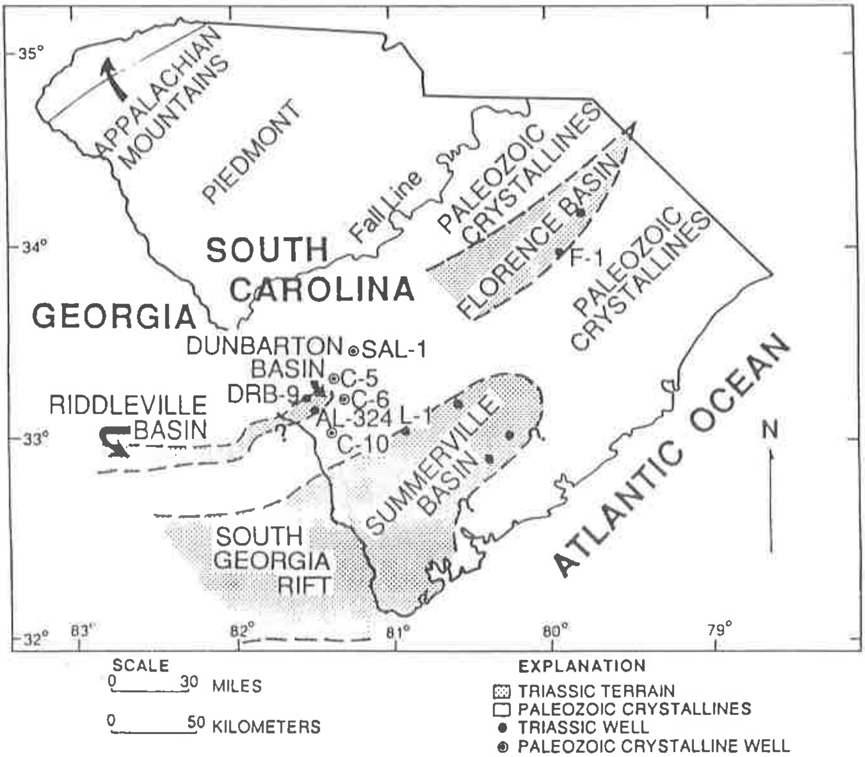


Figure 3. Location of Triassic basins beneath Coastal Plain sediments in South Carolina and southeastern Georgia. The Florence and Summerville basins are from Steele and Colquhoun (1985). The South Georgia Rift and Riddleville basin are after Daniels and others (1983). The postulated connection between the Riddleville and Dunbarton basin is after Daniels and others (1983). Wells F-1 and L-1 are from W.T. Griffin (S.C. Water Resources Commission, personal communication, 1989). Well Al-324 is from D. C. Prowell (U.S. Geological Survey, personal communication, 1990). Unnumbered wells are from Steele and Colquhoun (1985). Well SAL-1 is from Speer (1982). Wells C-5 and C-6 are from K. Sargent (1990, personal communication). Well C-10 and DRB-9 are from the present investigation.

the southeast of the Dunbarton basin (Fig. 3) does not support this interpretation. Wells C-6, C-7, and C-10 record Late Cretaceous clastic sequences nonconformably above crystalline rocks and the absence of early Mesozoic sediments (Robert Logan, personal communication). Well C-6 encountered gneissic basement at the elevation of -1123 ft (-342 m), C-7 encountered saprolitic schist at -1,146 ft (-349 m), and calcalkaline granite was encountered in C-10 at an elevation of -1,480 ft (-451 m). A whole rock Rb-Sr model age determined from this granite brackets the age between 440 - 550 Ma (Kish, 1992). Based on this evidence it appears that the Dunbarton basin is isolated from the larger rift basins to the southeast.

The Dunbarton basin is approximately 30 mi (50 km) long and 5,500 ft (1,700 m) deep, based on simultaneous gravity and magnetic modeling (Cumbest and others, 1992). The northern boundary of the basin has been accurately located based on seismic reflection and drill core data. At least one well, DRB-9 (Fig. 4), penetrated the basin sediments and encountered crystalline rock (Marine and Siple, 1974). The southeastern boundary of the basin is not as precisely located; however, based on an ongoing investigation by the writers, it is probably the Martin fault of Oldham (1981) and Logan and Euler (1989). This boundary is constrained by only two wells, Al-324, which bottomed in early Mesozoic red beds, and C-7,

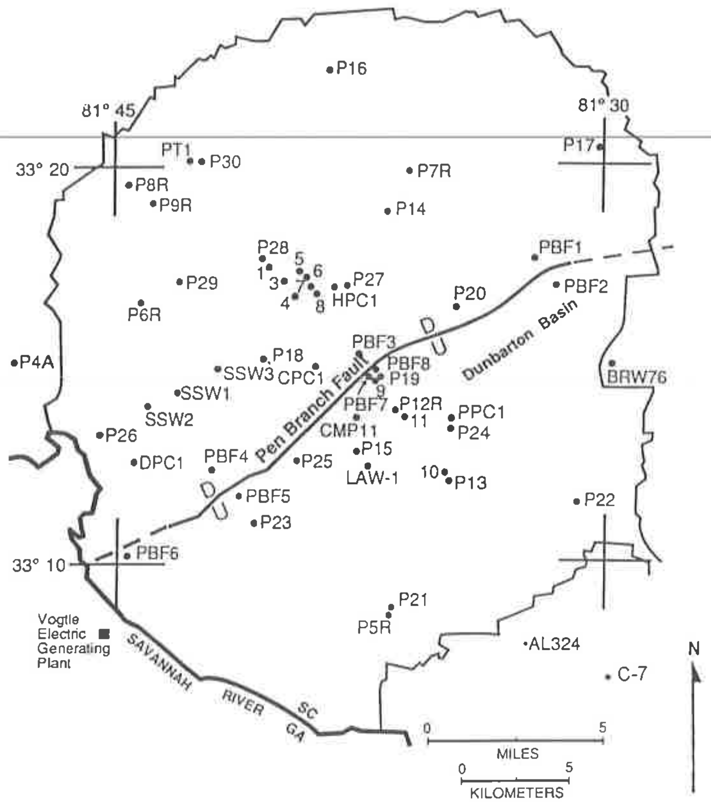


Figure 4. Locations of drill core used as geologic control for this study. P4A is from Siple (1967). Al-324 is from Prowell (personal communication, 1990).

which bottomed in Paleozoic schist. Gravity and magnetic modeling indicated that the basin may be asymmetric in cross section with a steeply dipping northwestern boundary (Cumbest and others, 1992).

Coastal Plain Stratigraphy

Coastal Plain stratigraphy in the central Savannah River area is the focus of many previous and ongoing studies (for instance Sloan, 1908; Cooke, 1936; LaMoreaux, 1946a,b; Cooke and MacNeil, 1952; Siple, 1967; Marine and Siple, 1974; Nystrom and Willoughby, 1982; Colquhoun and others, 1983; Prowell and others, 1985; Huddlestun and Hetrick, 1985; Steele, 1985; Fallaw and Price, 1992). The stratigraphic section used at the time of this study is in Table 1, although some recent modifications have been presented in Fallaw and Price (1992).

The lower part of the Coastal Plain section at SRS consists of Upper Cretaceous strata divided into, from older to younger, the Cape Fear, Middendorf, Black Creek, and Peedee formations, with the latter three constituting the Lumbee Group. The Cape Fear, which lies on igneous and metamorphic rocks of the Appalachian orogen and Triassic red beds, is composed of poorly sorted, silty and clayey quartz sands and interbedded clays. The sands are commonly medium and coarse and are arkosic and pebbly in places. The Cape Fear Formation is more indurated than the other Cretaceous formations.

The Middendorf Formation consists mostly of medium and coarse quartz sand. Sorting is generally moderate to good. Pebbly zones are common within the sand, and clay clasts occur in places. Some parts of the unit are feldspathic, and micaceous and lignitic

Table 1. Lithology and ages of the stratigraphic units recognized in 1989 at SRS. For revisions to this stratigraphic column, see Fallaw and Price (1992).

AGE		UNIT		LITHOLOGY
TERTIARY	Miocene(?)	"upland unit"		Clayey, silty sands, conglomerates, pebbly sands, and clays; clay clasts common
	Late Eocene	Barnwell Group	Tobacco Road Sand	Red, purple, and orange, poorly to well-sorted sand and clayey sand with abundant clay laminae
			Dry Branch Formation	Tan, yellow, and orange, poorly to well-sorted sand with tan and gray clay layers near base; calcareous sands and clays and limestone in lower part downdip
			Clinchfield Formation	Biomoldic limestone, calcareous sand and clay, and tan and yellow sand
	Middle Eocene	Santee Limestone and correlatives		Micritic, calcarenitic, shelly limestone, and calcareous sands; interbedded yellow and tan sands and clays; green clay and glauconitic sand near base
	Early Eocene	Congaree Formation		Yellow, orange, tan, and greenish-gray, fine to coarse, well-sorted sand; thin clay laminae common
	Paleocene	Williamsburg Formation		Light gray, silty sand interbedded with gray clay
		Ellenton Formation		Black and gray, lignitic, pyritic sand and interbedded clays with silt and sand laminae
LATE CRETACEOUS	Maestrichtian	Lumbee Group	Peedee Formation	Gray and tan, slightly to moderately clayey sand; gray red, purple, and orange clays common in upper part
	Campanian		Black Creek Formation	Tan and light to dark gray sand; dark clays common in middle and oxidized clays at top
	Santonian		Middendorf Formation	Tan and gray, slightly to moderately clayey sand; gray red, and purple clays near top
		Cape Fear Formation		Gray, clayey sand with some conglomerates, and sandy clay; moderately to well indurated
LATE TRIASSIC		Newark Supergroup		Boulder conglomerate, red, arkosic, poorly sorted sandstone and red shale
PALEOZOIC and CRYPTOZOIC(?)		"crystallines"		Biotite gneiss, mica schist, amphibolite, chlorite schist, and granitoid rocks

zones occur. Over much of SRS, a kaolinitic clay or a clay-and-interbedded-sand zone up to 50 ft (15 m) thick forms the top of the unit. In most wells the contact between the Middendorf and the underlying Cape Fear is sharp and often marked by a pebbly zone. The younger unit has cleaner sands and lacks the repetitive sand-clay sequences of the Cape Fear. It contains less feldspar, is not as well indurated, and the color is less variable. A sharp increase in resistivity going upward across the contact occurs on geophysical logs.

