SOUTHEASTERN GEOLOGY

PUBLISHED AT DUKE UNIVERSITY DURHAM, NORTH CAROLINA

VOL.21. NO.4 NOVEMBER, 1980

SOUTHEASTERN GEOLOGY

PUBLISHED QUARTERLY

AT

DUKE UNIVERSITY

This journal welcomes original papers on all phases of geology, geophysics, and geochemistry as related to the Southeast. Transmit manuscripts to S. DUNCAN HERON, JR., BOX 6665, COLLEGE STATION, DURHAM, NORTH CAROLINA 27708. 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 as finished copy.
4) Make certain that all photographs are sharp, clear, and of good contrast.
5) Stratigraphic terminology should abide by the code of Stratigraphic Nomenclature (AAPG, v. 45, 1961).

Proofs will be sent authors.

Reprints must be ordered prior to publication; prices available upon request.

Subscriptions to Southeastern Geology are $7.00 per volume (US and Canada) or $8.00 per volume (foreign). Inquiries should be sent to: SOUTHEASTERN GEOLOGY, BOX 6665, COLLEGE STATION, DURHAM, NORTH CAROLINA 27708. Make checks payable to: Southeastern Geology.

SOUTHEASTERN GEOLOGY

Table of Contents

Vol. 21, No. 4

November 1980

SEISMIC REFLECTION, REFRACTION AND GRAVITY MEASUREMENTS FROM THE CONTINENTAL SHELF OFFSHORE FROM NORTH AND SOUTH CAROLINA

By

M.S. Grim
U.S. Geological Survey
Denver, Colorado 80225

W.P. Dillon
U.S. Geological Survey
Woods Hole, Massachusetts 02543

and

R.E. Mattick
U.S. Geological Survey
Reston, Virginia 22092

ABSTRACT

A compilation of geophysical data collected on the coastal plain and offshore from North Carolina and South Carolina raises questions about the nature of the pre-Cretaceous basement rocks in the area. An interpretation of the data from the Continental Shelf of South Carolina and southern North Carolina must take into account the presence of local Bouguer gravity anomalies in excess of +80 mgals and -60 mgals, the presence of a smooth seismic reflecting horizon at depths where seismic velocities range from 5.6 to 6.1 km/sec, and the presence of deeper refractors having velocities in excess of 6.3 km/sec. Our interpretation of these data suggests that the smooth reflector represents basalt flows that overlie both a complex crystalline basement and areas of Triassic(?) sediments.

INTRODUCTION

The term "Southeast Georgia embayment" is used in this paper to define the recession into the Atlantic Coastal Plain that interrupts the uniform slope of the basement surface away from the south flank of the Cape Fear Arch. This definition follows the usage of Maher (1971) and Dillon and others (1975). The embayment extends southwestward from the Cape Fear Arch to the Peninsular Arch in Florida (Figure 1) and opens toward the sea.

Numerous onshore wells, some of which have penetrated basement rocks, have been drilled in the embayment. On the basis of data from these wells, the rocks in the onshore part of the Southeast Georgia embayment can be divided into two general categories: (1) Coastal Plain sediments of Cretaceous and younger age, and (2) underlying pre-Cretaceous rocks of the basement complex. In terms of present sediment deposition, the Southeast Georgia embayment lies in a transitional zone between a predominantly clastic depositional province north of Cape Hatteras and a carbonate province that includes Florida and the Bahamas.

A stratigraphic section (Figure 2) from the Peninsular Arch in Florida to the Cape Fear Arch in North Carolina shows the general distribution of Coastal Plain sediments along the Atlantic coast. Maximum thickness of Cretaceous and younger Coastal Plain sediments is about 1.4 km. Lower Cretaceous rocks, which are absent in the vicinity of the Cape Fear Arch, may be represented by thin (30-90 m) nonmarine clastic sediments penetrated near the bottom of the California Oil Buie No. 1 well (Brown, 1974). In general, Upper Cretaceous rocks grade from terrigenous sands and shales

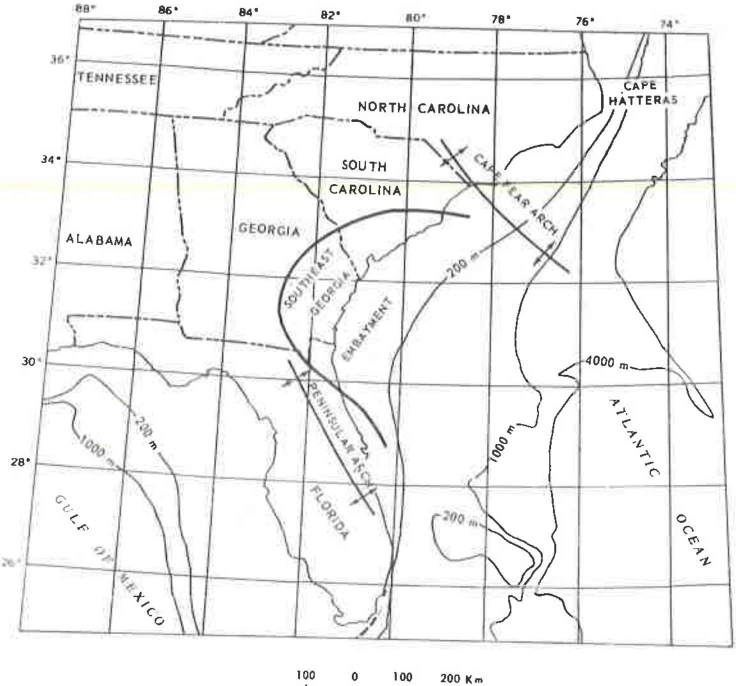

Figure 1. Index map showing general outline of the Southeast Georgia embayment (Dillon and others, 1975). Bathymetric contours are in meters.

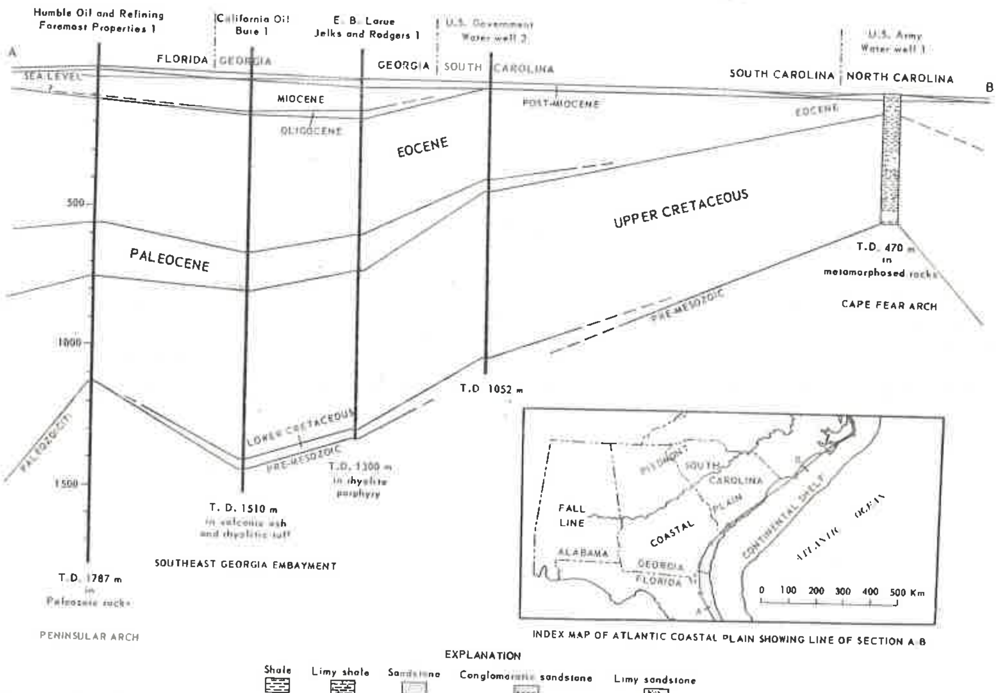

Figure 2. Stratigraphic section across the Southeast Georgia embayment. Section was modified from Maher (1971).

inland in Georgia and South Carolina to more marine chalks, limestones, and dolomites near the coastline (Dillon and others, 1975). The Cenozoic rocks, predominently shallow-water marine carbonates in Florida, grade northward into marginal marine clastic facies composed of sands, marls, and limestone (Dillon and others, 1975).

Pre-Cretaceous rocks of the basement complex can be divided into: (1) early Paleozoic and late Precambrian igneous and metamorphic rocks similar to those exposed in the Appalachian Piedmont province (Spivak and Shelburne, 1971); and (2) Triassic deposits of nonmarine arkoses, sandstones, shales, basic flows, and diabase intrusions. Where exposed from northern South Carolina to Nova Scotia, Triassic rocks occur in northeast-southwest trending downfaulted grabens (Dillon and others, 1975).

In the offshore Southeast Georgia embayment, where data from deep wells are unavailable, early authors (e.g., Woollard and others, 1957) defined "Pre-Cretaceous basement" on the basis of seismic refraction velocities greater than 5.6 km/sec. Recently collected multichannel seismic reflection data suggest that some of these high velocities could represent a layer of Mesozoic volcanic rocks. These volcanic rocks may overlie sedimentary rocks, Paleozoic basement rocks containing mafic intrusions, and possibly Triassic basins and granitic intrusions.

This paper presents newly acquired multichannel seismic reflection and seismic refraction data together with previously published gravity data of Krivoy and Eppert (1977). These new data were collected in 1975 during a cruise jointly sponsored by the U.S. Geological Survey (USGS) and l'Institut Francais du Petrole (IFP) (Dillon and others, 1979). Multichannel seismic reflection profiles were collected using the IFP Flexichoc acoustic source, a large-low pressure airgun that produces a relatively simple pulse equivalent to about a 1000 in^3 array of airguns. Reflection data were processed using a 2400 percent common-depth-point (CDP) stack. The same sound source was used with disposable sonobuoys to record four refraction profiles on the Continental Shelf.

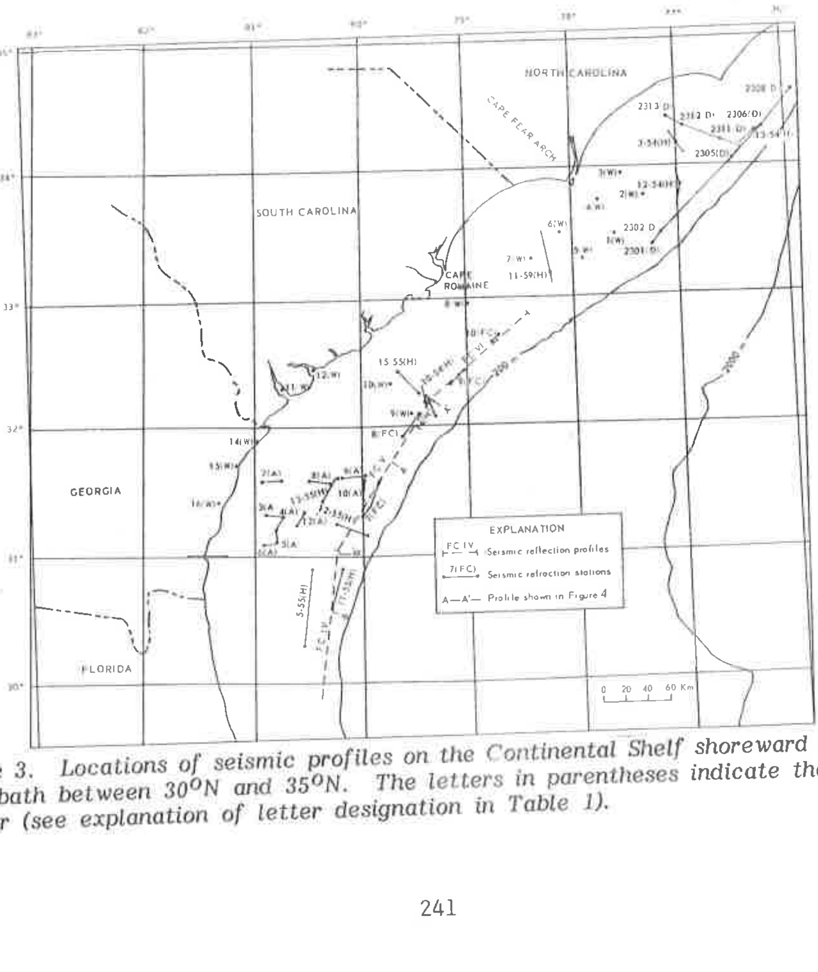

Figure 3. Locations of seismic profiles on the Continental Shelf shoreward of the 200 m isobath between 30°N and 35°N. The letters in parentheses indicate the principal worker (see explanation of letter designation in Table 1).

GEOPHYSICAL STUDIES

Seismic Reflection

The locations of the USGS-IFP seismic reflection profiles discussed in this paper are shown in Figure 3 and two seismic sections are shown in Figure 4. All records show horizontal reflectors above a smooth, strong reflector that was recorded at depths of between 1.5 and 2.0 seconds of two-way travel time. On the records from Lines FC-V and FC-IV (Figures 4a and 4b) there are weak indications that suggest the existence of reflectors beneath the smooth, strong reflector. On FC-IV (Figure 4b), the smooth, strong reflector loses its distinctive character at about 31°N and a rough acoustic horizon is apparent at 2.0 to 3.0 seconds depth.

Seismic Refraction

Figure 3 shows the locations of seismic refraction profiles obtained during the 1975 USGS-IFP program, 7(FC), 8(FC), 9(FC), and 10(FC), as well as the earlier work of Woollard and others (1957), Hersey and others (1959), Antoine and Henry (1965), and Dowling (1968). The results of these surveys are given in Table 1. The station numbering system is the same as that used by the original authors. Most seismic refraction studies have indicated the presence of four layers, and these were defined on the basis of similar depths and velocities. When the results from all the studies were tabulated together, we found it convenient to divide them into six groups based on velocity. Our primary interest in this paper is the nature of the rocks represented by the velocities in columns V and VI in Table 1.

Seven velocities greater than 6.3 km/sec are listed in column VI (Table 1). At three additional onshore stations, Woollard and others (1957) measured velocities equal to or in excess of 6.3 km/sec. The locations of these ten stations and the depths to the high velocity refractor are shown in Figure 5.

The two lowest of these high velocity values, 6.30 km/sec (Hersey and others, 1959) and 6.36 km/sec (Antoine and Henry, 1965), are believed to be the most accurate because they were obtained at locations where both forward and reverse recording was employed. The remaining velocities listed in column VI, recorded on unreversed lines, might be somewhat high. On lines, FC-8 and 9(W), velocities of 6.59 km/sec were recorded. These extremely high velocities were recorded on essentially perpendicular lines, thereby partially eliminating the problem of increased apparent velocity caused by dip of the refracting surface along unreversed profiles.

These velocities indicate the presence of high density rock. It should be noted, however, that they were generally recorded in areas of negative gravity anomalies (Figure 6). If the volumes of these high-velocity rocks are small, their effect on the measured gravity would be minimal.

Gravity

The most striking features on the Bouguer gravity map of the Southeast Georgia embayment (Figure 6) are the north-south elongate positive anomalies in excess of 80 mgals at 78.7°W and the adjacent areas where negative anomalies are lower than -40 mgals. The location of the gravity highs coincides with the location of large magnetic highs (Taylor and others, 1968; Klitgord and Behrendt, 1977).