The Black Creek Formation consists of quartz sands, silts, and clays. It is generally darker, more micaceous, and more lignitic than the other Cretaceous units. The lower part of the formation consists of fine to coarse sands with moderate to poor sorting. The sand is micaceous and becomes lignitic in the central and southeastern parts of SRS. In the central and downdip part of SRS, a southeastwardly-thickening wedge of dark, fissile, lignitic, pyritic, micaceous clay with dark, interbedded sands and silts occurs in the middle and upper parts of the formation. The topmost part of the formation consists mostly of sands. In many wells sands of the Black Creek lie on oxidized clay beds at the top of the Middendorf. Where the clays are missing, it is difficult to pick a contact, but a pebbly zone occurs in some wells. In general, the Black Creek contains more dark clays, lignite, and muscovite than the Middendorf.

An updip facies of the Peedee Formation overlying the Black Creek is made up of quartz sands and mostly oxidized, kaolinitic clays. The sediments in the lower part of the formation are poorly to well-sorted, fine to coarse quartz sand and silty sand, in places very micaceous. The upper part of the Peedee in most places at the Site is oxidized, kaolinitic clay interbedded with sands in places. In most wells, the basal contact can be placed at the bottom of a coarse sand below which are sands interbedded with dark clays and above which are sands interbedded with variegated clays.

The Paleocene Ellenton Formation in most wells at SRS is composed of gray, poorly and moderately sorted, micaceous, silty and clayey quartz sands and pebbly sands with interbedded, thick, dark gray clays. The sands are locally feldspathic, especially those near the base, and iron sulfides and lignite are common in the darker parts of the section. The clays are fissile in places, micaceous, lignitic, and contain silt and fine sand laminae. Iron sulfides are common. Both sands and clays are glauconitic in places, especially in the southeastern part of the Site. Basal sands, often pebbly, of the Ellenton lie with a sharp, unconformable contact on oxidized clays of the Peedee in most wells at SRS. In general, the Ellenton has more feldspar and iron sulfide than the Peedee and is darker and sorting is poorer. The clays of the Ellenton are more fissile than those of the Peedee. Where the oxidized clay at the top of the Peedee is missing, it is difficult to pick the contact. Where the Ellenton in these places is better sorted and lighter in color than is typical, it is similar to moderately to well-sorted sands in the Peedee. In some cores, the sands of the Peedee are micaceous, poorly sorted, and dark, similar to typical Ellenton sands. A pebbly layer occurs in the base of the Ellenton in some of the problem wells.

An updip facies of the upper Paleocene Williamsburg Formation consists of silty, micaceous, medium to coarse quartz sands and pebbly sands interbedded with kaolinitic clays. Sorting in the sands is generally poor, but well-sorted sands are present. Dark, micaceous, lignitic sands also occur. The clays are oxidized in some places but dark in others. In most wells basal, light colored, micaceous Williamsburg sands lie on dark clays, glauconitic in places, of the Ellenton. Williamsburg sands are usually lighter in color than Ellenton sands, and the unit contains less lignite, iron sulfide, and glauconite.

Lying above the Paleocene strata is the Congaree Formation, which contains fossil assemblages indicating an early Eocene age, at least for most of the unit. The lower part of the Congaree is moderately to well-sorted, fine to coarse quartz sand with clays a few feet thick in the middle and at the top in places. Glauconite, muscovite, and iron sulfide are common accessories. The lower part of the Congaree, as the term is used in this paper, correlates biostratigraphically with the Fishburne Formation, a downdip carbonate (Gohn and others, 1983). The upper Congaree consists of moderately and well-sorted, fine to coarse quartz sands. Thin clay laminae are present in places. In the northwestern part of SRS, the Congaree overlies the dark clays and sands of the Ellenton Formation. In the southeast, the underlying unit is the Williamsburg. The basal contact is sharp in both areas. In general, going upward across the Ellenton contact, the sands become cleaner, iron sulfide and lignite content decreases, colors become lighter, and clay bed thickness decreases. Going upward across the Williamsburg/Congaree contact, sands become cleaner, glauconite increases, and clay bed thickness decreases. On geophysical logs, an increase in resistivity and a decrease in gamma ray count occur going upward across the Congaree/Paleocene contact.

Above the Congaree is the middle Eocene Santee Limestone and correlatives. In the center of SRS, it is composed of light-colored calcarenite and calcilutite with indurated calcareous nodules. Indurated, moldic limestone also occurs in many SRS cores. An updip facies consists of quartz sands, silts and clays. Typically, the sands of the siliciclastic facies are finer than the sands above and below, contain more heavy minerals, and are more likely to contain glauconite. Clay beds and laminae are more abundant than in underlying and overlying units. In the southeastern part of the Site, light-colored Santee carbonates interfinger with and grade into gray and pale green, clayey, laminated calcilutite, calcarenite, and calcareous silt and clay. Where the siliciclastic facies overlies the Congaree, going upward in the section colors become darker, grain size decreases, sorting becomes poorer, green clays become more common, and heavy minerals become abundant. A pebbly zone occurs at the base in places. On geophysical logs, the gamma ray count for the Santee is usually higher than for overlying and underlying units. In general, resistivities are low in the Santee, especially in comparison with the overlying unit.

The Clinchfield, Dry Branch, and Tobacco Road formations constitute the upper Eocene Barnwell Group at SRS. The Clinchfield consists mostly of quartz sand and clay, calcareous in places, and carbonates. The siliciclastics are poorly to well-sorted, and fine to coarse. The carbonate is an indurated, bioclastic and biomoldic, glauconitic limestone in some places, and in others a calcareous sand and calcarenite. In places the lower contact of the Clinchfield is marked by a change from calcareous sediments of the Santee to poorly to well-sorted quartz sand. In general, where Clinchfield siliciclastics overlie Santee siliciclastics, above the contact the sediments are coarser, there are fewer heavy minerals, and manganese-stained sediments are more common.

Part of the Dry Branch Formation consists of the calcareous Griffins Landing Member, composed of calcilutite, calcarenite, bioclastic and biomoldic limestone, calcareous sand, and shelly, calcareous clay. At SRS, the Griffins Landing is less glauconitic than the carbonates of the underlying Clinchfield and Santee. A pebbly layer at the contact occurs in some cores. The remainder of the Dry Branch Formation within SRS is made up of the Irwinton Sand Member. It is composed of moderately sorted quartz sand, with interlaminated and interbedded clays, typically tan, abundant in places. Pebbly layers and zones rich in clay clasts occur. Irwinton sands are generally coarser than those of the underlying Santee siliciclastics, and glauconite and heavy minerals are less abundant. Tan clays are more common above and green and gray clays are more common below the contact.

The Tobacco Road Sand consists of moderately to poorly sorted quartz sands and clayey quartz sands. In general, the sands of the Tobacco Road are muddier, more micaceous, and more highly colored than those of the underlying Dry Branch. The base of the Tobacco Road is marked in places by a coarse layer that contains flat quartz pebbles. A peak in gamma radiation occurs at the contact on many geophysical logs. The upper surface of the unit is irregular because of incision that preceded deposition of the overlying "upland" unit.

The "upland" is an informal term applied to Miocene(?) deposits that occur at higher elevations in many places in the southwestern South Carolina Coastal Plain. The sediments are poorly-sorted, clayey and silty, fine to coarse sands, with lenses and layers of gravels, pebbly sands, and oxidized, massive clays. Clay clasts are abundant. Cross-bedding is prominent in places, and muscovite and flecks of weathered feldspar are locally abundant. In general, going upward across the base of the "upland", sorting becomes poorer, weathered feldspar grains become larger and more common, clay beds become more abundant and thicker, sands become more argillaceous and indurated, pebbles are larger, and, in places, muscovite increases. The lower surface of the unit is very irregular because of erosion of underlying deposits.

Coastal Plain Faulting

The distribution of faults of Cretaceous and Cenozoic age east of the Mississippi river has been compiled by Prowell (1983). Most of this compilation was based on field observation of fault exposures. Out of about 130 localities where one or more faults were reported, 80 were located along the Atlantic margin. According to Prowell (1988) most of the post - Jurassic faults in the southeastern Atlantic Coastal Plain are northeast-trending reverse faults (see for instance Wentworth and Mergner-Keefer, 1983; Zupan and Abbott, 1975). However, many workers have reported Coastal Plain faulting involving other than reverse movement sense. Brown and others (1972) reported displacements producing half-grabens in the Coastal Plain of North Carolina and Virginia and there are many reported occurrences of normal faulting in Coastal Plain sequences (Cramer and Arden, 1978; Reinhardt and others, 1984; Howell and Zupan, 1974; Inden and Zupan, 1975). Many studies have concluded that faults with both normal and reverse movement sense have been active in the Coastal Plain since Late Cretaceous to Tertiary resulting in displacements of a few tens of meters (Higgins and others, 1978; Zoback and others, 1978; Behrendt and others, 1981; Behrendt and others, 1983; Hamilton and others, 1983; Talwani, 1986; Shedlock and Harding, 1988; Schilt and others, 1983).

Several authors have reported oblique-slip or strike-slip movement on post-Cretaceous Coastal Plain faults. Prowell and O'Conner (1978) and Bramlett and others (1982) have described, from outcrop and subsurface information, Late Cretaceous and post-Eocene oblique-slip reverse faulting near Augusta, Georgia. Seismic profiling has also identified faulting of Coastal Plain sequences in places inaccesable to field study. Although only vertical offset can be resolved easily on the seismic sections, Behrendt and Yaun (1986) inferred strike-slip movement based on the orientations of fault planes relative to the regional state of stress and on the occurrence of en echelon fault strands connected by normal faults. Although the movement sense reported for post-Cretaceous faults on the U. S. Atlantic margin spans the full range from reverse to strike-slip to normal, many of these faults show decreasing offset up section in Coastal Plain sequences indicating that they are growth faults.