INTERPRETATION OF DATA

The data in Table 1 were used to construct a contour map showing depth to refractors having velocities between 5.6 and 6.1 km/sec (Figure 7). On the basis of onshore refraction results, where well control was available, Woollard and others (1957) were able to show that velocities in the 5.6-6.1 km/sec range represented pre-Cretaceous rocks, and this relationship was extended to the offshore area where drill

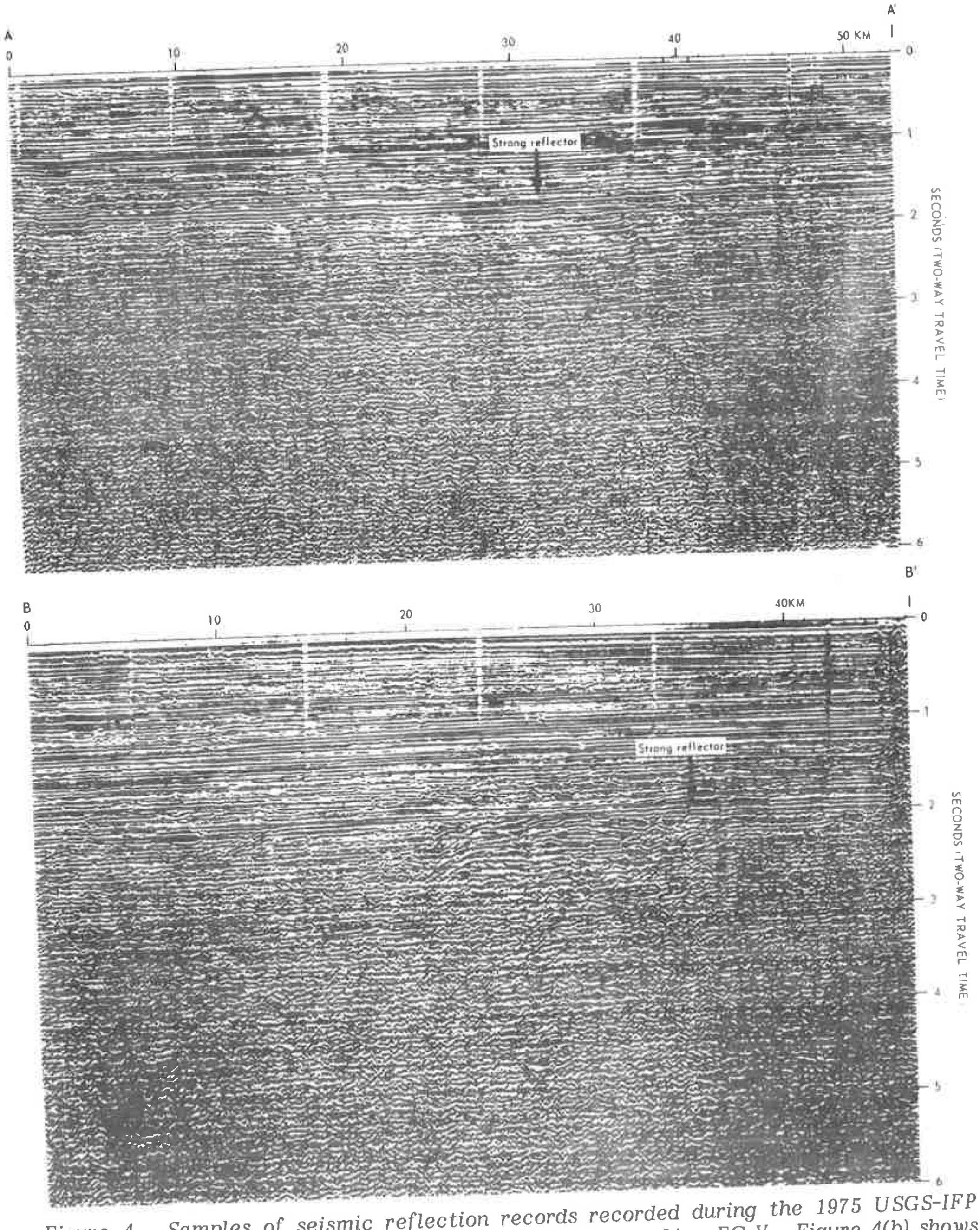

Figure 4. Samples of seismic reflection records recorded during the 1975 USGS-IFP cruise. Figure 4(a) shows the smooth, strong reflector on Line FC-V. Figure 4(b) shows reflections from horizons beneath this strong reflector on Line FC-IV.

hole data were unavailable. The term "basement" was applied to this pre-Cretaceous surface. In the Cape Fear area it was generally assumed that velocities between 5.6 and 6.1 km/sec represented crystalline Paleozoic rocks.

Superimposed on the contour map of Figure 7 are depths to the smooth, strong reflector (Figure 4). The excellent agreement in depth between the two horizons suggests that seismic arrivals of velocities of 5.6-6.1 km/sec were refracted along the smooth, strong reflector or a horizon at approximately the same depth. This smooth,

Table 1. Compilation of the results of seismic refraction surveys in the Southeast Georgia embayment. Results are grouped by measured velocity. Velocities of less than 5.3 km/sec are believed to represent Cretaceous and younger sediments. Depths are given in km below sea level and velocities in km/sec. Letters in parentheses indicate principal investigator: (A) Antoine and Henry (1965), (D) Dowling (1968), (H) Hersey and others (1959), (W) Woollard and others (1957), (FC) recently acquired USGS-IFP data. Velocities (v) and depths (d).

Station	1.0-2.2 v	k/s d	2.2-2.8 v	k/s d	2.8-3.1 v	k/s d	3.4-4.3 v	k/s d	5.3-6 v	k/s d	6.3-7+ v	k/s d
2313(D)	2.0*	.02	2.44	.38			3.43	.76	5.96	1.07		
2312(D)	2.0*	.03	2.44	.35			3.43	.84	5.96	1.47		
2311(D)	2.0*	.04	2.44	.21			3.43	1.04	5.96	2.72		
2301(D)	2.0*	.04	2.74	.67					5.82	1.68		
2302(D)	2.0*	.04	2.74	.67					5.82	1.83		
2305(D)	2.0*	.04	2.67	.39			3.74	1.39	5.94	2.54		
2306(D)	2.0*	.04	2.67	.69			3.74	1.58	5.94	3.16		
2308(D)	2.0*	.04	2.67	.98			4.27	1.84				
13-54(H)	1.84	.03	2.41	.53			3.88	1.32				
3-54(H)	1.86	.03	2.48	.25					5.49	1.16		
12-54(H)	1.88	.04	2.46	.40					5.90	1.47		
2(W)	1.62	.03	2.37	.35					5.88	1.31		
3(W)	2.18	.02	2.51	.08					5.91	.77		
4(W)	2.10	.02	2.36	.21					5.91	.71	7.01	1.29
1(W)	1.71	.02							5.79	.88		
6(W)	2.23	.02	2.33	.32					5.79	.64	6.46	1.28
5(W)	1.67	.03	2.32	.33					5.82	.91		
11-54(H)	2.05	.03	2.26	.09					5.89	.86		
7(W)	1.95	.03							6.25	.73		
8(W)	2.00	.02	2.43	.30					5.76	.94		
10(FC)	1.87	.05	2.19	.31								
9(FC)	1.82	.04			2.85	.70			5.94	1.61	6.87	3.50
15-55(H)$^{N}_{S}$	1.90	.03 .03	2.44	.35 .48					5.98	1.28 1.77		
10-54(H)	1.66	.03	2.71	.50					6.04	1.78		
14-55(H)$^{N}_{S}$	1.77	.05 .05	2.65	.62 .61					5.96	2.15 2.40		
10(W)	1.74	.02	2.48	.23					6.13	1.42		
9(W)	1.76	.04	2.28	.35					5.64	2.04	6.59	3.39
8(FC)	1.78	.04	2.68	.59							6.59	2.12
12(W)	2.12	.006	2.34	.08					6.04	1.18		
11(W)	1.80	.01							6.07	.93		
14(W)	1.89	.01	2.59	.22					5.88	1.23		
15(W)	1.87	.006	2.56	.16					5.88	1.30		
7(A)$^{W}_{E}$	1.61	.03 .03	2.36	.13 .13	2.85	.41 .48			5.85	1.63 1.65		
8(A)$^{W}_{E}$	1.76	.03 .04	2.47	.20 .25	3.03	.60 .70			5.96	1.79 1.87		
13-55(H)$^{S}_{N}$	1.83	.03 .03	2.56	.31 .25			3.45	.73 1.09	5.95	2.59 2.27		
9(A)$^{W}_{E}$	1.72	.045 .05	2.61	.23 .24	3.09	.69 .78			5.76	1.99 2.21		
10(A)$^{S}_{N}$	1.72	.05 .05	2.66	.28 .27	3.05	.72 .68			5.45	2.20 2.23		
7(FC)	1.86	.04	2.70	.32					5.89	2.31		
16(W)	1.76	.006	2.47	.14	2.99	.31			5.30	1.24		
3(A)$^{W}_{E}$	1.72	.02 .03	2.32	.12 .06	2.96	.42 .35					6.36	1.64 1.72
4(A)$^{N}_{S}$	1.67	.03 .03	2.31	.14 .15	2.96	.45 .48			5.52	1.68 1.65		
5(A)$^{N}_{S}$	1.63	.03 .03	2.24	.14 .17	3.04	.43 .45			5.39	1.78 1.94		
6(A)$^{E}_{W}$	1.72	.03 .03	2.38	.18 .20	3.00	.52 .48			5.81	1.93 1.80		
12(A)$^{S}_{N}$	1.60	.05 .05	2.45	.17 .18	2.92	.47 .46			5.94	1.91 1.78		
12-55(H)$^{W}_{E}$	1.95	.04 .04			2.88	.34 .44					6.30	2.79 2.71
5-55(H)$^{N}_{S}$	1.68	.03 .03	2.57	.26 .14			3.65	.98 .48	5.86	3.74 4.92		
11-55(H)$^{N}_{S}$	1.92	.04 .04	2.34	.14 .26	2.85	.55 .55	3.43	1.01 1.06	5.96	3.05 3.10		

* assumed velocity

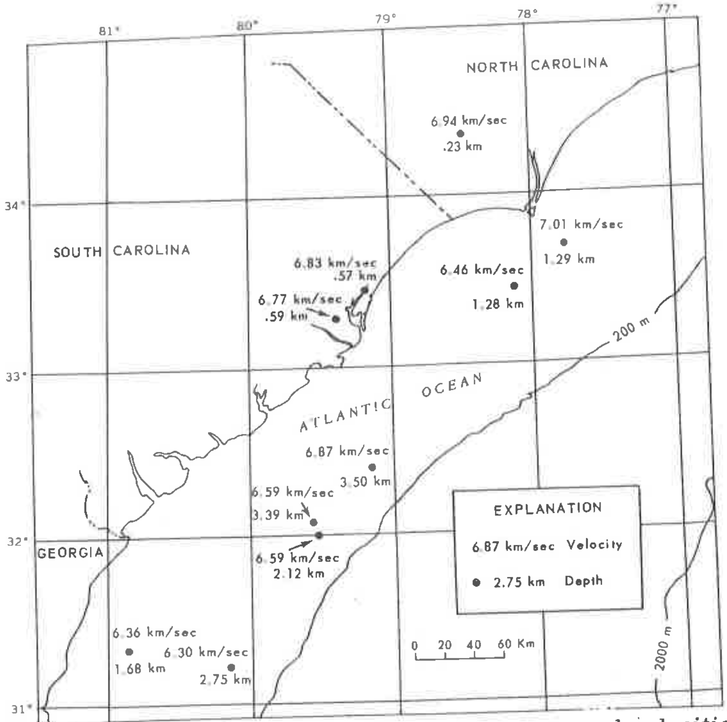

Figure 5. Locations where seismic refraction surveys measured velocities equal to or in excess of 6.3 km/sec. Measured velocities and depths of high-velocity refractor are indicated on the map.

strong reflective horizon has been traced on seismic reflection records over an extensive area of the offshore Southeast Georgia embayment between 31ºN and 35ºN, where it appears to be continuous and is overlain by about 1 km of essentially flat-lying undisturbed sediments (Dillon and others, 1979).

Drilling near Charleston, South Carolina, has revealed basalt flows at a depth of 0.8 km (Gohn, Higgins, and others, 1978). K-Ar dates indicate that the basalt is about 160-200 m.y. old (Triassic-Early Jurassic age) (Gohn, Gottfried, and others, 1978). By projecting the smooth, strong reflector onshore, Dillon and others (1979) correlated the reflector with these basalt flows.

South of 31ºN reflections are recorded from a horizon about 1 to 1.5 km deeper than the smooth, strong reflector. These deeper reflections could represent Paleozoic metasediments or rocks of a crystalline basement complex. The interval between the strong reflector and this deeper set of reflections might represent Triassic or Lower Jurassic deposits. Although we cannot be sure that such lower Mesozoic deposits exist beneath the smooth reflector north of 31ºN, their presence would help explain the large negative gravity anomalies that are observed there and are discussed in the next paragraph.

On the Bouguer gravity map of the Southeast Georgia embayment (Figure 6), a north-south elongate positive anomaly appears between 32ºN and 34ºN. According to Harold Krivoy (USGS, written commun., 1977), the amplitude of this positive gravity anomaly, its gradient, and its association with a large positive aeromagnetic anomaly indicate that the source of the anomaly is a high-density mafic body of rock that has been intruded to relatively shallow depths (possibly 1-2km). To the east of this positive gravity feature, Bouguer values are as low as -60 mgals. The mass deficit that must correspond to such low gravity values is difficult to explain. The short horizontal extent of the steep gradients suggests a shallow source for the negative anomalies. Because post-Triassic sediments appear to simply thicken seaward, the source is believed to be a mass deficit occurring immediately below or at shallow depths below the smooth, strong reflector.

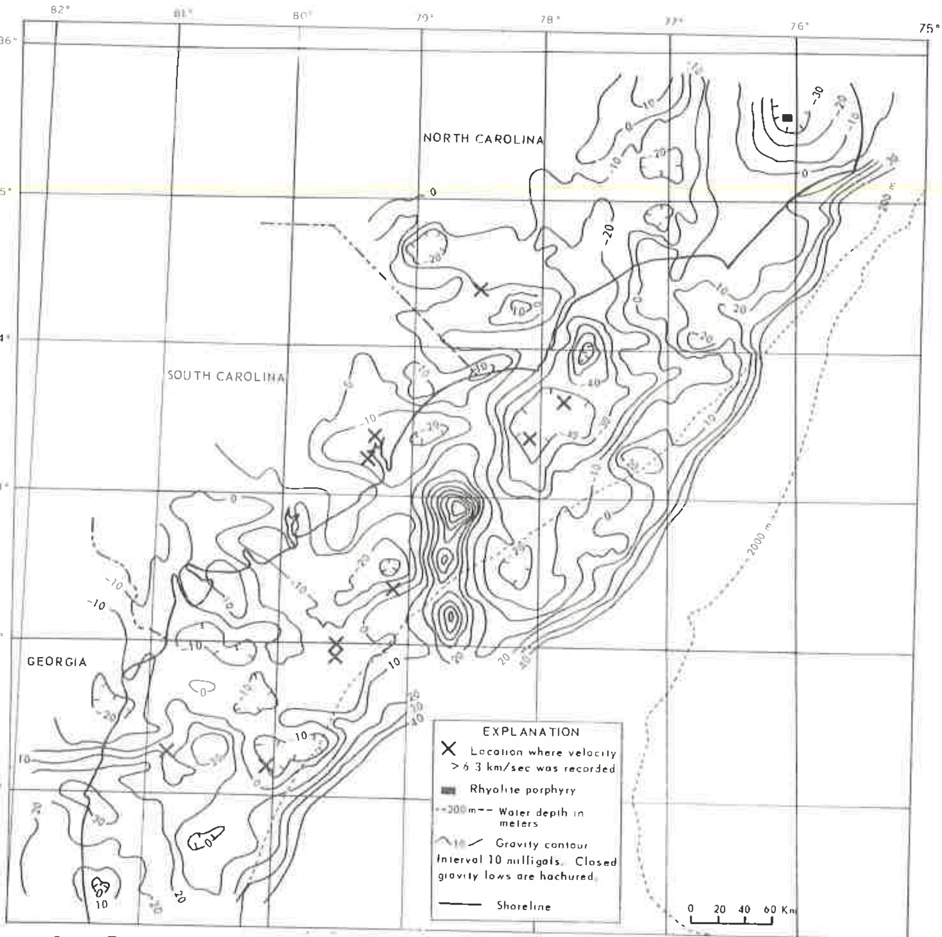

Figure 6. Bouger gravity map of the Continental Shelf in the Southeast Georgia embayment area. Figure is modified from Krivoy and Eppert (1977).

Sedimentary rocks or low density granitic rock could account for the mass deficit. Buried Triassic basins, containing chiefly nonmarine sediments, are present in the region. These include the Dunbarton Basin (Marine and Siple, 1974), the Florence Basin (MacCarthy, 1936; Bonini and Woollard, 1960) and the Durham-Wadesboro Basin (Randazzo and others, 1970; Ackermann and others, 1976).

Several onshore wells in coastal North Carolina have penetrated rhyolite basement (Maher, 1971; Brown and others, 1973). North of Cape Hatteras, on the northern shore of Albemarle Sound near 36°N, 76°W (Figure 6), Denison and others (1967) reported an altered rhyolite porphyry, which is apparently associated with a negative Bouguer gravity anomaly lower than -30 mgals.