A characteristic of many Coastal Plain faults is their close association with early Mesozoic extensional basins. Shomo (1980, 1982) studied Coastal Plain faulting associated with a sub-Coastal Plain extensional basin in Virginia. She concluded that high-angle reverse faults which offset Cretaceous and Tertiary sediments were caused by reactivation of Triassic normal faults. Based on seismic data Behrendt and others (1988) documented late Cenozoic movement above several Triassic basins in South Carolina. The close association naturally leads to speculation about the relationship of basement controlled faulting to Coastal Plain structures.

GEOLOGICAL AND GEOPHYSICAL INVESTIGATIONS

Interpretations of the Coastal Plain stratigraphy and structure at SRS are based largely on well control augmented by seismic reflection data (Chapman and Di Stefano, 1989). A total of 57 wells are employed in this investigation. The lithologic data on 46 wells are from the present study, information on three of the wells was obtained from the literature (Siple, 1967; Marine and Siple, 1974) and data from the remaining eight wells are from unpublished SRS reports. Geophysical logs from 46 wells

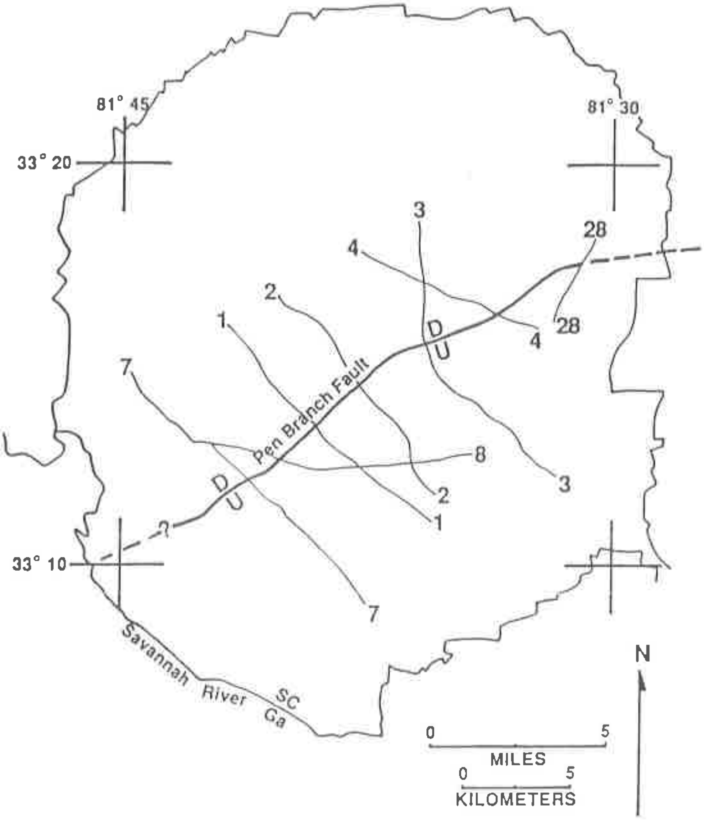


Figure 5. Locations of seismic survey lines in relation to the Pen Branch fault (after Chapman and Di Stefano, 1989).

were studied, cores were examined from 36 wells, and paleontological analyses were made on 19 wells. For each well geophysical logs included two or more of the following: natural gamma, spontaneous potential and resistivity. Eight well sites were located and drilled specifically to obtain additional control on the Pen Branch Fault. In the summer of 1989, these wells were cored and geophysically logged from the ground surface through the sub-Cretaceous unconformity into crystalline basement (3 wells) or early Mesozoic basin sedimentary fill (5 wells).

In addition to subsurface investigations, studies of topographic maps (1:24000) and geological field reconnaissance were utilized to investigate the possibility of recent fault movement. These investigations revealed no surface expression of the Pen Branch Fault. During this investigation a Quaternary light tan soil horizon, which has a thickness of 10 - 20 ft (3 - 6 m) was examined in railroad cuts which occur above the projected trend of the fault. This soil horizon reveals no detectable offset, indicating that there has been no recent Pen Branch Fault activity.

Reflection Seismic Control

Conoco, Inc. conducted an extensive reflection seismic survey of the Savannah River Site in 1989 (Chapman and Di Stefano, 1989). Seven of the survey lines were located and oriented correctly to image the Pen Branch Fault (Figure 5). Two of the seismic lines are shown in Figures 6 and 7. The seismic data show distinctive laterally extensive events in the upper 0.5 sec. of the seismic records. This part of the section represents the Coastal Plain seismic stratigraphy and the regional dip can be

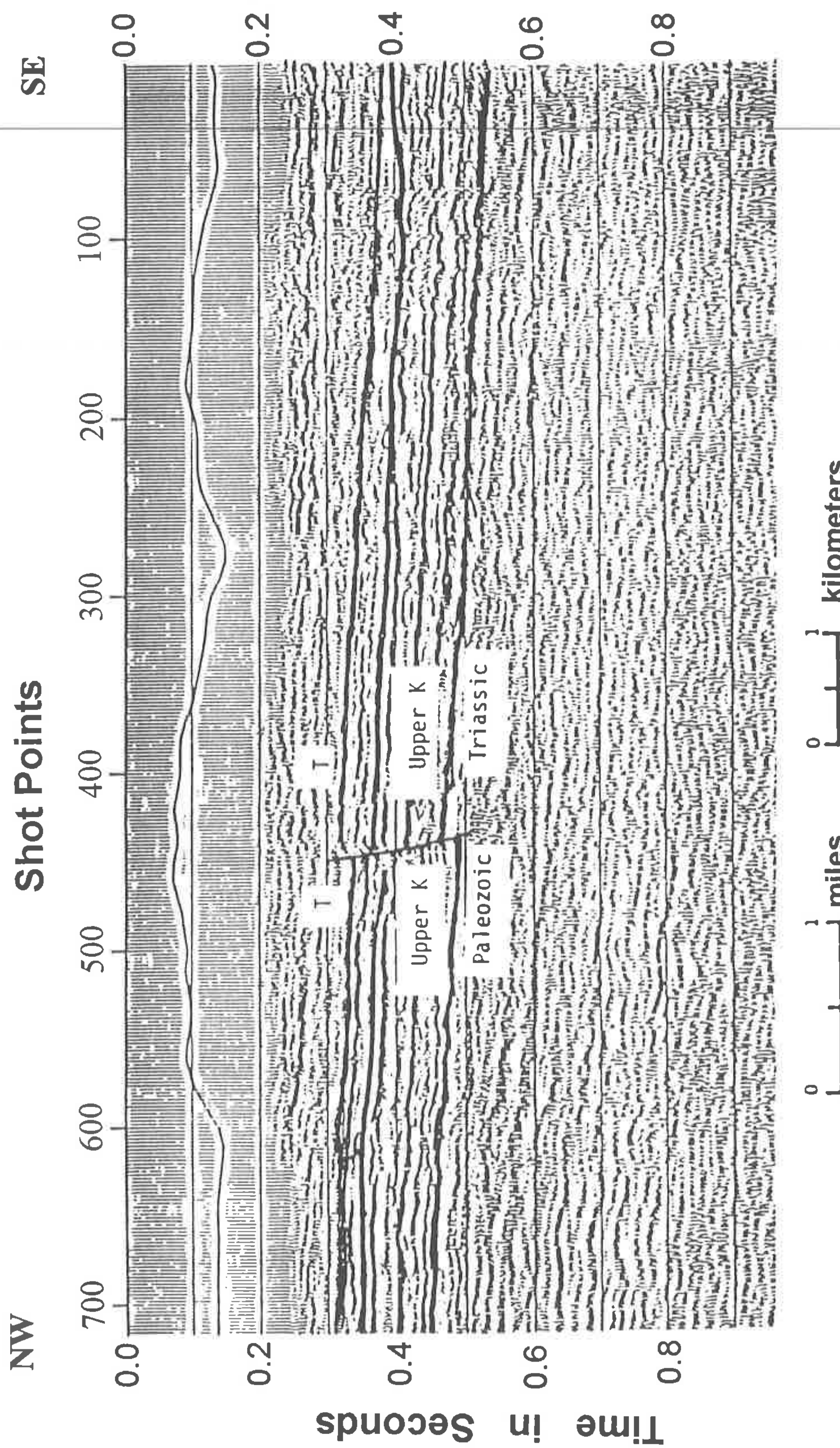


Figure 6. Migrated seismic reflection profile (line 7, Figure 5) showing offset of Upper Cretaceous reflectors and lateral changes in Tertiary events. The discontinuous nature of the Tertiary reflectors makes offset difficult to demonstrate. Based on these data the Pen Branch fault appears to dip steeply to the southeast.