To estimate the thickness of the low density rock the following assumptions were made. The surrounding basement was assigned a density of 2.73 gm/cc. This density, assigned to the Appalachian-Atlantic coast area by Woollard (1962), was chosen because it allows plausible density contrasts in calculations involving either the negative anomalies of -60 mgals or the positive anomalies of +80 mgals. An average density of 2.6 gm/cc was assigned to the rocks causing the low anomalies. This density is appropriate for either sedimentary rocks or low-density granitic rock (Daly and others, 1966). A thickness of about 7.5 km of 2.6 gm/cc rock is necessary to produce a negative anomaly of about -42 mgals, which, except for one small area, approximates the lowest negative anomaly values. If the density contrast were assumed to be 0.18, then the source of the anomaly would be about 5.5 km thick.

To explain the Georgetown, South Carolina, gravity low, Talwani and others (1975)

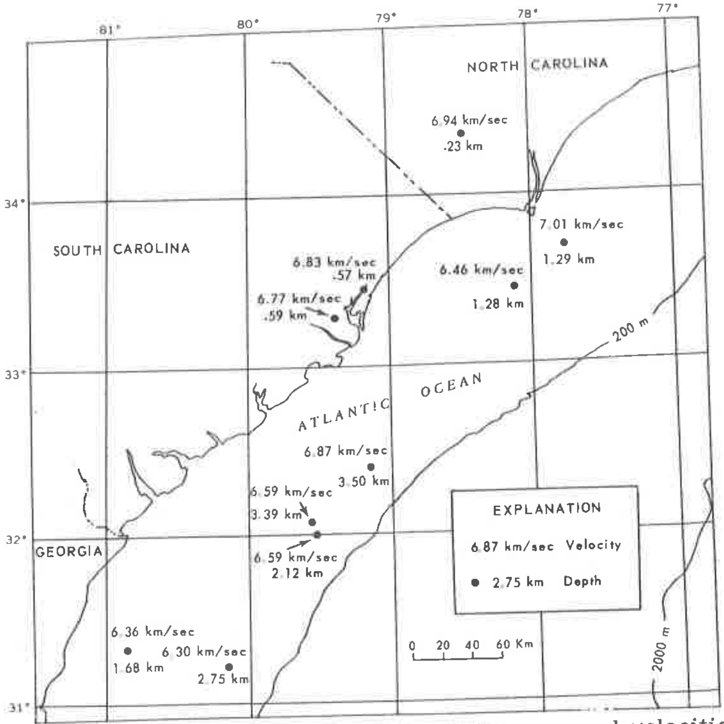

Figure 5. Locations where seismic refraction surveys measured velocities equal to or in excess of 6.3 km/sec. Measured velocities and depths of high-velocity refractor are indicated on the map.

strong reflective horizon has been traced on seismic reflection records over an extensive area of the offshore Southeast Georgia embayment between 31°N and 35°N, where it appears to be continuous and is overlain by about 1 km of essentially flat-lying undisturbed sediments (Dillon and others, 1979).

Drilling near Charleston, South Carolina, has revealed basalt flows at a depth of 0.8 km (Gohn, Higgins, and others, 1978). K-Ar dates indicate that the basalt is about 160-200 m.y. old (Triassic-Early Jurassic age) (Gohn, Gottfried, and others, 1978). By projecting the smooth, strong reflector onshore, Dillon and others (1979) correlated the reflector with these basalt flows.

South of 31°N reflections are recorded from a horizon about 1 to 1.5 km deeper than the smooth, strong reflector. These deeper reflections could represent Paleozoic metasediments or rocks of a crystalline basement complex. The interval between the strong reflector and this deeper set of reflections might represent Triassic or Lower Jurassic deposits. Although we cannot be sure that such lower Mesozoic deposits exist beneath the smooth reflector north of 31°N, their presence would help explain the large negative gravity anomalies that are observed there and are discussed in the next paragraph.

On the Bouguer gravity map of the Southeast Georgia embayment (Figure 6), a north-south elongate positive anomaly appears between 32°N and 34°N. According to Harold Krivoy (USGS, written commun., 1977), the amplitude of this positive gravity anomaly, its gradient, and its association with a large positive aeromagnetic anomaly indicate that the source of the anomaly is a high-density mafic body of rock that has been intruded to relatively shallow depths (possibly 1-2km). To the east of this positive gravity feature, Bouguer values are as low as -60 mgals. The mass deficit that must correspond to such low gravity values is difficult to explain. The short horizontal extent of the steep gradients suggests a shallow source for the negative anomalies. Because post-Triassic sediments appear to simply thicken seaward, the source is believed to be a mass deficit occurring immediately below or at shallow depths below the smooth, strong reflector.

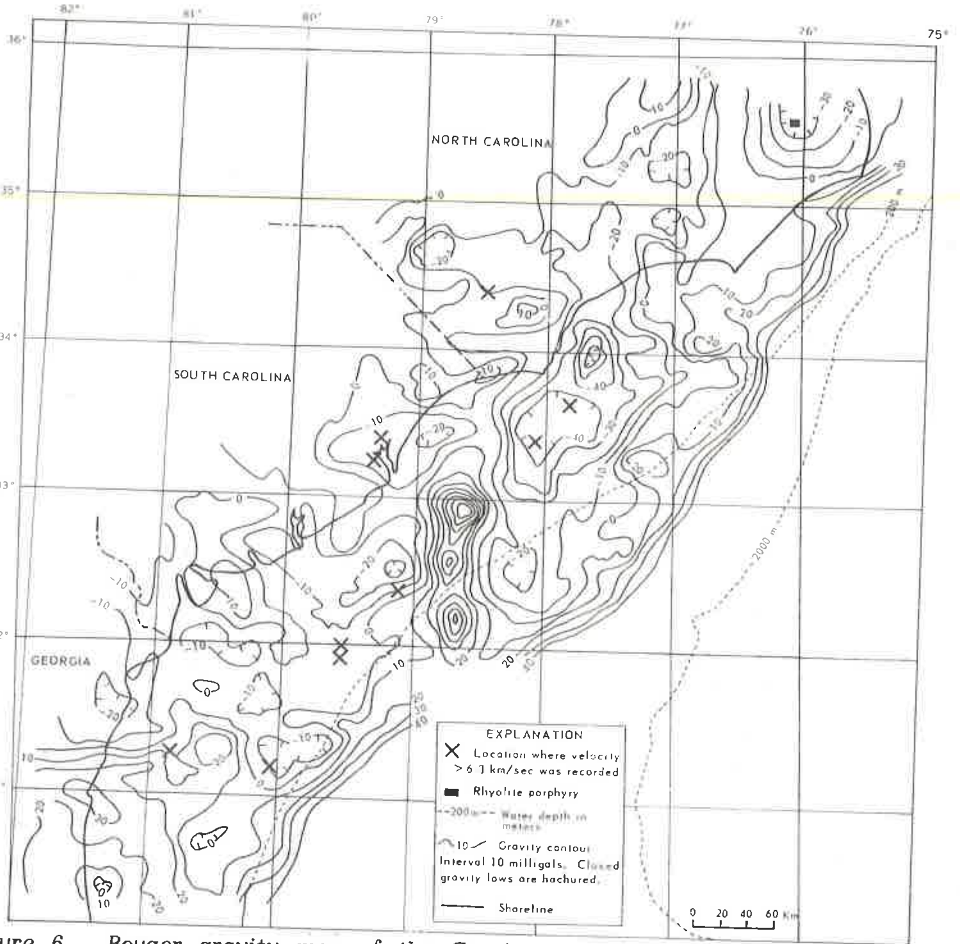

Figure 6. Bouger gravity map of the Continental Shelf in the Southeast Georgia embayment area. Figure is modified from Krivoy and Eppert (1977).

Sedimentary rocks or low density granitic rock could account for the mass deficit. Buried Triassic basins, containing chiefly nonmarine sediments, are present in the region. These include the Dunbarton Basin (Marine and Siple, 1974), the Florence Basin (MacCarthy, 1936; Bonini and Woollard, 1960) and the Durham-Wadesboro Basin (Randazzo and others, 1970; Ackermann and others, 1976).

Several onshore wells in coastal North Carolina have penetrated rhyolite basement (Maher, 1971; Brown and others, 1973). North of Cape Hatteras, on the northern shore of Albemarle Sound near 36°N, 76°W (Figure 6), Denison and others (1967) reported an altered rhyolite porphyry, which is apparently associated with a negative Bouguer gravity anomaly lower than -30 mgals.

To estimate the thickness of the low density rock the following assumptions were made. The surrounding basement was assigned a density of 2.73 gm/cc. This density, assigned to the Appalachian-Atlantic coast area by Woollard (1962), was chosen because it allows plausible density contrasts in calculations involving either the negative anomalies of -60 mgals or the positive anomalies of +80 mgals. An average density of 2.6 gm/cc was assigned to the rocks causing the low anomalies. This density is appropriate for either sedimentary rocks or low-density granitic rock (Daly and others, 1966). A thickness of about 7.5 km of 2.6 gm/cc rock is necessary to produce a negative anomaly of about -42 mgals, which, except for one small area, approximates the lowest negative anomaly values. If the density contrast were assumed to be 0.18, then the source of the anomaly would be about 5.5 km thick.

To explain the Georgetown, South Carolina, gravity low, Talwani and others (1975)

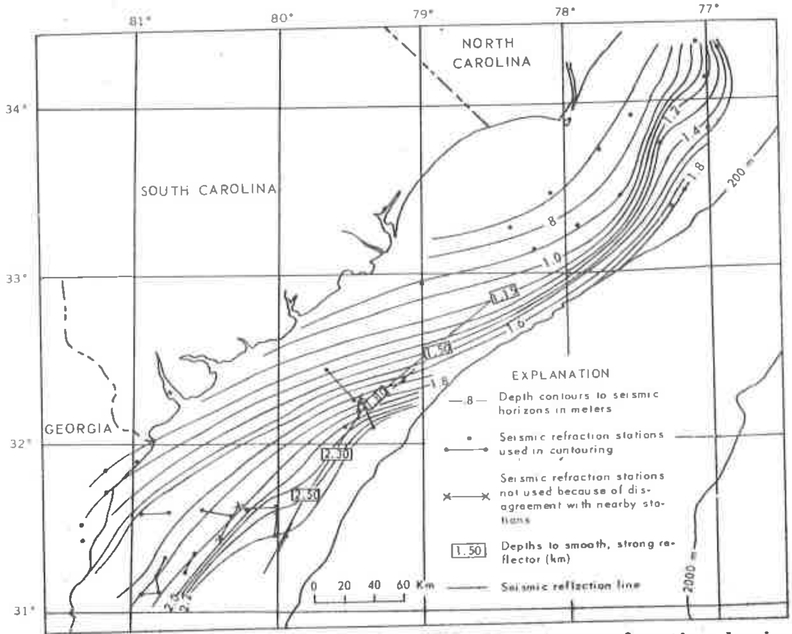

Figure 7. Contour map showing depth to 5.6-6.1 km/sec refracting horizon in the Southeast Georgia embayment area. Note the excellent agreement between depths to 5.6-6.1 km/sec horizon and the smooth, strong reflector recorded on USGS-IFP cruise.

proposed the presence of a sedimentary basin and calculated its thickness using a surrounding rock density of 2.67 gm/cc. Their estimate of density was based on that of Marine (1974) for the Dunbarton Basin. If we were to use their basement-density value, the source of the anomaly would have a thickness in excess of 15 km.

These thicknesses are unusually large for a sedimentary basin. The size of the low-density mass necessary to produce these large negative anomalies suggests that it is a granitic intrusion. On the other hand, the large numbers of Triassic (and/or Jurassic) basins found on- and off-shore along the east coast of North America (Marine and Siple, 1974) suggest the possible existence of others. Perhaps the low densities might be best explained by a combination of granitic intrusions and sedimentary basins.

CONCLUSIONS

Compilation of seismic reflection, seismic refraction and gravity data from the area between 32°N and 34°N latitude on the U.S. Atlantic Continental Shelf suggest that the pre-Cretaceous basement in the area is quite complex. The data obtained to date cannot be unequivocally interpreted. Many of the compressional wave velocities ranging from 5.6 to 6.1 km/sec, which have been recorded in seismic refraction surveys appear to represent a basalt flow or series of such flows of large areal extent. In places these flows may rest directly on Paleozoic rocks, but, particularly in areas of negative Bouguer gravity anomalies, they may overlie sediments.

Offshore from South Carolina two unusually strong negative anomalies suggest the presence of extremely thick Triassic-Early Jurassic basins, unusually low density granitic intrusions, or a combination of the two. High positive Bouguer gravity anomalies between the two areas of negative anomalies are believed to be caused by dense mafic intrusions.

ACKNOWLEDGMENTS

We are grateful to P. Popenoe and R.E. Sheridan for their reviews of, and suggestions for changes in this manuscript.

REFERENCES CITED

Ackermann, H.C., Bain, G.L., and Zohdy, A.A.R., 1976, Deep exploration of an east coast Triassic basin using electrical resistivity: Geology, v. 4, no. 3, p. 137-140.

Antoine, J.W., and Henry, V.J., Jr., 1965, Seismic refraction study of shallow part of continental shelf off Georgia coast: Am. Assoc. Petroleum Geologists Bull., v. 49, no. 5, p. 601-609.

Bonini, W.E., and Woollard, G.P., 1960, Subsurface geology of North Carolina-South Carolina Coastal Plain from seismic data: Am. Assoc. Petroleum Geologists Bull., v. 44, no. 3, p. 298-315.

Brown, P.M., 1974, Subsurface correlation of Mesozoic rocks in Georgia, in Stafford, L.P., Symposium on the petroleum geology of the Georgia coastal Plain: Georgia Geol. Survey Bull., v. 87, p. 45-59.

Brown, P.M., Miller, J.A., and Swain, F.M., 1973, Structural and stratigraphic framework and spatial distribution of permeability of the Atlantic Coastal Plain, North Carolina to New York: U.S. Geol. Survey Prof. Paper 796, 7 p.

Daly, R.A., Manger, G.E., and Clark, S.P., Jr., 1966, Density of rocks, sec. 4 of Handbook of physical constants (revised edition): Geol. Soc. America Mem. 97, p. 19-26.

Denison, R.E., Raveling, H.P., and Rouse, J.T., 1967, Age and descriptions of subsurface basement rocks, Pamlico and Albemarle Sound areas, North Carolina: Am. Assoc. Petroleum Geologists Bull., v. 51, no. 2, p. 268-272.

Dillon, W., Girard, O., Jr., Weed, E., Sheridan, R., Dolton, G., Sable, E., Krivoy, H., Grim, M., Robbins, E., and Rhodehamel, E., 1975, Sediments, structural framework, petroleum potential, environmental conditions, and operational cosiderations of the United States south Atlantic outer continental shelf: U.S. Geol. Survey Open-file report 75-411, 262 p.

Dillon, W.P., Paull, C.K., Buffler, R.T., and Fail, J.P., 1979, Structure and development of the southeast Georgia embayment and Northern Blake Plateau: Preliminary analysis: Geological and Geophysical Investigations of Continental Margins: Am. Assoc. Petroleum Geologists Mem. 29, p. 27-41.

Dowling, J.J., 1968, The East Coast onshore-offshore experiment, II. Seismic refraction measurements on the Continental Shelf between Cape Hatteras and Cape Fear: Seismol. Soc. America Bull., v. 58, no. 3, p. 821-834.

Gohn, G.S., Gottfried, D., Schneider, R.R., Lanphere, M.A., Higgins, B.B., Hess, M.M., Force, L.M., and Perlman, S.H., 1978, Preliminary report on the geology of two deep test holes, Clubhouse Crossroads #2 and #3, near Charleston, South Carolina (abs): Geol. Soc. America Abs. with Programs, v. 10, no. 4, p. 169.

Gohn, G.S., Higgins, B.B., Smith, C.C., and Owens, J.P., 1978, Preliminary report on the lithostratigraphy of the deep core hole (Clubhouse Crossroads Corehole #1) near Charleston, South Carolina, in Geological, geophysical and seismologica studies related to the Charleston, South Carolina, earthquake of 1886: A Preliminary report: U.S. Geol. Survey Prof. Paper 1028, p. 59-70.

Hersey, J.B., Bunce, E.T., Wyrick, R.F., and Dietz, F.T., 1959, Geophysical investigations of the continental margin between Cape Henry, Virginia, and Jacksonville, Florida: Geol. Soc. America Bull., v. 70, no. 4, p. 437-466.

Klitgord, K.D., and Behrendt, J.C., 1977, Aeromagnetic anomaly map of the U.S. Atlantic Continental Margin: U.S. Geol. Survey Misc. Field Studies Map MF-913, 2 sheets, scale 1:1,000,000.