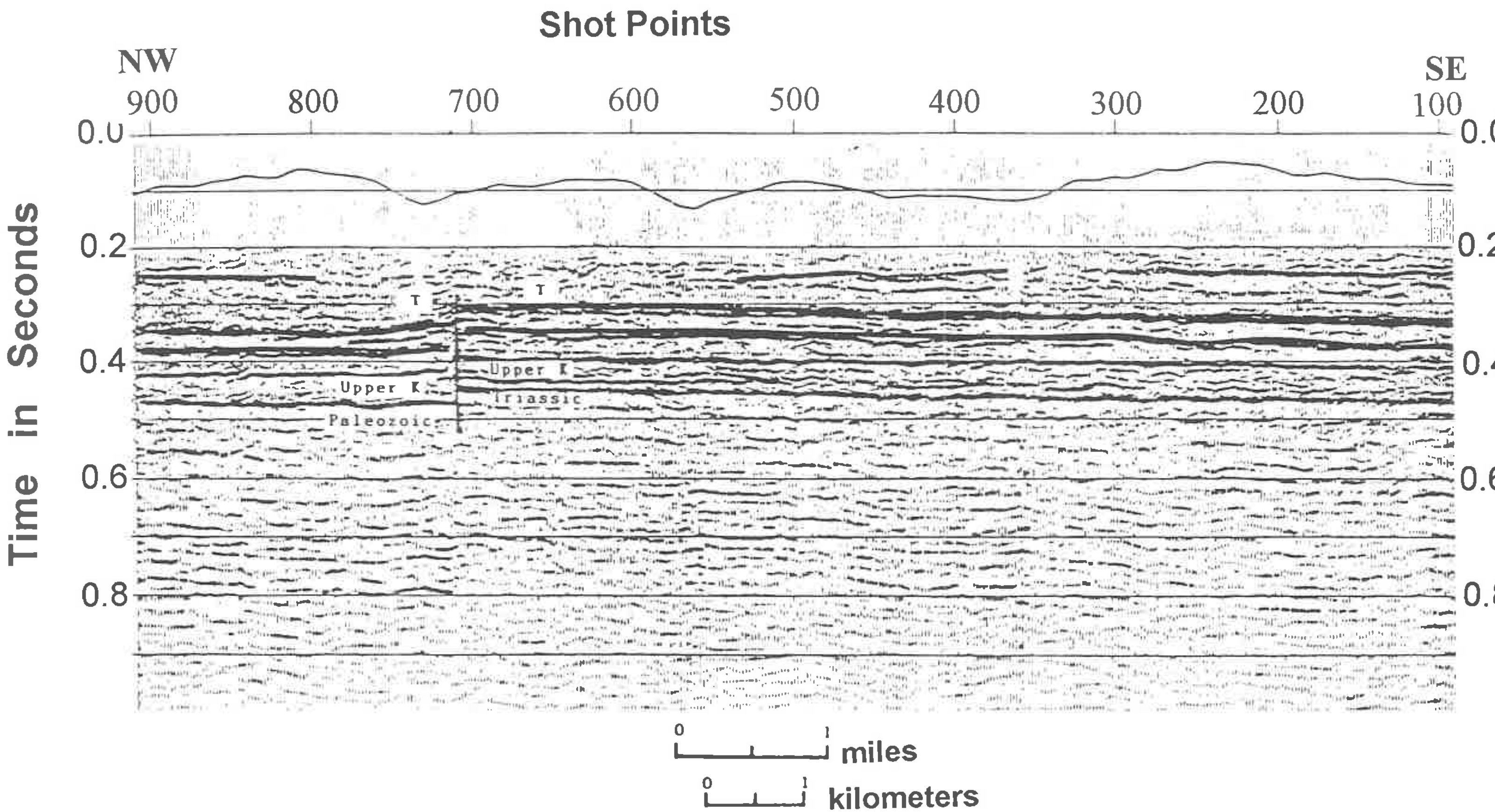


Figure 7. Migrated seismic reflection profile (line 8, Figure 5) showing offset of the Upper Cretaceous reflectors associated with the Pen Branch fault. Due to lateral changes offset of the Tertiary reflectors is difficult to demonstrate. Based on these data the dip of the Pen Branch fault appears to be vertical.

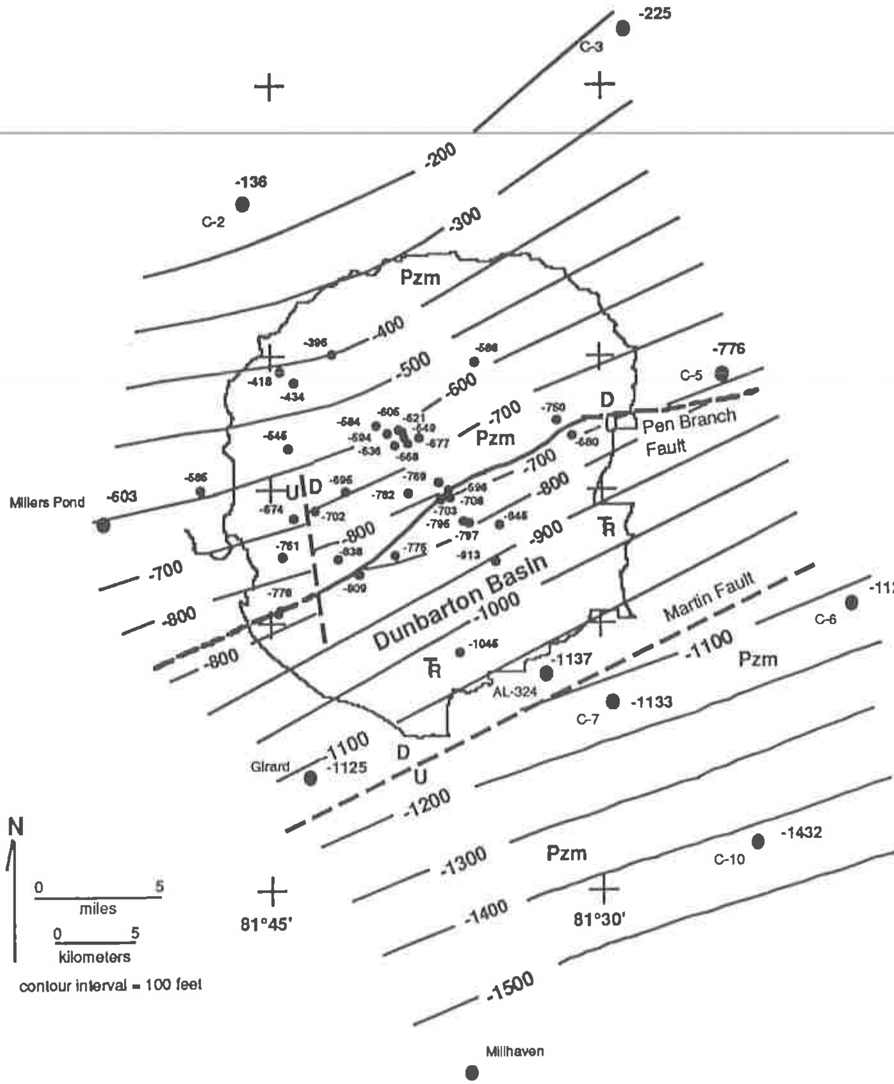


Figure 8. Structure contour map on the sub-Cretaceous unconformity. (Pzm = Paleozoic crystalline rocks; T_R = Triassic red beds and conglomerates; well cores marked by solid circles; "D" on downthrown block; "U" on upthrown block).

easily discerned. The lower most of the events can be correlated with the aid of drill core to the sub-Cretaceous unconformity at the base of the Coastal Plain section. Unfortunately the seismic data do not clearly image the basement.

A distinct offset of the event corresponding to the sub-Cretaceous unconformity with down to the northwest movement sense can be clearly seen in the seismic data. This offset represents an estimated throw at the top of basement of 100 ft (30 m). The seismic data also show offset in superposed reflection events that indicate that the fault is sub-vertical. However, the discontinuous nature of the upper reflectors make the orientation of the dip (i.e. northwest or southeast) and the extent of the offset in the upper part of the section uncertain.

Subsurface Geologic Control

Based on correlations of lithologic and

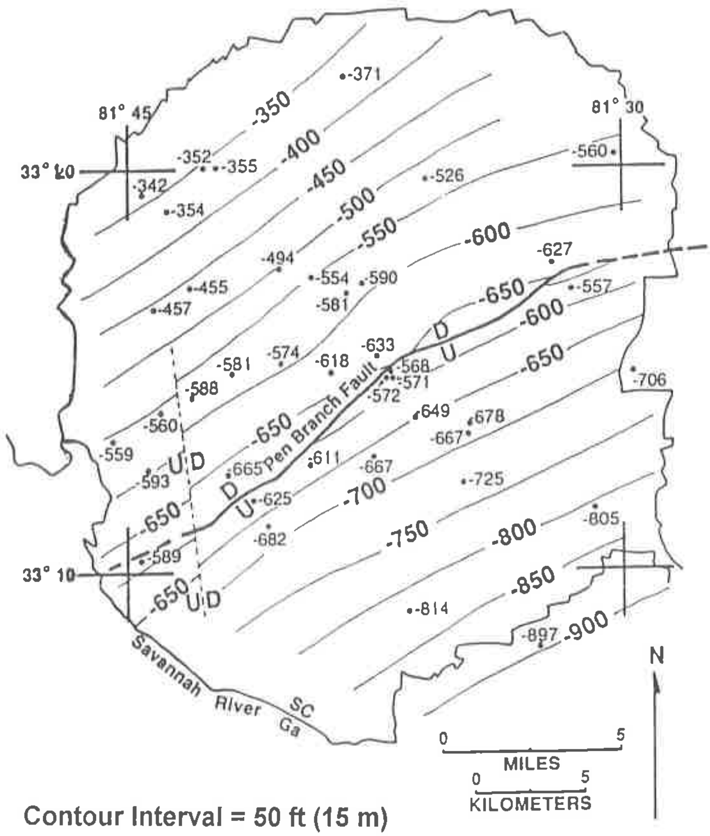


Figure 9. Structure contour map on the top of the Upper Cretaceous Cape Fear Formation.

geophysical logs, together with paleontological data, four structural contour maps, two isopach maps and two cross sections were constructed. Structure contour maps include the sub-Cretaceous unconformity, top of the Upper Cretaceous Cape Fear Formation, top of the Upper Cretaceous Lumbee Group, and top of the middle Eocene Santee Limestone and correlatives. The average dip of the sub-Cretaceous unconformity, as determined from the structure contour map (Figure 8) is 50 ft/mi (9 m/km) southeast. Based on control provided by the numerous wells adjacent to the fault the crystalline rocks of the basement on the northwest side of the fault can be tightly constrained to be 80 - 100 ft (24 - 30 m) lower than the early Mesozoic clastic sediments of the basement to the southeast. The top of the Late Cretaceous Cape Fear Formation (Figure 9) has a regional dip of 36 ft/mi (7 m/km) and shows similar amounts of offset as the sub-Cretaceous unconformity. These observations are consistent with offset estimates made for these markers from the seismic data.

The top of the Upper Cretaceous Lumbee Group (Upper Cretaceous - Paleocene unconformity; Figure 10) is more irregular than the surfaces mapped lower in the section and varies in dip from 18 ft/mi (3 m/km) to 40 ft/mi (8 m/km) with an average dip of about 27 ft/mi (5 m/km). The displacement at the top of the Upper Cretaceous - Paleocene unconformity varies from 30 ft (9 m) to 70 ft (21 m) with and average of approximately 40 ft (12 m). The average offset is one half that determined from the sub-Cretaceous unconformity and top of the Cape Fear Formation.

The top of the middle Eocene Santee

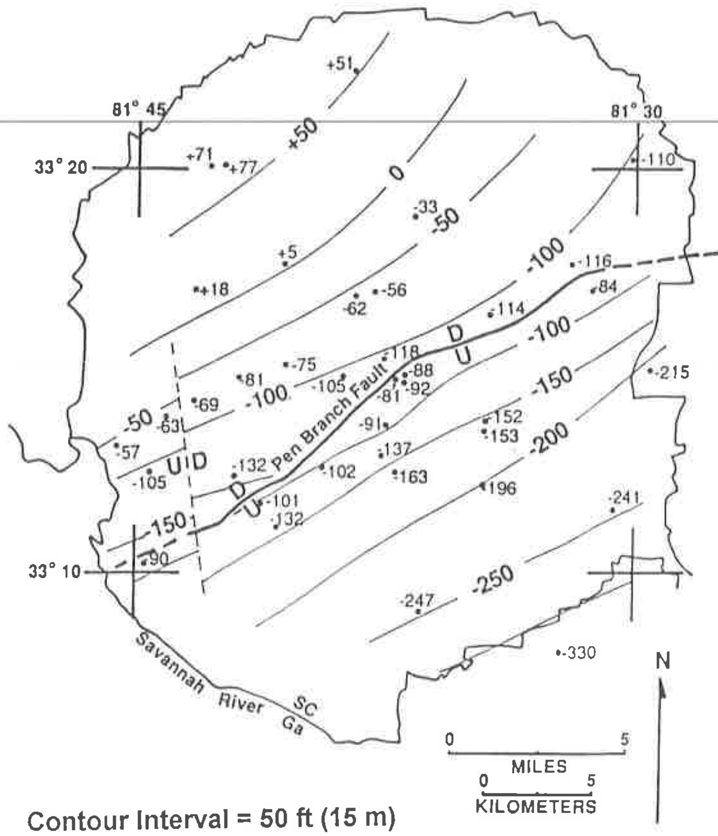


Figure 10. Structure contour map on the top of the Upper Cretaceous Lumbee Group.

Limestone and correlatives (middle Eocene - late Eocene unconformity; Figure 11) is also very irregular in dips ranging from 10 ft/mi (2 m/km) to 30 ft/mi (6 m/km) and an average of 18 ft/mi (3 m/km). The displacement of the unconformity varies along strike from 30 - 70 ft (9 - 21 m). This amount of offset is similar to that seen at the Upper Cretaceous - Paleocene unconformity.

The thickness of Upper Cretaceous units (Figure 12), including the Cape Fear Formation and the Lumbee Group, increases uniformly from 450 ft (140 m) near the northern boundary of SRS to about 800 ft (240 m) near the southern boundary with the exception of the areas adjacent to the Pen Branch Fault. In this location the Upper Cretaceous clastic sequences are 30 - 70 ft (9 - 21 m) thicker on the northwest side than the southeast side. The Tertiary strata between the top of the Upper Cretaceous and the top of the middle Eocene Santee Limestone and correlatives (Figure 13), including the Paleocene Ellenton and Williamsburg Formations, the middle and lower Eocene Congaree Formation and the middle Eocene Santee Limestone and correlatives, vary in thickness from 150 ft (46 m) to 300 ft (91 m) from northwest to southeast. Significant local variations in thickness of this interval make comparisons of thicknesses across the fault equivocal.

Cross section A-A' (Figure 14) was constructed across the central portion of SRS perpendicular to the strike of the fault. Several of the wells illustrated on this cross section are too close together to be drawn separately. The general southeastward thickening and increase in dips of the Coastal Plain sedimentary units are clearly discerned in addition to offset of the various stratigraphic tops. Well DRB-9 intersected the contact between the early Mesozoic

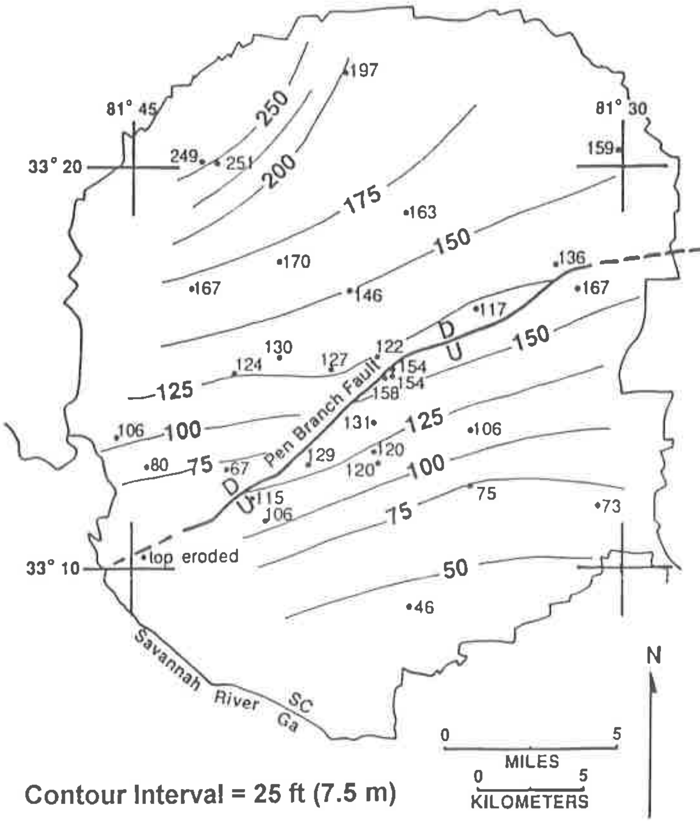


Figure 11. Structure contour map on the top of the middle Eocene Santee Limestone and correlatives.

clastic basin fill and crystalline rocks at an elevation of -2330 ft (-710 m). A specimen of core from just below the contact was a mylonitic granite indicating that this contact is probably both an unconformity and a fault. Since DRB-9 is the only well that intersects this surface the magnitude of the dip cannot be determined from the available well control. However, PBF-3 and PBF-8 constrain the location of the fault in higher portions of the section indicating that the fault probably dips to the southeast.

Cross section B-B' (Figure 15) is coincident with a short section of A-A' and illustrates the upper part of the section in this area with control provided by shallow wells E-24 and E-26. This cross section documents the amount of offset at the top of the upper Eocene Dry Branch Formation. Based on this cross section the throw at the top of the Dry Branch is about 30 ft (9 m).

DISCUSSION

Faulting History and Cause of Faulting

Offset of the sub-Cretaceous unconformity and Upper Cretaceous through lower upper Eocene stratigraphic units are demonstrated based on well control and reflection seismic data. This offset lies directly above and follows the trend of the northwestern border of the Dunbarton basin. Along this trend Coastal Plain sediments are thicker on the northwest side than they are on the southeast side, counter to the regional thickening to the southeast. These observations are interpreted as proof of the presence of a fault, named the Pen Branch Fault (Snipes and others, 1992). Pen Branch is close to the trend of the fault, and flows southeast to the Savannah River.

The fault can be traced at least 15 mi (24

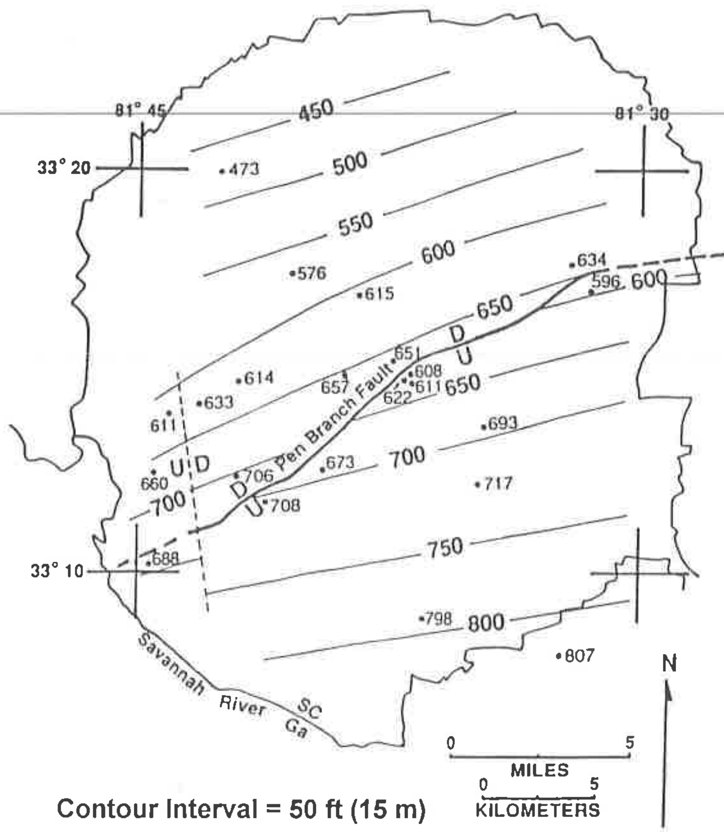


Figure 12. Isopach map of the Upper Cretaceous section.

km) across SRS. However, its extent to the southwest across the Savannah River and to the northeast is uncertain. Based on control provided by PBF-6, which is situated 1.4 mi (2.2 km) east of the Savannah River, the projected fault trace should pass about 2 mi (3 km) north of Vogtle Electric Generating Plant, Georgia.