Krivoy, H., and Eppert, H., 1977, Simple Bouguer anomaly representation over a part of the Atlantic Continental shelf and adjacent land areas of Georgia, the Carolinas, and northern Florida: U.S. Geol. Survey Open file report 77-316, 3 p.

McCarthy, G.R., 1936, Magnetic anomalies and geologic structures of the Carolina Coastal Plain: Jour. Geology, v. 44, no. 3, p. 396-406.

Maher, J.C., 1971, Geologic framework and petroleum potential of the Atlantic Coastal Plain and continental shelf: U.S. Geol. Survey Prof. Paper 659, 98 p.

Marine, I.W., 1974, Geohydrology of buried Triassic basin at Savannah River plant, South Carolina: Am. Assoc. Petroleum Geologists Bull., v. 58, no. 9, p. 1825-1837.

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

Randazzo, A.F., Swe, W., and Wheeler, W.H., 1970, A study of tectonic influence on Triassic sedimentation, the Wadesboro basin, central piedmont: Jour. Sed. Petrology, v. 40, no. 3, p. 998-1006.

Spivak, J., and Shelburne, O.B., 1971, Future hydrocarbon potential of Atlantic coastal province, in Cram, I.H., Sr., ed., Future petroleum provinces of the United States--their geology and potential: Am. Assoc. Petroleum Geologists Mem. 15, v. 2, p. 1295-1310.

Talwani, P., Ressetar, R., McAleer, J., Holmes, T., Grothaus, B., Findlay, M., Cable, M., and Amick, D., 1975, Gravity and magnetic profiles across the Georgetown gravity low: Carolina Div. Geology Geol. Notes, v. 19, no. 2, p. 24-32.

Taylor, P.T., Zietz, I., Dennis, L.S., 1968, Geologic implications of aeromagnetic data for the eastern continental margin of the United States: Geophysics, v. 33, no. 5, p. 755-800.

Woollard, G.P., 1962, The relation of gravity anomalies to surface elevation, crustal structure and geology: Univ. Wisconsin Research Rept. ser. no. 62-9.

Woollard, G.P., Bonini, W.E., and Meyer, R.P., 1957, A seismic refraction study of the sub-surface geology of the Atlantic Coastal Plain and continental shelf between Virginia and Florida: Univ. Wisconsin Tech. Rept., contract No. N7onr-28512, 128 p.

PETROLOGY OF THE BALSAM GAP DUNITE, JACKSON COUNTY, NORTH CAROLINA[1]

By

F. M. Honeycutt[2]

and

R. A. Heimlich
Department of Geology
Kent State University
Kent, Ohio 44242

ABSTRACT

The Balsam Gap dunite is enclosed in Precambrian gneiss composed of biotite-quartz-feldspar ± garnet and hornblende-quartz-feldspar ± garnet assemblages indicative of amphibolite facies metamorphism. Ground magnetic survey data and geologic mapping show the body to have maximum plan dimensions of 240 by 390 meters and a thickness of 150-170 meters. Its long axis is oriented roughly parallel to the strike of foliation in the enclosing gneiss which strikes NE-SW and dips steeply either SE or NW.

Mineralogically the rock is composed of olivine (Fo93) and chromite, altered in varying degrees to serpentine, talc, chlorite, tremolite, vermiculite, anthophyllite, and calcite. Typically the dunite texture is characterized by equant olivine grains (0.1-5 mm. in diameter) which meet at 120° triple-point junctions. Olivine also occurs as large irregular grains (3-6 mm in diameter) which commonly display undulatory extinction and strain bands. The amount of alteration varies throughout the body, but it generally increases from the center toward the margins with a maximum alteration of 40%. In conjunction with the alteration, there is a slight increase in silica and slight decrease in magnesia as the margins of the body are approached.

Petrofabric data indicate a preferred orientation of olivine crystallographic axes in one of two samples investigated. There is no direct relationship between these orientations and the shape of the body nor the attitude of foliation in the country rock.

Collectively, the data suggest that the Balsam Gap dunite was emplaced as a cold, solid mass of mantle material. Hydrous alteration of the dunite occurred during and/or after emplacement.

INTRODUCTION

Alpine ultramafic bodies have drawn considerable attention in recent years due to development of the plate tectonic theory and interest in the nature of the mantle. Such bodies are found in folded mountain belts near continental margins which are considered to be or to have been active plate boundaries. The ultramafic rock is thought to be upper mantle material which was emplaced in subduction (Coleman, 1971; Ragan, 1963; Chidester and Cady, 1972; Stevens et al., 1974) or obduction (Church and Stevens, 1971; Dewey and Bird, 1971; Williams, 1971; Laurent, 1975) zones.

---

[1] Contribution No. 187, Department of Geology, Kent State University, Kent, Ohio 44242

[2] Present address: Ohio Division of Geological Survey, Fountain Square, Columbus, Ohio 43224

The nature of the ultramafic material during emplacement has been debated by different workers. Some feel the bodies were emplaced as fresh peridotite which underwent serpentinization and other hydrous alteration during and/or after emplacement (Dribus et al., 1980; Hahn and Heimlich, 1977; Kingsbury and Heimlich, 1978; Ragan, 1963; Condie and Madison, 1969; Yurkovich, 1977; Coleman, 1971; and Dribus et al., 1976). Others believe that the material was emplaced as serpentinite which underwent partial or near total dehydration to peridotite after emplacement (Astwood et al., 1972; Carpenter and Phyfer, 1976; Vance and Dungan, 1977).

Many of the earlier studies of southern Appalachian ultramafites were directed primarily toward their economic potential with respect to chromite, vermiculite, asbestos, and olivine (Hunter et al., 1942; Murdock and Hunter, 1946; Conrad et al., 1963; Hunter, 1941). The more recent studies tend to be restricted to a single aspect such as geochemistry (Carpenter and Phyfer, 1976; Carpenter and Chen, 1978), petrofabrics (Astwood et al., 1972; Sailor and Kuntz, 1973; Bluhm and Zimmerman, 1977) or serpentinization (Condie and Madison, 1969; Alcorn and Carpenter, 1976; Swanson and Raymond, 1976).

The goals of this study of the Balsam Gap body were to apply field mapping, geophysics, petrography, petrofabric analysis, and geochemistry for the purpose of determining its three-dimensional shape and size, lithology, textures, fabric, and its mineral and chemical composition. Regarding earlier work, the Balsam Gap body was included in Hunter's (1941) reconnaissance study, two samples were studied by means of petrofabric analysis by Astwood et al. (1972), and several olivine samples were subjected to electron microprobe analysis by Carpenter and Phyfer (1976).

Financial assistance for this study was provided by the Society of the Sigma Xi. Mr. R. Henry and the International Minerals and Chemical Corporation gave permission to work on the Balsam Gap property and Mr. E.B. Conrad and the Parks family provided access to their lands nearby. Additionally, we are grateful to Donald van Buskirk, James Kluttz, and Steven P. Yurkovich for their assistance and advice during the field work.

LOCATION AND REGIONAL GEOLOGIC SETTING

The Balsam Gap ultramafic body is located in Jackson County, North Carolina approximately 65 kilometers WSW of Asheville (Figure 1) near the town of Balsam. It occurs within the Blue Ridge Province of the Appalachian Mountains, and is one of a large number of ultramafic bodies which define a discontinuous but persistent belt that extends from Alabama northeastward to Newfoundland.

In western North Carolina the ultramafic rocks occur as relatively small, subcircular to lenticular bodies typically associated with quartzofeldspathic ("Carolina") and mafic ("Roan") gneisses of Precambrian age (Misra and Keller, 1978). These rocks, as well as some of the ultramafic bodies, were intruded by granitic pegmatites of the Spruce-Pine type during middle Paleozoic time, suggesting that the ultramafics may have been emplaced during the Taconian (Late Ordovician) Orogeny (Misra and Keller, 1978).

METHODS

The body was mapped by the pace-and-compass method and by means of magnetic surveys (Honeycutt et al., 1980). An enlargement of the U.S.G.S. Hazelwood Quadrangle (1:24000 scale, 1941) was used as a base map (1 inch equals 250 feet) for the pace-compass data and for sample locations. Due to generally poor exposure in this area, a relatively small number of samples was obtained (Figure 2). An earlier open pit mine provided sampling access to the interior of the dunite.

A total of 29 thin sections was studied, including 18 samples of the ultramafic rock and 11 samples of the gneiss. Modes were obtained for each thin section based on counts of 1200 points minimum. In conjunction with standard petrographic analysis, a luminoscope was used to identify carbonate minerals and to determine the presence of potassium feldspar in the country rock. In addition, two oriented ultramafic

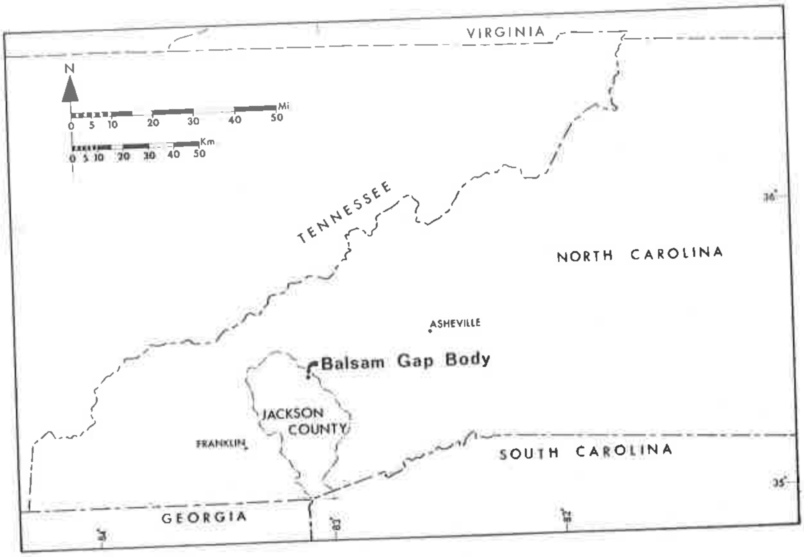

Figure 1. Location map.

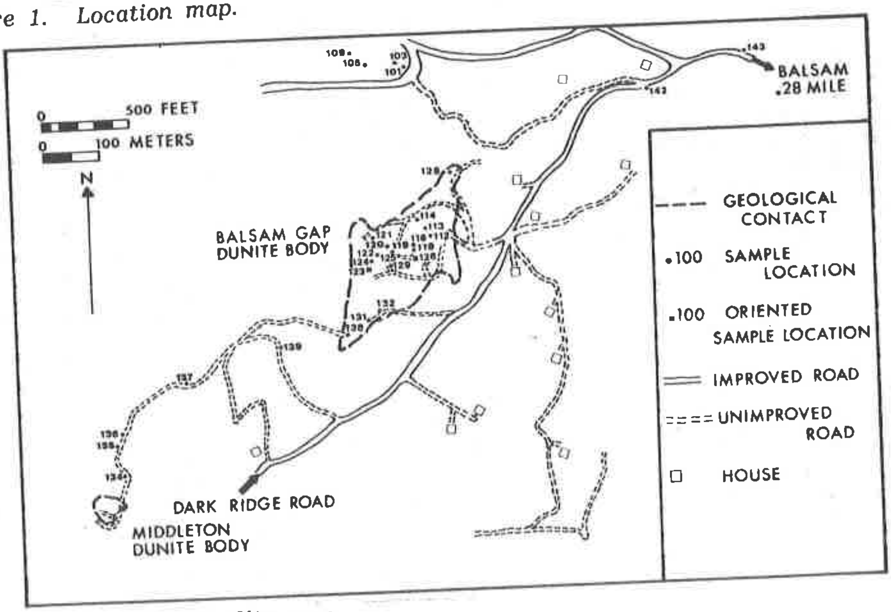

Figure 2. Sample locality map.

samples were collected and studied with a five-axis universal stage, following Emmons (1943), to determine if any preferred orientation of the olivine is present. For each section 150 grains were measured and the results were plotted on equal-area stereographic nets.

X-ray diffraction was used for mineral identification and to determine forsterite content of the olivine, within a maximum error of ±4% (Yoder and Sahama, 1957). Plagioclase compositions were estimated by the Michel-Levy method. Whole rock chemical analysis of 12 ultramafic samples was performed with a Perkin-Elmer Model 403 atomic absorption spectrophotometer for selected major and trace elements. Duplicate samples were analyzed, including those for whole rock standards DTS-1 and PCC-1, to determine precision (better than 4%). Accuracy was determined to be within 4% by comparing the values obtained for DTS-1 and PCC-1 with their accepted values.

COUNTRY ROCK

As is true throughout the southern Appalachian Mountains, the country rock in the Balsam Gap area is generally poorly exposed and typically highly weathered. A thick saprolite is the more common indicator of country rock rather than actual outcrops. Where exposed, the rock is clearly a gneiss composed of alternating layers of mafic and felsic minerals. Its foliation strikes generally northeast-southwest (Figure 3). The gneiss is composed of biotite-quartz-feldspar ± garnet and hornblende-quartz-feldspar ± garnet assemblages. Table 1 lists the minerals identified in thin section by means of their optical properties and cathodoluminescent characteristics.

The feldspar is primarily twinned plagioclase which occurs as subidioblastic grains possessing an average composition of An35 (range = An26-42). This is the dominant mineral in all but three samples of the gneiss (Table 1). With the exception of one sample, potassium feldspar is present only as inclusions in the plagioclase.

A major constituent in several samples, hornblende occurs as subidioblastic pleochroic grains (light to dark green) which exhibit well-defined cleavage and incipient alteration to fine-grained magnetite. Among the micas, biotite occurs as relatively large grains concentrated in bands while muscovite occurs as small scattered flakes. The biotite is strongly pleochroic (light to dark brown) and typically altered to magnetite along cleavage traces and grain boundaries.

Quartz, a prominent constituent of the gneiss (Table 1), is typically xenoblastic, finer grained than the feldspar, and characterized by undulatory extinction.

Accessory minerals include garnet, calcite, apatite, zircon, rutile, and magnetite. Garnet occurs as small polygonal prophyoblasts associated with the hornblende or biotite bands. Irregular calcite grains and idioblastic apatite prisms are scattered throughout the thin sections in small amounts. Zircon and primary magnetite are

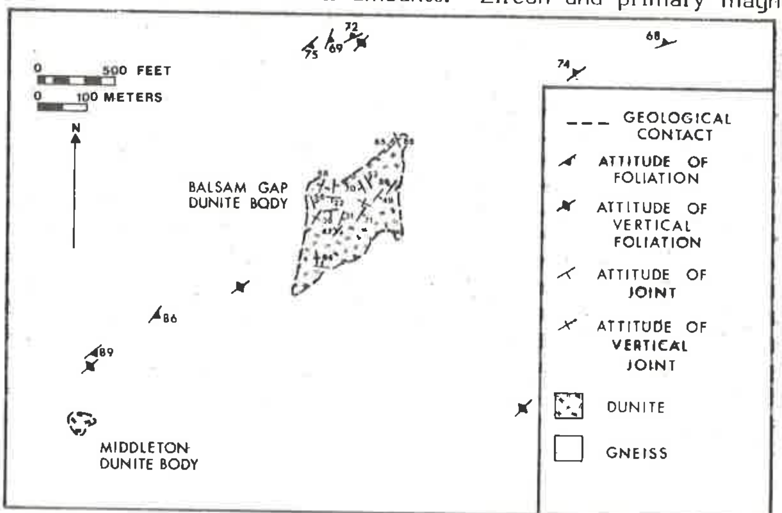

Figure 3. Geologic map.

Table 1. Modal analyses of the gneiss.

Sample Number	101	103	105	109	134	135	136	137	139	142	143
Plagioclase	52.6	43.6	34.4	60.5	32.0	62.2	46.9	50.6	34.9	55.3	48.3
Microcline					tr	3.8				tr	
Hornblende	16.2	12.4	41.0		59.5				54.4		
Biotite	7.8	14.2	7.9	10.1	0.6	6.6	14.8	16.2	tr	12.0	21.8
Quartz	21.6	28.2	10.9	28.2	4.1	27.0	36.2	31.3	5.2	28.8	25.7
Muscovite	0.4	0.4	3.5	0.5	0.2	tr	0.2	0.6	2.5	3.0	1.3
Magnetite	0.4	0.1	tr	tr	0.3	tr	0.2	0.1	tr	0.1	0.4
Garnet		0.1	0.2		0.5		0.4	0.1	0.2	0.1	1.3
Apatite	0.1	0.1	0.3	0.2	0.3	0.3	0.2	0.2	0.1	0.3	0.5
Zircon		tr		tr			tr	tr		tr	tr
Rutile	tr	tr			tr	tr	tr		tr		
Calcite	0.8	0.8	1.7	0.5	2.4	tr	1.0	0.7	2.6	0.3	0.6
Hematite		tr	tr				tr	0.1			

commonly associated with hornblende and biotite typically as inclusions in the latter. Rutile is present only in trace amounts associated with the hornblende and biotite. In the more weathered samples, hematite is present along cleavage traces and grain boundaries, and in some cases it discolors the feldspar.