Offset of the sub-Cretaceous unconformity and Coastal Plain stratigraphic markers all indicate down-to-the-northwest movement sense. This movement sense is consistent with thickening of stratigraphic units on the northwestern side indicating that the Pen Branch Fault is a growth fault. The sub-vertical nature of the fault and poor seismic data in the upper part of the section make the dip direction of the fault difficult to determine. However, based on wells near the fault trace and one well that penetrated the fault at depth, the fault probably ranges in dip from vertical to southeast. This conclusion is predicated upon the assumption that the fault in the Coastal Plain is a continuation of the fault that separates crystalline rocks from early Mesozoic clastic basin infill. We consider this assumption valid based on the coincident location of these two features and the fact that the surface that separates the basement from the Coastal Plain section (sub-Cretaceous unconformity) is also faulted. However, the high angle of the fault dip and uncertainty in its dip direction make assignment of a reverse or normal nomenclature difficult.

The increase of offset of stratigraphic markers and increase in thickening of stratigraphic units on the down-thrown side of the Pen Branch Fault indicate that fault movement exerted a control on deposition of Coastal Plain sedimentary sequences. Fault movement began at least as early as Late Cretaceous time (or

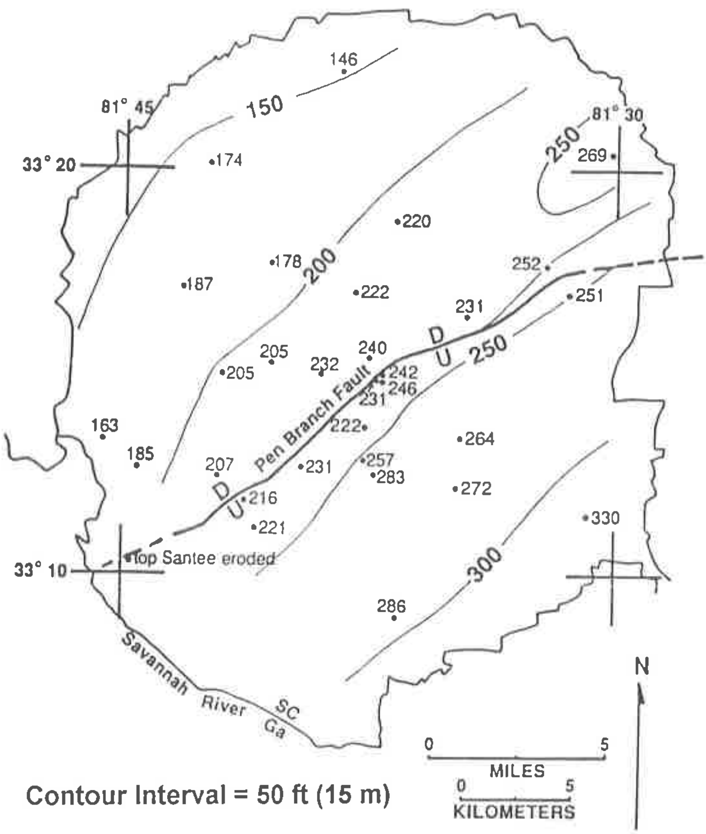


Figure 13. Isopach map of the Tertiary interval between the top of the Upper Cretaceous and the top of the middle Eocene Santee Limestone and correlatives.

earlier) and continued intermittently through the early part of late Eocene time. The youngest stratigraphic unit that can be shown to be offset is the Dry Branch Formation (Price, 1990), i.e., the lower part of the late Eocene Barnwell Group. Demonstration of offset in the overlying section is problematical due to the irregular nature of the erosional unconformity. Therefore, the question of displacement at the top of the upper Eocene section has not been resolved.

Estimated slip rates on the Pen Branch Fault from Late Cretaceous to Tertiary time, based on chronological ages of Coastal Plain stratigraphic units combined with corresponding throw at the top of these units (Table 2), vary from 0 to 1.5 m/m.y. The average is about 0.4 m/m.y. over the last 85 m.y. These results correspond closely with those of Prowell (1988) for Cretaceous and Cenozoic faults on the Atlantic Coastal margin.

The down-to-the-northwest movement sense exhibited by the Pen Branch Fault is opposite to that expected from faulting associated with development of a basin to the southeast. If movement of the basin-bounding fault in the basement is the cause of faulting in the Coastal Plain, this implies that the basement fault has been reactivated in a sense opposite to its original movement.

Chapman and Di Stefano (1989) suggested that the Pen Branch Fault was a reverse fault resulting from reactivation of a normal fault associated with the Triassic Dunbarton basin. They postulated that the cause of this activity was compression created by Late Cretaceous plate movement. This explanation is similar to the proposed by Wentworth and Mergner-

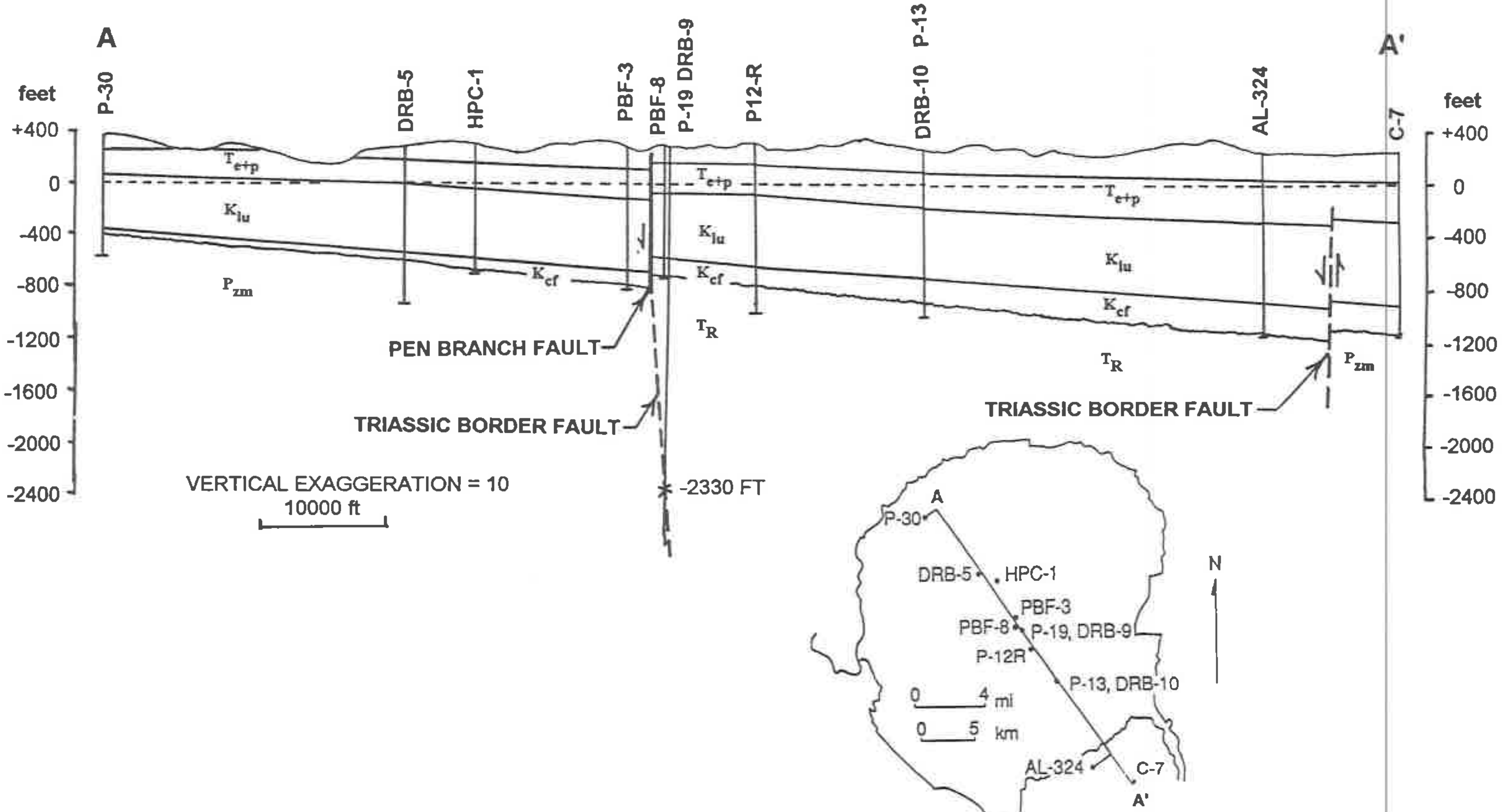


Figure 14. Northwest-southeast cross section across the central portion of the Savannah River Site. (T_{e+p} = Middle Eocene Santee Limestone and correlatives, middle (end lower?) Eocene Congaree Formation, upper Paleocene Williamsburg Formation and lower Paleocene Ellenton Member of Rhems Formation; K_{lu} = Upper Cretaceous Lumbee Group, Peedee, Black Creek and Middendorf Formations; K_{cf} = Upper Cretaceous Cape Fear Formation; T_R = Triassic red beds; P_{zm} = Paleozoic crystalline rocks).

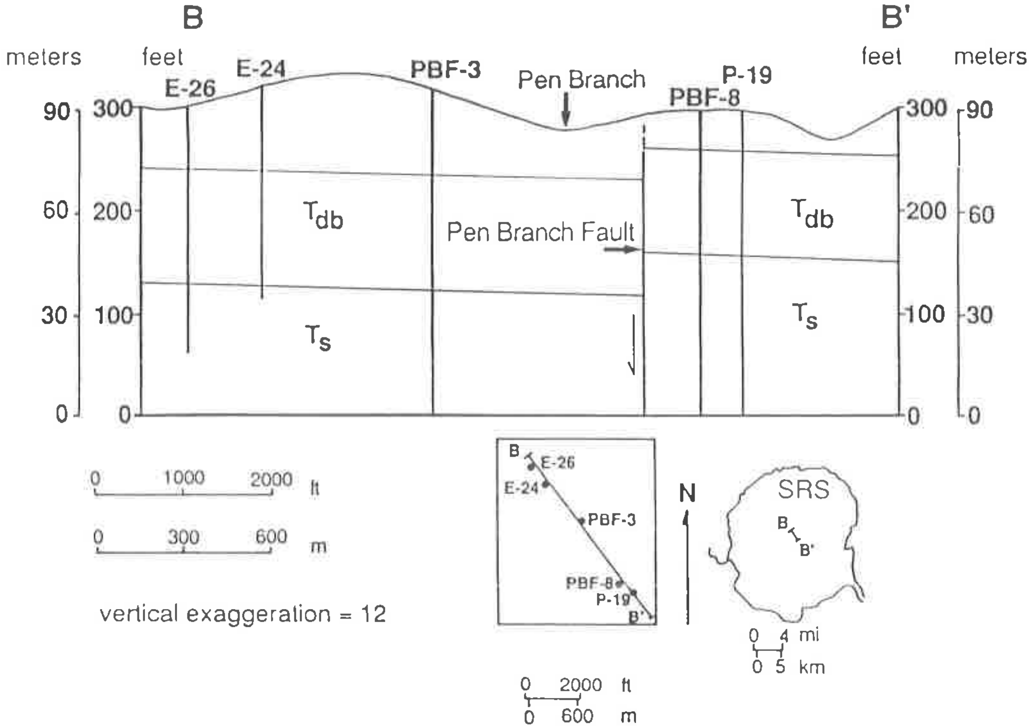


Figure 15. Detailed northwest-southeast cross section (B-B') showing displacement by the Pen Branch fault of the upper Eocene Dry Branch Formation (Tdb) and the middle Eocene Santee Limestone and correlatives (ts).

Table 2. Estimated slip rates on the Pen Branch fault during Late Cretaceous and Tertiary time. Geologic ages based on dinoflagellates, spores and pollen, calcareous nannofossils, mollusks, ostracodes and foraminifera. Chronologic ages are from Haq and others (1987).

STRATGRAPHIC UNIT	GEOLOGIC AGE	ESTIMATED AGE (m.y.)	AGE DIFFERENCE (m.y.)	THROW @ TOP (m)	SLIP DIFFERENCE	SLIP RATE (m/m.y.)
"upland unit"	Miocene (?)	10	28	0-9	0-9	0-0.3
Dry Branch Fm.	Late Eocene	38	2	9	3	1.5
Santee Ls.	Middle Eoc.	40	26	12	0	0
Peedee Fm.	Maestrichtian	66	19	12	18	1
Cape Fear Fm.	Santonian	85	----	30		

Keefer (1983), who suggested that subsurface faults near the South Carolina coast were reverse faults developed by reactivation of Triassic normal faults. Cretaceous-Tertiary east-west trending vertical faults characterize west Georgia and central Florida (?) with vertical displacements in early Tertiary strata as great as 60 m (Prowell, 1988). This region was described by Prowell (1988) as a zone of transition between the reverse fault and normal fault provinces of the Atlantic and Gulf Coast regions. Based on field observations, Prowell

(1988) reported that most northeast-trending faults of Late Cretaceous-Tertiary age exposed in the Atlantic Coastal Plain were reverse faults, and he concluded that they were caused by northwest-southeast compression. Recent upper crustal stress measurements in the central Savannah River Area on SRS (Zoback and others, 1989) indicate that the maximum compressive stress is oriented northeast. This is not a favorable orientation in which to activate northeast-southwest trending fault planes in compression. Also, vertical and steeply dipping reverse faults can be formed as a result of either compression or extension. (McClay and Ellis, 1987). Finally, the present northeast orientation of maximum compressive stress could cause strike slip movement on the Pen Branch fault. However, at this writing we find no seismic or core data which indicate strike slip displacement. Therefore, the origin of the stresses that resulted in the formation of the Pen Branch Fault and its post-Late Cretaceous movement history is unclear.

Capability

The presence of several nuclear installations at SRS near the Pen Branch Fault makes it important to note that we find no evidence to suggest that it is a "capable fault" as defined in 10 CFR-100-Appendix A (1988). The term "capable fault" has come into use in the regulatory literature to indicate a fault associated with seismicity or with the potential for seismicity. We reiterate that no disrupted soil profiles, scarps or linear features which could be ascribed to the Pen Branch Fault were observed in this study.

Further Investigations

Additional investigations at SRS and its environs are in progress which entail: (a) shallow reflection surveys; (b) well-drilling; and (c) reprocessing deep seismic data. The goals of these studies are: (a) to determine the age of the youngest stratigraphic unit which is cut by the Pen Branch Fault; (b) to provide additional information on the relationship between the Dunbarton basin and Pen Branch faults; and (c) to increase our knowledge of the relationship between the structure of the basement rocks and overlying Coastal Plain strata.

CONCLUSIONS

Constraints provided by well control and seismic reflection data show that the Pen Branch Fault extends northeast across the central portion of SRS for at least 15 mi (24 km). Based on a study of the aeromagnetic map of Petty and others (1965) the fault may continue northeast for 10 mi (16 km) beyond the boundary of SRS.

Movement on the Pen Branch Fault began in Late Cretaceous time (or earlier) and continued through the early part of the late Eocene (or later). The thickness of the Upper Cretaceous section is about 670 ft (20 m) on the down thrown side and about 610 ft (185 m) on the up thrown side. The throw of the fault varies from about 80 ft (24 m) to 100 ft (30 m) at the base of the Upper Cretaceous section to about 30 ft (9 m) at the top of the late Eocene Dry Branch Formation. These relationships indicate that the Pen Branch fault is a growth fault with estimated slip rates from 0 to 1.5 m/m.y. with an average of about 0.4 m/m.y. for the last 85 million years.

Based on its location and orientation it seems likely that the Pen Branch Fault is related to basement structures associated with the Dunbarton basin and is probably the result of reactivation of an early Mesozoic fault.

ACKNOWLEDGMENTS

This research was supported in part by funds from the U.S. Department of Energy and by the Westinghouse Savannah River Company, Contract number AA00878T. We thank the scientists listed below for their contributions to this paper. R.D. Hatcher, Jr. (University of Tennessee and Oak Ridge National Laboratory), A.L. Stievie (SRS), C.A. Eddy and R.K. Aadland read an early draft of the paper and made suggestions for its improvement. J. C. Behrendt and D.C. Colquhoun reviewed the manuscript, and we have incorporated several of their suggested revisions. D.E. Stephenson (SRS) discussed the results of an ongoing seismic reflection investigation at SRS. L.K. Price (SC Geological Survey) discussed the results of a drilling program adjacent to SRS. M.K. Harris (SRS) provided lithologic data on Upper Cretaceous and Tertiary cores. D.C. Prowell (U.S. Geological Survey) supplied data on the elevation of an Upper Cretaceous/Triassic contact. Tom Temples, USDOE and John Clarke supplied unpublished data The conclusions and opinions of the authors do not necessarily reflect those of the United States Department of Energy.

REFERENCES

Behrendt, J. C., Hamilton, R. M., Ackerman, H. D., and Henry, V. J., 1981, Cenozoic faulting in the vicinity of the Charleston, South Carolina, 1886 Earthquake: Geology, v.9, p.117-122.

Behrendt, J. C., Hamilton, R. M., Ackerman, H. D., Henry, V. J., and Bayer, K. C., 1983, Marine multi-channel seismic evidence for Cenozoic faulting and deep crustal structure near Charleston, South Carolina: in Gohn, G. S., ed., Studies Related to the Charleston, South Carolina Earthquake of 1886 - Tectonics and Seismicity: U. S. Geological Survey Professional Paper 1313, p. J1-J29.

Behrendt, J.C., Yaun, A., 1987, The Helena Banks strike-slip(?) fault zone in the Charleston, South Carolina, earthquake area: Results from a marine, high-resolution, multichannel, seismic reflection survey. Geological Society of America Bulletin, v.98, p.591-601.

Behrendt, J.C., Yaun, A., 1986, Strike slip on reactivated Triassic(?) basin boundary fault zones as sources of earthquakes near Charleston, S.C. in Proceedings of the Third U.S. National Conference on Earthquake Engineering, Aug. 24-28, 1986, Charleston, South Carolina.

Behrendt, J. C., Klitgord, K. D., and Hutchinson, D. R., 1988, Reactivated boundary fault zones of buried early Mesozoic basins as possible sources of seismicity in the Charleston, SC region: Geological Society of America Abstracts with Programs, v.20, p. 253.

Brown, P. M., Miller, J. A., and Swain, F. M., 1972, Structural and stratigraphic framework and spatial distribution of permeability of the Atlantic Coastal Plain, North Carolina to New York: U. S. Geological Survey Professional Paper 796, 79p.

Chapman, W. L., and Di Stefano, 1989, Savannah River Plant Seismic Survey, 1987-88, Research Report 1809-005-006-1-89: Conoco, Inc. Seismic Acquisition Section, 110p.

Colquhoun, D. J., Woolen, I. D., Van Nieuwenhuise, D. S., Padgett, G. G., Oldham, R. W., Boylam, D. C., Bishop, J. W., and Howell, P. D., 1983, Surface and subsurface stratigraphy, structure and aquifers of the South Carolina Coastal Plain: South Carolina Department of Health and Environmental Control, Governor's Office, and University of South Carolina, Columbia, SC, 78p.

Cooke, C. W., 1936, Geology of the Coastal Plain of South Carolina: U. S. Geological Survey Bulletin 867, 196p.

Cooke, C. W. and MacNeil, F. S., 1952, Tertiary stratigraphy of South Carolina: U. S. Geological Survey Professional Paper 243B, p. 19-29.

Cramer, H. R., and Arden, D. D., Jr., 1978, Faults in Oligocene rocks of the Georgia Coastal Plain: Geological Society of America Abstracts with Programs, v.10, p.166.

Cumbest, R. J., Price, V. and Anderson, E. E., 1992, Gravity and magnetic modeling of the Dunbarton Triassic Basin, South Carolina: Southeastern Geology, v. 33(1), p. 37-51.