Based upon the observed mineralogy, the country rock in this area lies within the amphibolite facies of regional metamorphism.

DUNITE

Field Characteristics

In plan view the Balsam Gap body measures 240 by 390 meters, as determined by field mapping and by a ground magnetic survey (Honeycutt et al., 1980). Its long axis is oriented N35E which is subparallel to the strike of foliation in the gneiss (Figure 3). In gross aspect the body appears to be concordant. However, lack of gneiss exposures near the contact prohibits a detailed comparison of the strike of the contact with that of the country rock. Modeling of the magnetic data suggests that the mass is pod-shaped and extends to a maximum depth of 150-170 meters (Honeycutt et al., 1980).

Megascopically, the rock varies from unaltered dunite to highly serpentinized dunite. The greatest degree of alteration by volume is generally along the margin, but alteration is also found at scattered locations throughout the body.

A network of fractures and joints characterize the dunite. Based on sparse data, there appear to be two major sets of joints, one striking northeasterly parallel to the long axis of the body, and another striking westerly to northwesterly roughly normal to its long axis (Figure 3). Dips range from 22° to 90°. Some of the fractures display slickensides and many are filled with serpentine or talc.

Although Astwood et al. (1972) make reference to the local presence of chromite lineations, none were found in the present study. Veins of calcite, quartz, and granitic pegmatite are also lacking in the dunite.

Petrography and Mineralogy

Petrographic study reveals that the primary minerals of the dunite are olivine and accessory chromite (Table 2). The olivine occurs characteristically as a mosaic of unstrained, polygonal grains possessing triple-point junctions which meet at approximately 120°. This texture is exhibited by all samples and it dominates each thin section. The polygonal grains range in diameter from 0.1 to 5.0 mm. The olivine also occurs as scattered irregular porphyroclasts which show deformational features

Table 2. Modal analyses and olivine composition for the dunite.

Sample Number	112	113	114	116	118	119	120	121	122
Olivine	55.5	60.9	66.5	92.4	92.1	85.2	97.5	72.9	95.2
Chromite	4.2	1.9	2.1	0.8	3.9	0.8	0.8	2.5	0.8
Serpentine	39.6	33.8	24.4	2.6	2.6	3.2	1.7	6.7	2.3
Talc	tr		4.2	3.8	0.6	9.3		13.9	0.6
Chlorite	0.7	3.3	1.9	0.4	0.5	1.2	tr	2.6	0.1
Tremolite			0.5			tr		0.9	0.9
Vermiculite	tr		0.2	tr	tr	0.1	tr	0.3	0.1
Anthophyllite			0.1	tr	tr	0.1		0.1	
Calcite					0.3	tr			
Olivine Fo%	92.1	90.7	91.8	95.1	93.5	93.4	95.2	90.3	93.3

Sample Number	123	124	125	126	128	129	131	132	138
Olivine	78.3	88.6	94.5	96.3	77.9	79.4	76.4	92.0	83.7
Chromite	1.1	0.6	0.8	0.6	0.4	0.7	0.9	1.3	1.1
Serpentine	3.4	2.2	4.1	2.0	3.4	5.4	5.9	5.9	2.9
Talc	12.8	8.3		0.5	15.7	9.9	14.9	0.1	10.1
Chlorite	3.0	0.1	tr	0.1	1.7	3.7	1.1	0.1	0.4
Tremolite	1.1				0.5	1.3	0.8	tr	1.7
Vermiculite	0.2	tr	0.1	tr	0.1		tr		tr
Anthophyllite		tr						0.2	
Calcite	0.1	0.1		0.5	0.2		tr	0.3	tr
Olivine Fo%	89.4	93.9	93.3	94.3	93.7	90.6	92.8	94.8	93.0

such as undulose extinction and strain bands. The porphyroclasts are typically larger (3-6 mm. in diameter) than the surrounding polygonal grains.

X-ray analysis of the olivine indicates that it is highly magnesian, averaging Fo92.9%, virtually the same as the compositions obtained by microprobe analysis (Fo 92.5%) by Carpenter and Phyfer (1976). Eighteen samples were analyzed and the results are consistently within the experimental error, suggesting the absence of olivine compositional zonation across the body.

Chromite occurs as disseminated euhedral grains which comprise 0.4 to 4.2% of the rock (Table 2). Its grain size ranges from 0.05 to 1.00 mm. Although one outcrop exhibits thin, irregular stringers of chromite, in general chromite occurs as scattered grains.

Modal analyses indicate that alteration of the dunite ranges from 1.7 to greater than 40%. In general the amount of alteration increases from the center of the body to its margins. Major alteration products, in decreasing order of abundance, are serpentine (chrysotile and antigorite), talc, chlorite, calcite, and tremolite. Traces of vermiculite and anthophyllite are present in addition. Chrysotile occurs as fracture fillings which transect olivine grains while antigorite occurs as irregular replacement patches along olivine grain boundaries. Both types of serpentine are characterized by disseminated, fine-grained secondary magnetite. Although associated with the olivine locally, chlorite has formed primarily at the expense of chromite. When all the alteration products are associated in a single thin section, there appears to be a sequence of alteration (chlorite replaced by serpentine, in turn replaced by talc) which is consistent throughout the body. Tremolite is present as a later stage of alteration, as indicated by its cross-cutting relationship with the other alteration products.

Chemical Data

Whole rock analysis of six representative samples of the dunite (Table 3) indicates that the pluton is relatively homogeneous with respect to the four major oxides: SiO_2, Al_2O_3, MgO, and Fe_2O_3. A slight increase in SiO_2 and a slight decrease in MgO are present locally as the margins of the body are approached. This suggestive zonation may reflect original chemical variations or may be the result of interaction between the dunite and the enclosing gneiss. The major oxide data are comparable to those for other North Carolina dunites (Hahn and Heimlich, 1977; Dribus et al., 1980; Kingsbury and Heimlich, 1978; Carpenter and Chen, 1978).

Nickel and chromium, present in trace amounts, fail to exhibit systematic variation within the body (Table 3). Average values are 2764 ppm nickel and 3953 ppm chromium. Although both values are higher than the averages for these elements in typical ultramafic rock, both fall within the range of values reported by Goles (1967). Both the nickel and chromite values exceed those in several other North Carolina ultramafites (Hahn and Heimlich, 1977; Kingsbury and Heimlich, 1978).

Petrofabric Data

Of the two oriented samples selected for petrofabric analysis, one (#126) is from very near the center of the body whereas the other (#123) was obtained 30 meters inward from its western contact (Figure 2). The texture of both samples is dominated by polygonal, unstrained olivine; less than 5% of the grains are strained porphyroclasts.

As shown in Figure 4, sample 126 shows an X point maximum (6%) with Y and

Table 3. Chemical analyses of the dunite.

Sample Number	112	114	118	120	121	123
SiO_2	45.39	45.99	42.23	41.19	46.09	45.43
MgO	41.91	42.35	48.46	51.32	43.39	43.21
Fe_2O_3*	8.45	7.79	8.10	7.35	7.35	7.84
Al_2O_3	0.35	0.46	0.15	0.03	0.82	0.98
Totals	96.10	96.59	98.94	99.89	97.65	97.46
Cr (ppm)	2541	3208	8044	3539	3576	3972
Ni (ppm)	2568	2582	3031	3071	2261	2092

*Total Fe as Fe_2O_3

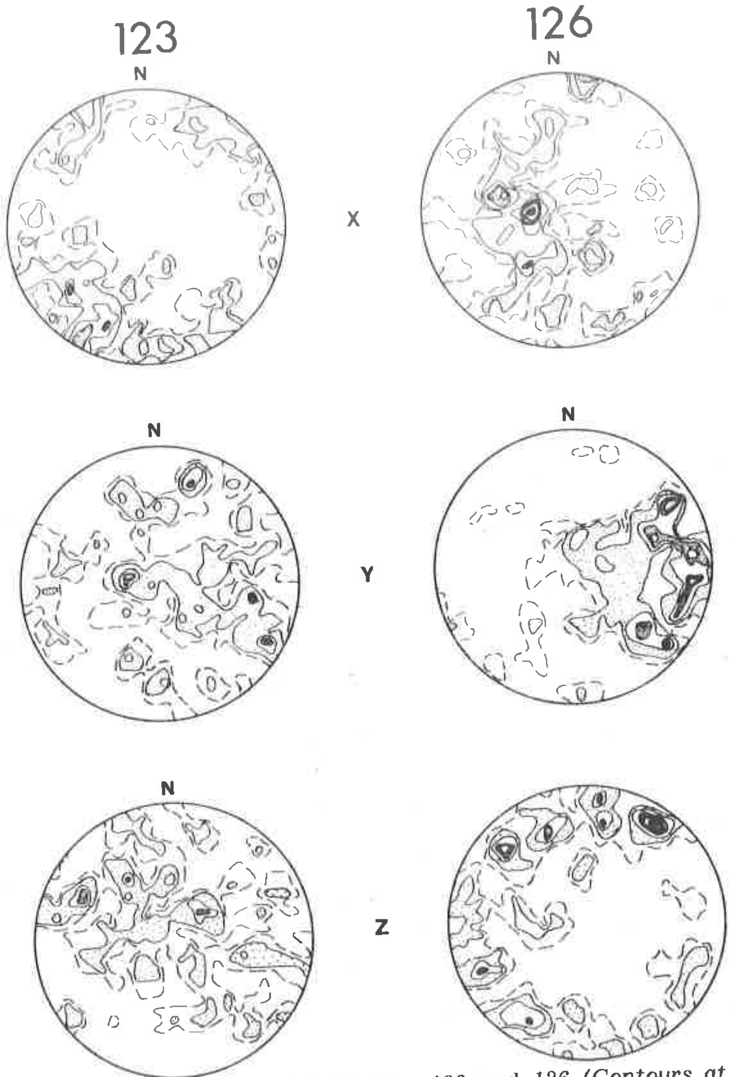

Figure 4. Petrofabric diagrams for samples 123 and 126 (Contours at 1, 2, 3, 4, 5, and 6%).

Z partial girdles normal to each other and normal to X. The olivine in sample 123 exhibits diffuse patterns for all three axes with maxima falling below 4% in all cases.

A comparison of these results with those of a previous petrofabric investigation (Astwood et al., 1972) suggests that their method of analysis, involving measurement of only 40 grains per thin section, may have failed to detect the presence of a preferred orientation of the olivine as is suggested by the 6% maximum (Ave' Lallemant, 1967) obtained in the present study.

CONCLUSIONS AND DISCUSSION

Field mapping combined with a ground magnetic survey (Honeycutt et al., 1980) indicates that the Balsam Gap dunite body is elongate and somewhat irregular in plan view and that it was emplaced in a terrain of gneisses which belong to the amphibolite facies of regional metamorphism. The long axis of the dunite is subparallel to the strike of foliation in the gneiss, suggesting that the dunite and the gneiss may be genetically related. However, the apparent absence of other stratigraphic units typical of ophiolite (Chidester and Cady, 1972; Dewey and Bird, 1971; Southwick, 1974) suggests that the dunite is not the ultramafic portion of an ophiolite sequence. Other data which support this are the absence of layering and megascopic preferred orientation of minerals in the dunite (Coleman, 1971).

Moreover, the magnetic data suggest that, in three dimensions, the dunite mass is pod-shaped and may plunge steeply northward (Honeycutt et al., 1980). Thus the body is not tabular, a form which might be expected if it were an ultramafic layer within an ophiolite sequence. The shape of the body supports the contention, based on the apparent absence of contact metamorphism, that it is an isolated pod of mantle material which was tectonically emplaced as a solid mass.

Modal analyses of the dunite suggest that serpentinization and other hydrous alteration increase generally toward the margins of the body. Such a pattern requires either alteration of dunite from the margins inward or dehydration of serpentinite from the center outward. Petrofabric data are consistent with, but do not prove, initial emplacement of solid dunite rather than serpentinite. Although the number of Balsam Gap samples is far less than that analyzed by Ave' Lallemant (1967) in the French Pyrenees, the maximum concentration obtained here in at least one of the two samples is the same number which Ave'Lallemant interpreted as indicating the presence of a preferred orientation.

Deformational features such as strain bands, undulose extinction and prophyroclasts surrounded by mosaics of smaller recrystallized olivine grains also argue for solid-state emplacement of the Balsam Gap dunite. Specifically, strain-banded porphyroclastic olivine grains are suggestive of incomplete recrystallization, movement, and deformation of a solid mass (Ragan, 1963; Nicolas et al., 1971; Mercier and Nicolas, 1975).

The prevalent polygonal texture of the dunite suggests recrystallization before (Ave'Lallemant and Carter, 1970), during (Ragan, 1969), or after (Spry, 1969) emplacement. Because this texture is identical to that in upper mantle peridotite tectonite inclusions in basalts (Mercier and Nicolas, 1975) and because it duplicates the experimental results of Ave'Lallemant and Carter (1970), we believe that the recrystallization occurred in the upper mantle source region prior to emplacement. The results of petrofabric analysis support this conclusion in that they show no direct relationship between preferred orientation of the olivine and the overall shape of the body nor the attitude of foliation in the country rock.

Petrographic data suggest three possible stages of dunite alteration. The first was characterized by extensive development of chlorite and serpentine, with the latter being the major alteration product. Its occurrence as cross-cutting veinlets and interstitial patches suggests alteration of olivine rather than dehydration of serpentine. If the latter was the case one might expect cross-cutting veinlets of secondary olivine in primary serpentine, as seen by Vance and Dungan (1977) in the Cascade Mountains of Washington. The second stage of alteration was steatization, characterized by the development of talc and some calcite. Textural relationships between the serpentine and talc-calcite suggest that the talc and calcite formed later. The third stage of alteration, amphibolization, is indicated by the presence of tremolite and trace amounts of anthophyllite and vermiculite, each of which transects olivine and minerals of the other two alteration stages. The most likely source of the hydrous fluids is the enclosing gneiss which represents eugeosynclinal sediments once rich in water.

The high magnesium content of the olivine and its large concentrations of nickel and chromium support the idea of a mantle origin for the dunite (Ragan, 1963; Goles, 1967).

The slight increase in silica and slight decrease in magnesia along the margins of the body appear to be the result of alteration of the dunite (See modes in Table 2). Similar results were noted by Hahn and Heimlich (1977), Palmer et al. (1977), Kingsbury and Heimlich (1978), and Dribus et al. (1980). The numerous joints and fractures in the dunite may have acted as avenues for the hydrous fluids during alteration of the dunite, which explains the presence of talc and serpentine fillings along these structures. Moreover, the absence of chromium and nickel zoning supports the theory of alteration zonation rather than a primary chemical zonation.

REFERENCES CITED

Alcorn, S.R., and Carpenter, J.R., 1976, Serpentinization of the Holcombe Branch Dunite: Southeastern Geol., v. 17, p. 243-256.

Astwood, P.M., Carpenter, J.R., and Sharp, W.E., 1972, A petrofabric study of the Dark Ridge and Balsam Gap dunites, Jackson County, North Carolina: Southeastern Geol., v. 14, p. 183-194.

Ave'Lallemant, H.G., 1967, Structural and petrofabric analysis of an "Alpine-type" peridotite: The lherzolite of the French Pyrenees: Leidse Geol. Medel., v. 42, p. 1-57.

Ave'Lallemant, H.G., and Carter, N.L., 1970, Syntectonic recrystallization and modes of flow in the upper mantle: Geol. Soc. America Bull., v. 81, p. 2203-2220.

Bluhm, C.T., and Zimmerman, J., 1977, Structure of the Frank ultramafic tectonite, Avery County, North Carolina (Abs.). Geol. Soc. America Abstracts with Programs, v. 9, p. 121.

Carpenter, J.R., and Phyfer, D.W., 1976, Olivine compositions from Southern Appalachian ultramafics: Southeastern Geology, v. 16, p. 169-172.

Carpenter, J.. and Chen, H.., 1978, Petrology and bulk geochemistry of the Frank ultramafic body, Avery County, North Carolina and associated other ultramafic rock bodies of the southern Appalacians: Southeastern Geol., v. 20, p. 21-25.

Chidester, A.H., and Cady, W.M., 1972, Origin and emplacement of alpine-type ultramafic rocks: Nature Phy. Science, v. 240, p. 27-31.

Church, W.R., and Stevens, R.K., 1971, Early Paleozoic complexes of Newfoundland Appalachians as mantle-oceanic crust sequences: Jour. Geophys. Res., v. 76, p. 1460-1466.

Coleman, R.G., 1971, Plate tectonic emplacement of upper mantle peridotites along continental edges: Jour. Geophys. Res., v. 76, p. 1212-1222.

Condie, K.C., and Madison, J.A., 1969, Compositional and volume changes accompanying progressive serpentinization of dunites from the Webster-Addie ultramafic body, North Carolina: Am. Mineral., v. 54, p. 1173-1179.

Conrad, S.G., Wilson, W.F., Allen, E.P., and Wright, T.J., 1963, Anthophyllite asbestos in North Carolina: North Carolina Dept. Cons. Devel., Bull. 77, 61p.

Dewey, J.F., and Bird, J.M., 1971, Origin and emplacement of the ophiolite suite: Appalachian ophiolites in Newfoundland: Jour. Geophy. Res., v. 76, p. 3179-3207.

Dribus, J.R., Godson, K.R., Hahn, K.R., Kammer, T.W., Lawton, D.L., Pesch, H.G., Schick, J.T., and Tata, S.S., 1976, Comparative petrography of some alpine ultramafic plutons in western North Carolina: Compass, v. 53, p. 33-45.

Dribus, J.R., Heimlich, R.A., and Palmer, D.F., 1980, The "Deposit No. 9" dunite, Macon County, North Carolina: Southeastern Geol., submitted.

Emmons, R.C., 1943, The universal stage: Geol. Soc. America Mem. 8, 205p.

Goles, G.G., 1967, Trace elements in ultramafic rocks in Wyllie, P.J., ed., Ultramafic and related rocks: John Wiley and Sons, N.Y., p. 352-362.

Hahn, K.R., and Heimlich, R.A., 1977, Petrology of the dunite exposed at the Mincy Mine, Macon County, North Carolina, Southeastern Geol., v. 19, p. 39-53.

Honeycutt, F.M., Heimlich, R.A., and Palmer, D.F., 1980, Magnetic survey of the Balsam Gap dunite, Jackson County, North Carolina: Southeastern Geol., submitted.

Hunter, C.E., 1941, Forsterite olivine deposits of North Carolina and Georgia, North Carolina Dept. Cons. Devel. Bull. 41, 177p.

Hunter, D.E., Murdock, T.G., and MacCarthy, G.R., 1942, Chromite deposits of North Carolina, North Carolina Dept. Cons. Devel. Bull. 42, 32p.

Kingsbury, R.H., and Heimlich, R.A., 1978, Petrology of an ultramafic body near Micaville, Yancey County, North Carolina: Southeastern Geol., v. 20, p. 33-46.

Laurent, R., 1975, Occurrences and origin of the ophiolites of southern Quebec, northern Appalachians: Canadian Jour. Earth Sci., v. 12, p. 433-455.

Mercier, J-C.C., and Nicolas, A., 1975, Textures and fabrics of upper-mantle peridotites as illustrated by xenoliths from basalts: Jour. Petrol., v. 16, p. 454-487.
Misra, C.K., and Keller, F.B., 1978, Ultramafic bodies in the Southern Appalachians: a review: American Jour. Science, v. 278, p. 389-418.
Murdock, T.G., and Hunter, C.E., 1946, The vermiculite deposits of North Carolina: North Carolina Div. Mineral Resources Bull. 50, 44p.
Nicolas, A., Bouchez, J.L., Boudier, F., and Mercier, J-C.C., 1971, Textures, structures, and fabrics due to solid flow in some European lherzolites: Tectonophysics, v. 12, p. 55-86.
Palmer, D.F., Heimlich, R.A., and Kolb, R.J., 1977, Petrofabric analysis of the Newfound Gap dunite, Haywood and Buncombe Counties, North Carolina (Abs.): Geol. Soc. America, Abstracts with Programs, v. 9, p. 304.
Ragan, D.M., 1963, Emplacement of the Twin Sisters dunite, Washington: American Jour. Sci., v. 261, p. 549-565.
Ragan, D.M., 1969, Olivine recrystallization textures: Mineralog. Mag., v. 37, p. 238-240.
Sailor, R.V., and Kuntz, M.A., 1973, Petrofabric and textural evidence for syntectonic recrystallization of the Buck Creek dunite, North Carolina (Abs.): Geol. Soc. America, Abstracts with Programs, v. 5, p. 791-792.
Southwick, D.L., 1974, Geology of the alpine-type ultramafic complex near Mount Stuart, Washington: Geol. Soc. America Bull., v. 85, p. 391-402.
Spry, A., 1969, Metamorphic Textures: Pergamon Press, New York, 305p.
Stevens, R.K., Strong, D.F., and Kean, B.F., 1974, Do some eastern Appalachian ultramafic rocks represent mantle diapirs produced above a subduction zone?: Geology, v. 2, p. 175-178.
Swanson, S.E., and Raymond, L.A., 1976, Alteration of the Day Book dunite, Spruce Pine district, western North Carolina (Abs.): Geol. Soc. America Abstracts with Programs, v. 8, p. 282.
Vance, J.A., and Dungan, M.A., 1977, Formation of peridotites by deserpentinization in the Darrington and Sultan areas, Cascade Mountains, Washington: Geol. Soc. America Bull., v. 88, p. 1497-1508.
Williams, H., 1971, Mafic ultramafic complexes in western Newfoundland Appalachians and the evidence for their transportation: A review and interim report: Geol. Asso. Canada Proc., v. 24, p. 9-25.
Yoder, H.S., Jr., and Sahama, T.G., 1957, Olivine X-ray determinative curve: American Min., v. 42, p. 475-491.
Yurkovich, S.P., 1977, The Corundum Hill dunite, Macon County, North Carolina: Southeastern Geol., v. 19, p. 55-68.

THE STRATIGRAPHY OF THE FLORIDAN AQUIFER SYSTEM EAST AND NORTHEAST OF LAKE OKEECHOBEE, FLORIDA

By

R. Thomas Mooney, III[1]
Department of Geology, Florida State University
Florida State University
Tallahassee, Florida 32306

Michael P. Brown
South Florida Water Management District
P.O. Box V
West Palm Beach, Florida 33402

and

Sherwood W. Wise, Jr.
Department of Geology
Florida State University
Tallahassee, Florida 32306

ABSTRACT

The geologic formations comprising the Floridan aquifer system in St. Lucie, Martin, and northern Palm Beach Counties, Florida, have been studied through lithologic analysis of cuttings and analysis of the geophysical signatures of the formations as shown on borehole geophysical logs. At least three units make up the Floridan aquifer system in the study area: an unnamed gray calcilutite, the Ocala Limestone of the upper Eocene, and the Avon Park Limestone of the upper middle Eocene.

The unnamed calcilutite is gray in color and contains varying amounts of quartz and phosphorite. Two members were observed in the Ocala Limestone. The upper member is a white coquina composed of foraminiferal tests such as *Lepidocyclina*, whereas the lower member is a cream colored bioclastic calcarenite containing smaller forams. The upper portion of the Avon Park Limestone is a white chalky calcilutite containing cone-shaped foraminifera such as *Dictyoconus cookei*. The lower portion consists of alternating beds of dolomitic limestones and dolomites.

A slightly undulating surface, dipping generally south to southeast, is developed on each of the three units of the aquifer system. No hard evidence was discerned for faults postulated in this area. An erosional surface at the top of the unnamed calcilutite may represent an Oligocene unconformity.

The unnamed calcilutite is less than thirty feet thick throughout most of the study area but reaches a thickness of 168 feet in easternmost St. Lucie County. More detailed study is needed to better define this unit. The Ocala Limestone is thickest along a linear feature trending NW-SE through St. Lucie and Martin Counties. Water wells penetrating the Floridan aquifer system along this trend record the highest composite transmissivities but yield the poorest quality water (highest composite total dissolved solids) within the study area.

INTRODUCTION

The term "Floridan aquifer" was originally defined on a regional basis to include the principal artesian aquifer which underlies the entire state of Florida. The Floridan

[1] Current address: Exxon Co., USA, 925 Common Street, New Orleans, Louisiana 70112

aquifer was said to include "parts or all of the middle Eocene (Avon Park and Lake City limestones), upper Eocene (Ocala limestone), Oligocene (Suwannee limestone), and Miocene (Tampa limestone, and permeable parts of the Hawthorn Formation that are in hydrologic contact with the rest of the aquifer)" (Parker, et al., 1955, p. 189). Brown and Reece (1979) have since found in the area of interest that the aquifer actually consists of a number of producing zones separated by semi-permeable zones, all of which they designate as the Floridan aquifer system.

The purpose of the present study is to determine which formations make up the Floridan aquifer system in St. Lucie, Martin and northern Palm Beach Counties, Florida, and to determine their general stratigraphic and structural relationships. The geophysical signatures of these formations are identified in an effort to establish a useful stratigraphic tool within the study area. By means of these signatures, the formations are mapped and their stratigraphic relationships interpreted. Correlations are then made between some of the geologic information so derived and aquifer data obtained by companion studies (Brown, 1979; Brown and Reece, 1979). Overall, our study is intended to aid in the understanding of the Floridan aquifer system in southeastern Florida and to provide a starting point for more detailed studies in the future.

METHODS

The study area covers some 1800 square miles (4662 square kilometers) of St. Lucie, Martin and northern Palm Beach Counties in southeastern Florida (Figure 1). Latitudes, longitudes, and total depths below mean sea level (MSL) of the wells studied are given in Table 1. Wells were drilled primarily by the percussion or hydraulic rotary drilling method. No cores were available. Cutting samples for many of the wells were collected at uniform depth intervals, usually every ten or twenty feet. Most of these cutting samples are on file with the Florida Bureau of Geology, Tallahassee, Florida. Wells referred to herein are designated by Florida Bureau of Geology ("W" series) or South Florida Water Management numbers. Sample descriptions (lithologies and the presence of certain key foraminifers) for five wells were given by Mooney (1979, Appendix). Mineral identifications and their approximiate percentages were determined by X-ray diffraction. Key foraminifers were identified using Applin and Applin (1944), Applin and Jordan (1945), and Loeblich and Tappan (1964).

Three types of geophysical logs were used in the study: natural gamma-ray log, neutron porosity log, and short and long normal electrical resistivity logs. The geophysical signatures of the various formations were identified by comparing the cutting samples from a reference well (SLF-3A) with geophysical logs of the well. SLF-3A was chosen for this purpose as the senior author aided in the collection of the cutting samples and felt confident as to the depths which the cutting samples represent. Geophysical logs and cuttings of other wells were compared to check these signatures. Credence was placed primarily in the geophysical logs since discrepancies were found in the depths recorded for some cutting samples. These discrepancies seem to have resulted from errors in estimating the depths from which the cuttings came, a problem common to deep well studies of this kind. The formations are identified primarily by means of the gamma-ray logs, with the neutron and electrical resistivity logs reinforcing identifications. The geophysical logs used are lists by well in Table I and are on file at the Florida Bureau of Geology. The depths obtained from the geophysical logs were corrected to mean sea level (MSL). This information was then synthesized into cross-section, structure contour maps and isopach maps.

PREVIOUS STUDY AND EVOLUTION OF FORMATIONAL CONCEPTS

The stratigraphy and general geology of south Florida have been subjects of debate since the first attempts were made to explain the geologic development of Florida's peninsula. Agassiz (1852) and LeCount (1857, 1878) hypothesized that much of the Florida peninsula was constructed of Recent coral reef debris; however, a study of fossil mollusks by Heilprin (1887) showed that the peninsula was neither Recent in

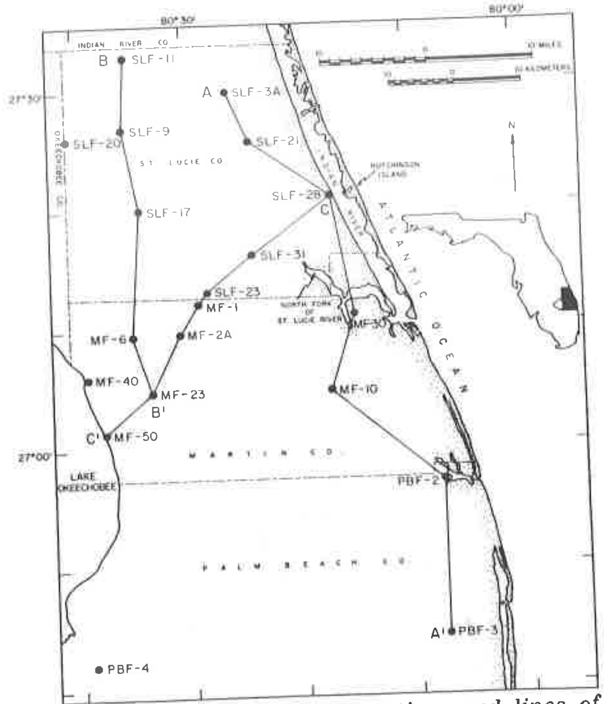

Figure 1. Map of study area showing well locations and lines of cross-section.

age nor formed of coral reef deposits.

Close similarities in the lithologies of Florida's limestones led early workers to subdivide these limestones into formations primarily on the basis of paleontological data. Initially, all of the older Tertiary rocks of the peninsula were included in the Vicksburg Limestone by Smith (1881). This included all limestones older than the greenish clay sequence presently regarded as the Hawthorn Formation. The Vicksburg Limestone of Smith was later divided by Dall and Harris (1892) on paleontological grounds. They distinguished a lower division which they called the Vicksburg Limestone, or *Orbitoides* limestone, and an upper division which they named the Ocala Limestone. The Ocala Limestone was distinguished as a yellow, friable limestone with many foraminifera, especially in the form of "nummulitic beds."

The Tampa Limestone was also named by Dall and Harris (1892). It was described as a hard white, earthy limestone that underlies the town of Tampa. The Tampa Limestone is a controversial formation recently restudied by King and Wright (1979). It is generally regarded as an impure limestone, usually in the form of a calcilutite. The quartz sand content of the Tampa as reported in the literature varies greatly. The range of lithologic characteristics attributed to the Tampa is sufficiently wide that almost any limestone could be regarded as the Tampa Limestone using published criteria. The age of the Tampa is also debatable. However, most workers seem to feel that it is late Oligocene, early Miocene, or both.

Dall (1903) proposed the replacement of the name Vicksburg Limestone with the term Peninsular Limestone. The names Peninsular Limestone and Ocala Limestone were adopted by Matson and Clapp (1909) and later by Matson and Sanford (1913). They followed Dall in regarding the Ocala Limestone as younger than the Peninsular Limestone although they considered both limestones to be in the Vicksburg Stage (Oligocene).

Cooke (1915) found that much of the Peninsular Limestone was identical to the Ocala Limestone and that the fauna of the Ocala is of the Jackson Stage (upper Eocene). Cooke also suggested that the term "Peninsular Limestone" not be used until the relationship between the Ocala Limestone became better understood. He listed the formations older than the Ocala as being buried and presumably unworkable at the time. This view of south Florida's older limestones is again reflected in Stringfield's

Table 1. Well Locations, Depth and Geophysical Logs Available for Study.

Well #	Depth (feet below MSL)	Latitude	Longitude	-ray	Neutron	Electrical Resistivity
SLF-3A (W-13850)	1200	27°29'48"	80°26'48"	X	X	X
SLF-9	1033	27°26'50"	80°35'28"	X	X	X
SLF-11	921	27°32'12"	80°35'11"	X	X	X
SLF-17	1260	27°20'14"	80°34'18"	X	X	X
SLF-20	867	27°26'04"	80°40'40"	X	X	X
SLF-21	686	27°25'37"	80°24'09"	X	X	X
SLF-23	868	27°13'11"	80°28'11"	X		X
SLF-28	852	27°20'28"	80°16'35"	X		
SLF-31	982	27°16'14"	80°23'50"	X	X	X
MF-1	801	27°14'12"	80°29'00"	X	X	X
MF-2A	852	27°09'37"*	80°30'39"*	X		X
MF-6	1017	27°09'39"	80°35'00"	X	X	X
MF-10	970	27°04'32"	80°17'23"	X	X	
MF-23	1089	27°04'25"	80°33'47"	X	X	X
MF-30 (W-12556)	2990	27°11'46"	80°15'02"	X		
MF-40 (W-5441)	992	27°05'58"*	80°39'27"*	X		X
MF-50 (W-5442)	997	27°01'22"*	80°38'04"*	X		X
PBF-2	1337	26°56'42"	80°07'23"	X		
PBF-3 (W-13000)	3314	26°44'21"	80°07'32"	X		
PBF-4 (W-8077)	2063	26°42'20"*	80°38'45"*	X		

*approximate values; accurate to the nearest minute

(1933) study of the ground water in the Lake Okeechobee area. He reported the Ocala Limestone in well cuttings from the area, but was uncertain as to the bottom of the Ocala. Stringfield, therefore, referred to all limestones from the top of the Ocala down to the bottom of the wells as being part of the Ocala Limestone. Stringfield (1936, p. 125) wrote, "The lithology of the Ocala and the underlying Eocene rocks is similar, and it is therefore necessary to distinguish the two units on the basis of a study of fossils collected from the well cuttings. No diagnostic fossils have been reported near the contact and the lower limit of the Ocala has therefore not been definitely determined."