Daniels, D. L., Zietz, I., and Popenoe, P., 1983, Distribution of subsurface lower Mesozoic rocks in the southeastern United States as interpreted from regional aeromagnetic and gravity maps: in Gohn, G. S., ed., Studies Related to the Charleston, South Carolina Earthquake of 1886 - Tectonics and Seismicity: U. S. Geological Survey Professional Paper 1313K, p. K1-K24.

Fallaw, W. C., and Price, V., 1992, Outline of stratigraphy at the Savannah River Site: in W. Fallaw and V. Price, eds., Geological Investigations of the Central Savannah River Area, South Carolina and Georgia, Carolina Geological Society Field Trip Guide Book, Nov. 13-15, 1992.

Fallaw, W. C., Snipes, D. S., Daggett, J. S., Furlow, L., James, A., Moore, J. P., Price, S., Price, V., 1992, Stratigraphic relationships in Eocene outcrops along Upper Three Runs at the Savannah River Site: in W. Fallaw and V. Price, eds., Geological Investigations of the Central Savannah

River Area, South Carolina and Georgia, Carolina Geological Society Field Trip Guide Book, Nov. 13-15, 1992.

Gohn, G. S., Hazel, J. E., Bybell, L. M., and Edwards, L. E., 1983, The Fishburne Formation (lower Eocene), a newly defined subsurface unit in the South Carolina Coastal Plain: U. S. Geological Survey Bulletin, 1537-C, 16p.

Hamilton, R. M., Behrendt, J. C., and Ackerman, H. D., 1983, Land multichannel seismic-reflection evidence for tectonic features near Charleston, South Carolina: in Gohn, G. S., ed., Studies Related to the Charleston, South Carolina Earthquake of 1886 - Tectonics and Seismicity: U. S. Geological Survey Professional Paper 1313K, p. I1-I18.

Higgins, B. B., Gohn, G. S., and Bybell, L. M., 1978, Subsurface geologic evidence for normal faults in the South Carolina Coastal Plain near Charleston: Geological Society of America Abstracts with Programs, v.10, p.171.

Howell, D. E., and Zupan, A. W., 1974, Evidence for post-Cretaceous tectonic activity in the Westfield Creek Area, north of Cheraw, South Carolina: Geologic Notes, South Carolina State Development Board, v.18, p.98-105.

Huddleston, P. F., and Hetrick, J. H., 1985, Upper Eocene stratigraphy of central and eastern Georgia: Georgia Geologic Survey Bulletin, v.95, 78p.

Inden, R. F., and Zupan, A. W., 1975, Normal faulting of upper Coastal Plain sediments, Ideal Kaolin Mine, Langley, South Carolina: Geologic Notes, South Carolina State Development Board, v.19, p.161-165.

Kish, S. A., 1992, An initial geochemical and isotopic study of granite from core C-10, Savannah River Site, S.C.: in W. Fallaw and V. Price, eds., Geological Investigations of the Central Savannah River Area, South Carolina and Georgia, Carolina Geological Society Field Trip Guide Book, Nov. 13-15, 1992.

LaMoreaux, P. E., 1946a, Geology of the Coastal Plain of east-central Georgia: Georgia Geological Survey Bulletin, v. 50, 26p.

LaMoreaux, P. E., 1946b, Geology and ground-water resources of the Coastal Plain of east-central Georgia: Georgia Geological Survey Bulletin, v. 52, 173p.

Logan, R. W., and Euler, G. M., 1989, Geology and groundwater resources of Allendale, Bamberg, and Barnwell Counties and part of Aiken County, South Carolina: S. C. Water Resources Commission, Report No. 155, 113p.

Marine, I. W. and Siple, G. E., 1974, Buried Triassic basin in the central Savannah River area, South Carolina and Georgia: Geological Society of America Bulletin, v.85, p.311-320.

McClay, K.R., 1989, Physical models of structural styles during extension. in Extensional Tectonics and Stratigraphy of the North Atlantic Margins. AAPG Memoir 46, p.95-110.

McClay, K. R., and Ellis, P. G., 1987, Geometries of extensional fault systems developed in model experiments: Geology, v.15, p.341-344.

Nystrom, P. G., Jr., and Willoughby, R. H., 1982, Cretaceous, Tertiary and Pleistocene (?) stratigraphy of Hollow Creek and Graniteville Quadrangles, Aiken County, South Carolina, in Nystrom, P. G., Jr. and Willoughby, R. H., eds., Geological Investigations Related to the Stratigraphy in the Kaolin Mining District, Aiken County, South Carolina: Carolina Geological Society Field Trip Guidebook, 1982, South Carolina Geological Survey, Columbia, SC, p. 80-113.

Petty, A. J., Petrafeso, F. A., and Moore, F. C., Jr., 1965, Aeromagnetic map of the Savannah River Plant area, South Carolina and Georgia: U. S. Geological Survey Geophysical Investigations Map GP-489.

Price, L. K., 1990, Geologic map of the Williston 7.5-minute quadrangle: South Carolina Geological Survey Open File Report 68.

Prowell, D. C., 1983, Index of faults of Cretaceous and Cenozoic age in the eastern U. S.: U. S. Geological Survey Field Studies Map MF-1269.

Prowell, D. C., Christopher, R. A., Edwards, L. E., Bybell, L. M., and Gill, H. E., 1985, Geologic section of the up dip Coastal Plain from central Georgia to western South Carolina: U. S. Geological Survey Miscellaneous Field Studies Map MF-1737, 11p.

Prowell, D. C., 1988, Cretaceous and Cenozoic tectonism on the Atlantic Coastal Margin: in Sheridan, R. E., and Grow, J. A., eds., The Geology of North America v. 1-2, The Atlantic Continental Margin, U. S., Geological Society of America, p. 557-564.

Reinhardt, J., Prowell, D. C., and Christopher, R. A., 1984, Evidence for Cenozoic tectonism in the southwest Georgia Piedmont: Geological Society of American Bulletin, v.95, p. 1176-1187.

Schilt, S. F., Brown, L. D., Oliver, J. E., and Kaufman, S., 1983, Subsurface structure near Charleston, South Carolina - Results of COCORP reflection profiling in the Atlantic

Coastal Plain: in Gohn, G.S., ed., Studies Related to the Charleston, South Carolina Earthquake of 1886 - Tectonics and Seismicity, U. S. Geological Survey Professional Paper 1313, p. H1-H19.

Shedlock, K. M., and Harding, S. T., 1988, Structure of the shallow crust near Charleston, South Carolina: Geological Society of America Abstracts with Program, v.20, p.313.

Shomo, S. J., 1980, Stratigraphic evidence for Cretaceous - Tertiary offset of Triassic strata beneath the Coastal Plain: Geological Society of America Abstracts with Program, v.12, p.521.

Shomo, S. J., 1982, Geology of buried Triassic - Jurassic basin, Coastal Plain, Virginia: M. S. Thesis, The University of North Carolina, Chapel Hill, NC, 57p.

Siple, G. E., 1967, Geology and ground water of the Savannah River Plant and vicinity, South Carolina: U.S. Geological Survey Water Supply Paper 1841, 113p.

Sloan, E., 1908, Catalogue of the mineral localities of South Carolina: South Carolina Geological Survey, Series IV, Bulletin 2, The State Co., Columbia, SC, Reprinted by the South Carolina Geological Survey, Columbia, SC 1979, 506 p.

Snipes, D. S., Fallaw, W. C., and Price, V., Jr., 1992, Structural geology of the Savannah River Site in the Coastal Plain of South Carolina, in Zullo, V. A., Harris, W. B., and Price, V., Jr. eds., University of North Carolina at Wilmington and U. S. Dept. of Energy, P. 33-36.

Speer, J. A., 1982, Descriptions of the granitoid rocks associated with two gravity minima in Aiken and Barnwell counties, South Carolina: South Carolina Geology, v. 26, p.15-24.

Steele, K. B., 1985, Lithostratigraphic correlation of Cretaceous and younger strata of the Atlantic Coastal Plain Province within Aiken, Allendale and Barnwell Counties, South Carolina: M. S. Thesis, University of South Carolina, Columbia South Carolina, 174p.

Steele, K. B., and Colquhoun, D. J., 1985, Subsurface evidence of the Triassic Newark Supergroup in the South Carolina Coastal Plain: South Carolina Geology, v. 28, p. 11-22.

Talwani, P., 1986, Current thoughts on the cause of the Charleston, South Carolina earthquakes: South Carolina Geology, v.29, p. 19-38.

Traverse, A., 1987, Pollen and spores date origin of rift basins from Texas to Nova Scotia as early Late Triassic: Science, v. 236, p. 1469-1472.

Wentworth, C. M., and Mergner-Keefer, M., 1983, Regenerate faults of small Cenozoic offset - Probable earthquake sources in the southeastern United States: in Gohn, G.S., ed., Studies Related to the Charleston, South Carolina Earthquake of 1886 - Tectonics and Seismicity, U. S. Geological Survey Professional Paper 1313, p. S1-S20.

Wernicke, B., Axen, G.J., 1988, On the role of isostasy in the evolution of normal fault systems. Geology, v.16, p.848-851.

Zoback, M. D., Healy, J. H., Roller, J. C., Gohn, G. S., and Higgins, B. B., 1978, Normal faulting and in situ stress in the South Carolina Coastal Plain near Charleston: Geology, v.6, p.147-152.

Zoback, M. D., Moos, D., and Stephenson, D. E., 1989, State of stress and the relation to tectonics in the central Savannah River area of South Carolina: in Khair, A. W., ed., Rock Mechanics as a Guide to Efficient Utilization of Natural Resources, Proceedings of the 30th U. S. Symposium, A. A. Balkema, Rotterdam, Netherlands, p.553-560.

Zupan, A. W., and Abbott, W. H., 1975, Clastic dikes: Evidence for post-Eocene (?) tectonics in the Upper Coastal Plain of South Carolina: Geologic Notes, South Carolina State Development Board, v.19, p.14-23.