At that time it was believed that no representative of the Vicksburg Stage occurred in the Florida peninsula (Stringfield, 1936). However, within the same year, Cooke and Mansfield (1936, p. 71) proposed the name "Suwannee Limestone" for a yellowish limestone exposed along the Suwannee River in Florida. They felt that this limestone should be included in the upper Vicksburg Stage on the basis of its contained fossils. The Suwannee is presently described as being a rather pure limestone composed primarily of limy particles of organic origin. Small amounts of quartz sand (<10%) may be present (Cooke, 1945, p. 86).

Based on cuttings and data obtained from deep oil wells, Applin and Applin (1944) mapped and described the stratigraphy of the entire state of Florida as well as southern Georgia. Using lithologic and faunal differences, they were able to subdivide the subsurface limestones of south Florida and, in the process, they named several new formations. Thus, they succeeded in establishing Florida's first good subsurface stratigraphic column from the Vicksburg Stage of the Oligocene Series down to the Lower Cretaceous, and this study provided the framework for all future stratigraphic work in the state.

Applin and Applin (1944) were able to separate the Ocala Limestone into upper and lower members. The upper member is the characteristic Ocala consisting of a soft, white, foraminiferal coquina (nummulitic beds). The lower member is a cream-colored calcarenite which is harder than the upper member and composed largely of molds of small miliolids. Unlike the upper member, the lower member contains very few large foraminifera. This concept of the Ocala is still used by the U.S. Geological Survey and can be applied in our study area.

The Avon Park Limestone, one of the new formations named by Applin and Applin (1944), is recognized in the present study. The Avon Park Limestone is described as a cream-colored, highly fossiliferous, chalky limestone containing interbedded sub-crystalline dolostones. It is also distinguished from the overlying Ocala Limestone by a difference in microfauna. The Avon Park contains many cone-shaped foraminifera such as *Coskinolina* and *Dictyoconus* (Applin and Applin, 1944). Applin and Applin listed the age of the Avon Park as late middle Eocene.

Subjacent to the Avon Park in Florida's peninsula is the Lake City Limestone, also named by Applin and Applin (1944). However, because only three wells used in the present study are deep enough to penetrate the Lake City, and since the geophysical logs of these wells are not of a high quality, the Lake City Limestone is not discussed in this study.

The water resources of Palm Beach County were investigated by Schroeder, Milliken and Lowe (1954). They listed the formations comprising the principal artesian aquifer as the lower Hawthorn limestones, and the Tampa, Suwannee, Ocala and Avon Park Formations. This aquifer system was then named the "Floridan aquifer" in the comprehensive study of the water resources of southeastern Florida prepared by Parker, et al. (1955).

The geology of Martin County was studied by Lichtler (1960) using well cuttings and electric logs. He postulated, in the eastern part of the county, a "major subsurface fault having a displacement of 300 to 400 feet and a strike that is approximately parallel to and about five miles inland from the coastline" (Lichtler, 1960, p. 16). Movement along the fault was theorized to have started in late to post Oligocene time and to have continued during early Miocene time, when the Suwannee Limestone was exposed and eroded. The fault was considered a scissorstype fault with the greatest development at the southern end. Lichtler showed the fault extending from Martin County into adjacent St. Lucie and Palm Beach Counties.

In a study based primarily on electric logs, Chen (1965) described the lithostratigraphy of the Paleocene and Eocene rocks of Florida. He presented his findings in the form of numerous structure maps, isopach maps and lithofacies maps. Chen's study was a broad regional study as evidenced by the fact that only five of his control points occur in all of St. Lucie, Martin and Palm Beach Counties.

Using hydrologic data, Vernon (1970, p. 7) further extended Lichtler's fault northward into St. Lucie County and southward through northern Palm Beach County. He showed the fault as intersecting the coastline somewhere near the city of West Palm Beach. Vernon used the top of the artesian aquifer as his measuring point rather than any particular geologic unit.

In 1975, Law Engineering Testing Company conducted a study of Hutchinson Island, St. Lucie County, for the purpose of locating a nuclear power plant (Anonymous, 1975a). During this study three deep geologic borings were drilled along a line west of the proposed plant site in order to obtain data on both sides of Vernon's (1970) extension of Lichtler's (1960) fault. In addition, seismic surveys were conducted along the Indian River and up the north fork of the St. Lucie River just north of the Martin County - St. Lucie County line (Figure 1). These surveys crossed the fault hypothesized by Lichtler. Utilizing the data obtained, Law Engineering concluded that no fault was present but that warping (folding) was responsible for the offset in marker beds.

A report on the drilling and testing of deep disposal and monitoring wells for the city of Stuart in Martin County was produced by another engineering company, Black, Crow and Eidseness, Inc. (Anonymous, 1975b). In this study, another fault was postulated, based on the offset of key beds indicated on gamma-ray logs, as lying parallel to and just west of Lichtler's (1960) fault. The configuration and timing of movement along this fault are similar to those suggested for Lichtler's (1960) fault. A brief description of the core taken from the Stuart disposal well is given in the appendix of Black, Crow, and Eidseness, Inc.'s 1975 report. The Stuart monitor well (W-12556) is used as a control point in the present study.

STRATIGRAPHY

Lithostratigraphy

As previously mentioned, well SLF-3A was chosen as the primary reference well for the study area. The stratigraphic section in this well, which is considered characteristic for the study area, is shown in Figure 2 and described below.

The base of the Hawthorn Formation occurs within the 483-503 foot (147-153 m) sample interval in SLF-3A, and is marked by a thin bed of unconsolidated sand. The sand is a mixture of phosphorite and quartz grains, and sand-sized fragments of chert and limestone. Traces of dolomitic limestone are also present. The thickness of the sand bed is probably less than the twenty foot sample interval as evidenced by the amount of limestone present.

The next lithologic unit encountered downhole is a gray calcilutite, similar to that found in the sand unit above, and about 20-30 feet (6-9 m) thick. Minor silt-sized phosphorite and traces of quartz sand are found in this limestone. No microfossils have been observed in this unit.

Below the gray calcilutite is a white foraminiferal coquina composed of large nummulitic genera such as *Lepidocyclina*. Some bryozoans and bioclastic debris are found in this coquina along with other foraminifera such as camerinids. This foraminiferal coquina is about 60 feet (18 m) thick.

Below the foraminiferal coquina is a cream-colored bioclastic calcarenite or "grapestone" type of limestone. Larger foraminifera are absent except as small fragments. However, smaller forms such as camerinids are still present and seem to be more abundant. Traces of dolomitic limestone and recrystallized limestone are also

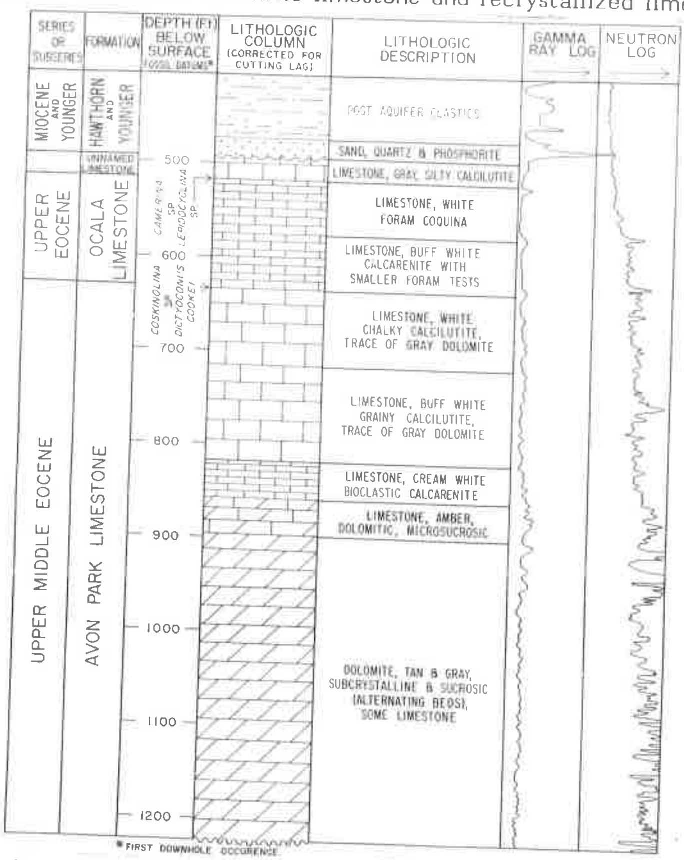

Figure 2. Detailed stratigraphic columnar section of well SLF-3A which is used herein as the reference section for the study area.

present. The bioclastic clacarenite is about 60 feet (18 m) thick.

The next lithologic unit encountered downhole is a white chalky clacilutite containing traces of dolomitic limestone. Occurring with this unit are cone-shaped foraminifera such as *Coskinolina*, and *Dictyoconus*. The thickness of this unit is about 80 feet (24 m).

Below the chalky calcilutite is a cream-colored grainy calcilutite. This gray calcilutite grades downward into a calcilutitic calcarenite which appears to be bioclastic in part. There seems to be an increase in the diversity of foraminifera, but the forams are too abraded and recrystallized to be accurately identified. There is also some evidence of recrystallized limestone and dolomite. The combined thickness of the gray calcilutite and the bioclastic calcarenite is about 140 feet (43 m).

Below these limestone units is a dolomitic limestone, amber in color and microsucrosic in texture. Dolomites extend to the total depth of this well and consist of alternating layers of sucrosic textured amber dolomites and massive, subcrystalline, grey dolomites.

These lithologic units in SLF-3A are representative of the stratigraphic sections in other wells where the cuttings were described (see Mooney, 1979, Appendix), and most of the units can be traced throughout the study area although depths of occurrence and thicknesses may vary.

Formations Encountered

As previously mentioned, the base of the Hawthorn Formation is readily recognized in the cutting samples, and corresponds to a sand unit with the highest concentration of phosphate found in the well. The sand is composed primarily of rounded, polished phosphorite and quartz grains. As stated above, in well SLF-3A the base of the Hawthorn occurs between 483 and 503 feet (147-153 m) below land surface.

The next lithologic unit is the gray, sandy calcilutite, which occurs between 503 and 523 feet (153-159 m) in well SLF-3A. The formational name and age of this unit are not known, as no microfossils or other diagnostic criteria have been observed by the authors. This limestone fits neither the usual definitions for the Suwannee nor Tampa limestone, although it could be considered closer to the Tampa (due largely to the Tampa's vague definition).

The white, foraminiferal coquina below the gray calcilutite corresponds to the upper member of the Ocala Limestone as described by Applin and Applin (1944). The foraminifera found here are large nummilitic types and therefore correspond to Dall and Harris's (1892) "nummulitic beds." The next unit down, the cream-colored, bioclastic calcarenite, or "grapestone" is probably the lower member of the Ocala as described by Applin and Applin (1944). In well SLF-3A, the Ocala has a thickness of about 120 feet (37 m). The Ocala Limestone has been placed in the Jackson Stage of the Eocene Series (Cooke, 1915).

Below the Ocala Limestone in well SLF-3A is a white, chalky clacilutite, containing cone-shaped foraminifera such as *Coskinolina* and *Dictyoconus*. This fits the description of part of the Avon Park Limestone of Applin and Applin (1944). Based on the cutting samples, the top of the Avon Park in SLF-3A is placed at 643-663 feet (196-202 m). The well apparently ends in the Avon Park Limestone, therefore, its thickness cannot be determined. Applin and Applin (1944) placed the Avon Park Limestone in the Claiborne Stage (upper middle Eocene).

Comparison of Formations and Geophysical Logs

The geophysical signatures of the formations discussed above were recognized by comparing the well cuttings and geophysical logs of well SLF-3A. This comparison is shown in Figure 2. The lithologic log obtained from examination of the cuttings from well SLF-3A was adjusted to correspond with changes in the geophysical logs. Characteristic geophysical signatures of the various formations then became evident. Before these geophysical signatures are described, a brief mention of the types of logs used is in order.

The natural gamma-ray log is the log which gave the most satisfactory correlation results and is the only geophysical log common to all the wells used in the present study. The gamma-ray log basically is a measurement of the natural radioactivity of the various formations. Since phosphates in Florida contain uranium, the gamma-ray log effectively shows the presence of phosphate. To a lesser extent, it can also mark the presence of clay units and in some cases dolomites, depending on their concentrations of radioactive elements.

The neutron porosity log responds most directly to hydrogen concentration and can be used to determine the porosity of a formation. It thus provides a useful check on units in the Florida aquifer system. Neutron logs are especially useful in delineating the deeper dolomites, many of which are sub-crystalline in texture and therefore have little porosity.

Electrical resistivity logs were used to delineate formations in wells where no neutron log was available. Although the resistivity log is highly affected by the quality of the water (fresh or saline) contained in a formation, it seemed to delineate the dolomites quite well because of their high resistivities.

The base of the Hawthorn Formation is easily recognized on the gamma-ray log. It corresponds with the deepest, and usually strongest, sharp gamma-ray intensity peak. This is due to the increase in rounded phosphorite grains which in well SLF-3A (Figure 2) occurs at about 488 feet (149 m).

Below the base of the Hawthorn, there is a 30-35 foot (9-11 m) interval of moderate radiation that occurs before the next good gamma-ray "kick." This zone corresponds to the unnamed gray calcilutite that commonly contains some quartz and minor phosphorite.

The next noticeable unit on the gamma-ray log is an interval of relatively low gamma-ray intensity. This unit is approximately 110 feet (34 m) thick in well SLF-3A and corresponds to the Ocala Limestone. On the gamma-ray log for well SLF-3A there is a peak of slightly higher intensity in the middle of this limestone where cuttings indicate the contact between the upper and lower Ocala. This division, however, is not everywhere evident; therefore the separation of the Ocala Limestone cannot always be made. The lower part of the Ocala shows an increase in the neutron log intensity. This may reflect the harder and less porous character of the lower Ocala.

Below the Ocala Limestone there is a noticeable increase in the background intensity of the gamma-ray log. This increase corresponds to the top of the Avon Park Limestone. The increase in the gamma-ray background intensity may be due to the dolomitic character of the Avon Park. Deeper in the Avon Park the dolomite beds can easily be recognized by strong intensity peaks on the neutron log. A slight increase in the gamma-ray log intensity is also associated with the dolomites. The electrical resistivity logs for other wells are also useful in delineating these dolomite beds.

STRATIGRAPHIC-STRUCTURAL INTERPRETATIONS AND CORRELATIONS

Using geophysical criteria, the formations comprising the Florida aquifer system were delineated in all wells used in this study. Table II lists the depths to the tops of these formations in the wells and the thicknesses for two of them: the unnamed gray calcilutite and the Ocala Limestone. The data in Table II were used to construct structure contour maps for the base of the Hawthorn Formation (Figure 3), the top of the Ocala Limestone (Figure 4), and the top of the Avon Park Limestone (Figure 5). Isopach maps for the unnamed gray calcilutite and the Ocala Limestone were constructed and are shown on Figures 6 and 7, respectively.

An examination of the structure contour maps shows a similarity between the tops of the unnamed calcilutite (bottom of Hawthorn), Ocala Limestone and Avon Park Limestone. All these units have a southerly or southeasterly general dip, and in northern and central St. Lucie County, they show a regular, uniform surface. However, in southern St. Lucie County and in Martin County, their surfaces are more irregular. A small, high area is present in east central Martin County on all three maps. The existence of this high area is probably the source of most of the faults proposed in this

*Table 2. Subsurface Data Used to Construct the Structure Contour and Isopach Maps.**

Well #	Base of Hawthorn	Top of Ocala	Top of Avon Park	Unnamed ls. thickness	Ocala thickness
SLF-3A (W-13850)	465	501	609	36	108
SLF-9	446	477	556	31	79
SLF-11	392	412	504	20	92
SLF-17	564	598	742	34	144
SLF-20	484	515	627	31	112
SLF-21	443	510	634	67	124
SLF-23	602	625	734	23	109
SLF-28	575	743	794	168	51
SLF-31	697	702	773	5	71
MF-1	637	671	757	34	86
MF-2A	601	633	718	32	85
MF-6	671	697	748	26	51
MF-10	628	645	779	17	135
MF-23	729	741	835	12	94
MF-30 (W-12556)	772	784	814	12	30
MF-40 (W-5441)	685	698	760	13	62
MF-50 (W-5442)	698	724	759	26	35
PBF-2	1079	1085	1197	6	112
PBF-3 (W-13000)	772	-	-	-	-
PBF-4 (W-8077)	801	831	933	30	102

*(Depths in feet below MSL)

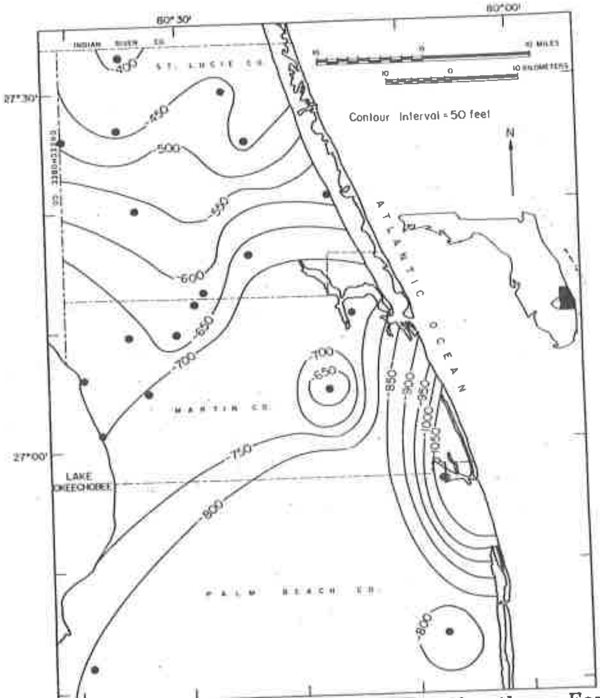

Figure 3. Structural contour map of the base of the Hawthorn Formation. Elevations are in feet below mean sea level.

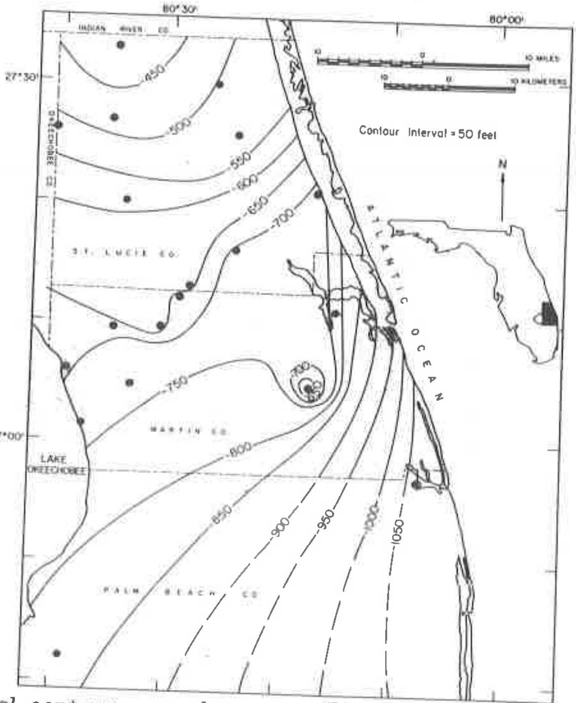

Figure 4. Structural contour map of the top of the Ocala Limestone. Elevations are in feet below mean sea level.

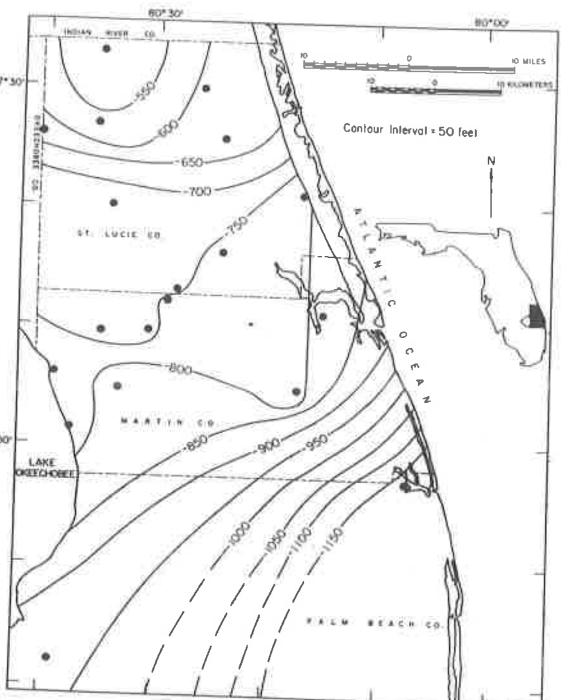

Figure 5. Structural contour map of the top of the Avon Park Limestone. Elevations are in feet below mean sea level.

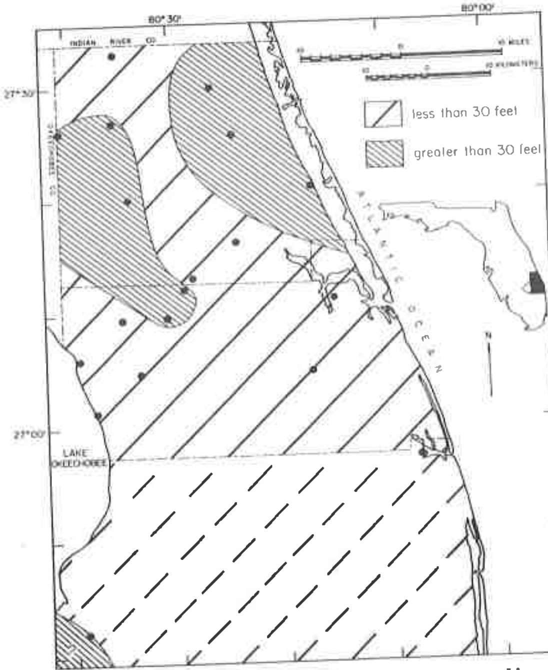

Figure 6. Isopach map of the unnamed calcilutite that overlies the Ocala Limestone.

area. It is our opinion, however, that this high probably is not caused by faulting but by some other means. This high area may represent only an erosional feature, or it may have been produced by tectonic activity such as warping, as suggested in the Law Engineering report (Anonymous, 1975a).

A structural low is shown at the base of the Hawthorn Formation in southeastern Martin and northeastern Palm Beach counties (Figure 3). It is not known whether this feature is present on the top of the Ocala or Avon Park Limestones. The gamma log on well PBF-3 is not of good quality and these units cannot be picked in this well with confidence.

The thickness of the unnamed gray calcilutite is plotted in two categories: greater than thirty feet and those less than thirty feet (Figure 6). The areas where the thickness is greater than thirty feet are located in eastern and southwestern St. Lucie County, with an indication that another thick area may be present in the southwesternmost corner of the study area.

The unnamed calcilutite has not yet been assigned to a specific formation although several possibilities exist. Minor phosphorite in the limestone suggests that it may be considered a lower unit of the Hawthorn Formation. If this is true, then the phosphorite sand used in the present study to mark the base of the Hawthorn would have to be moved up into the middle of the Hawthorn rather than marking its base. This would also place the Hawthorn Formation in direct contact with the Ocala Limestone. An alternative interpretation is that the calcilutite represents part of the Tampa Limestone. This is possible if the phosphorite in the calcilutite is the result of downhole contamination. Although the calcilutite does not closely resemble the Tampa at its type locality, the two would be similar if the phosphorite were not present (K.C. King, 1979, personal communication).

Another possibility is that the calcilutite is a facies of the Suwannee Limestone. The calcilutite, however, shows little resemblance to the Suwannee as originally described by Cooke and Mansfield (1936).

A fourth alternative is that the unnamed calcilutite does not belong to any of the above formations, but represents a new formation in itself. If this is indeed the case, then it needs to be formally named, described and mapped, an endeavor that is beyond the scope of this paper.

If one assumes that the phosphorite sand that is called the base of the Hawthorn

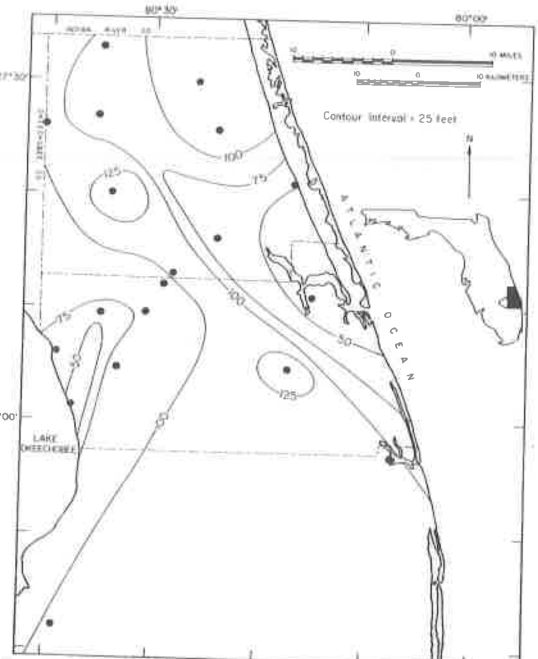

Figure 7. Isopach map of the Ocala Limestone.

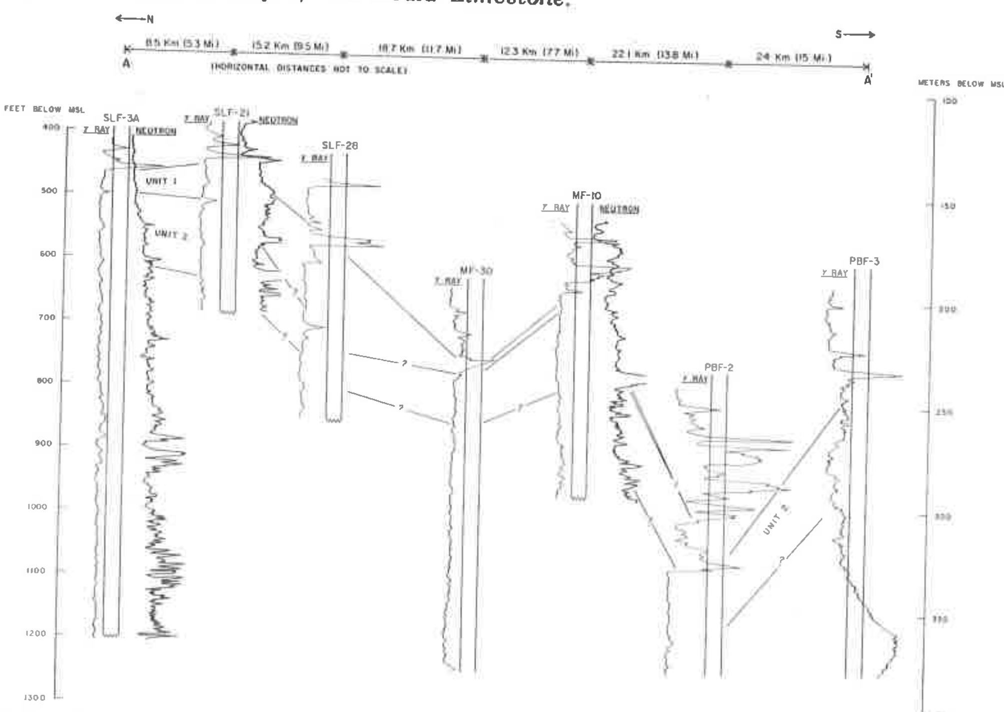

Figure 8. Stratigraphic cross-section A-A'.

in this paper marks the Oligocene unconformity of Vail (1978), then the resolution of the stratigraphic position of the unnamed calcilutite could have some interesting consequences. For example, if the unnamed calcilutite is in fact part of either the Hawthorn or Tampa Formations, then the last major unconformity in the study area

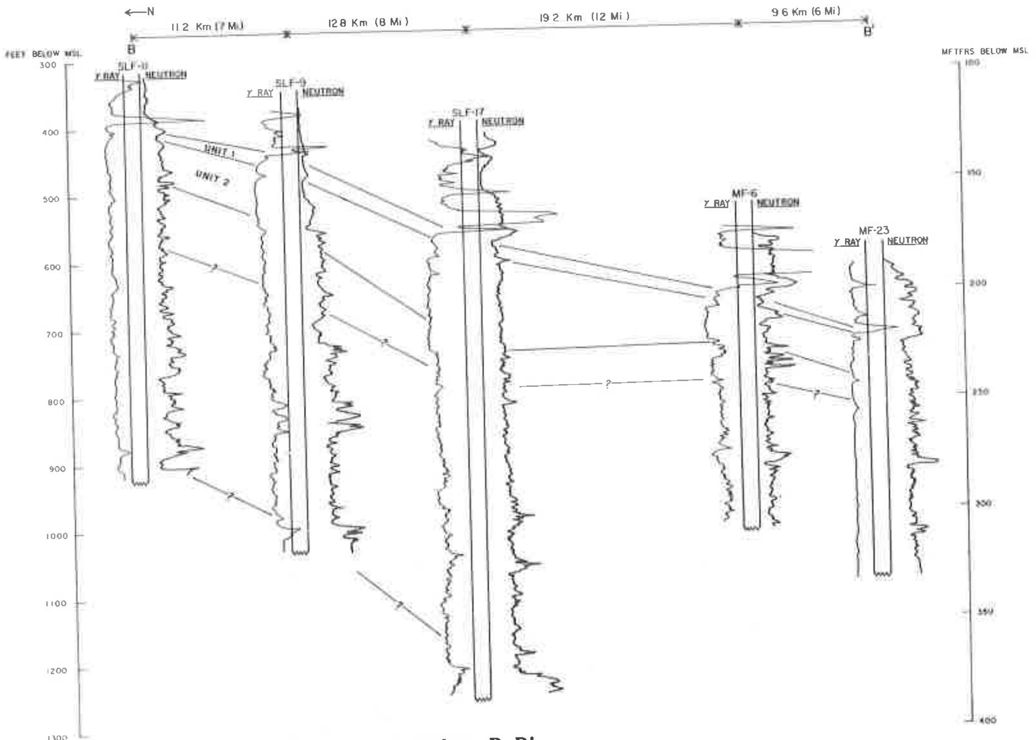

Figure 9. Stratigraphic cross-section B-B'.

formed during a hiatus in the deposition of the Miocene. Assuming that the Miocene age of the Hawthorn is correct, this would mean that the presumed Oligocene unconformity is absent in the study area.

On the other hand, if the Oligocene unconformity is marked by the phosphorite sand and the unnamed calcilutite is shown to be pre-Miocene in age, then other interesting conclusions are possible. The unnamed calcilutite can then be thought of as representing the youngest rock that survivied the erosion of the unconformity. This means that with the exception of two areas, less than 30 feet of post-Ocala rock survived the Oligocene erosion in the study area (Figure 6). Where it is recognized, the Oligocene section in South Florida is generally much thinner than in other parts of the state (Puri and Vernon, 1964).

The isopach map of the Ocala Limestone (Figure 7) shows two relatively thin areas (<50 feet of Ocala) in western and northeastern Martin County. The rest of the map shows a thicker section. The axis of this thicker section trends NW-SE through the study area with the thinner located on either side of the axis.

This suggests that a trough or low feature ran NW-SE through the area during deposition of the Ocala Limestone. This trough could have come about in many ways. More detailed study is necessary before the origin of this feature can be determined.

Three lines of cross-section are shown in Figure 1. Cross-sections A-A' and B-B' (Figures 8 and 9), running north-south, show the dip of the formations. Cross-section C-C' (Figure 10), running east-west, is approximately parallel to strike. On all the cross-sections, unit 1 refers to the interval on the geophysical logs that corresponds with the gray unnamed calcilutite in well SLF-3A. Unit 2 refers to the interval that corresponds with the Ocala Limestone. Below unit 2 is the Avon Park Limestone. While we believe that all of the geophysical logs as shown end in the Avon Park Limestone, correlations between beds within the Avon Park (primarily dolomite beds) have been proposed where possible.e. Due to the poor quality of the geophysical logs for well MF-30, PBF-2 and PBF-3, correlations involving these wells in cross-section A-A' (Figure 9) are speculative.

The surfaces of the formations are depicted as slightly undulating surfaces which

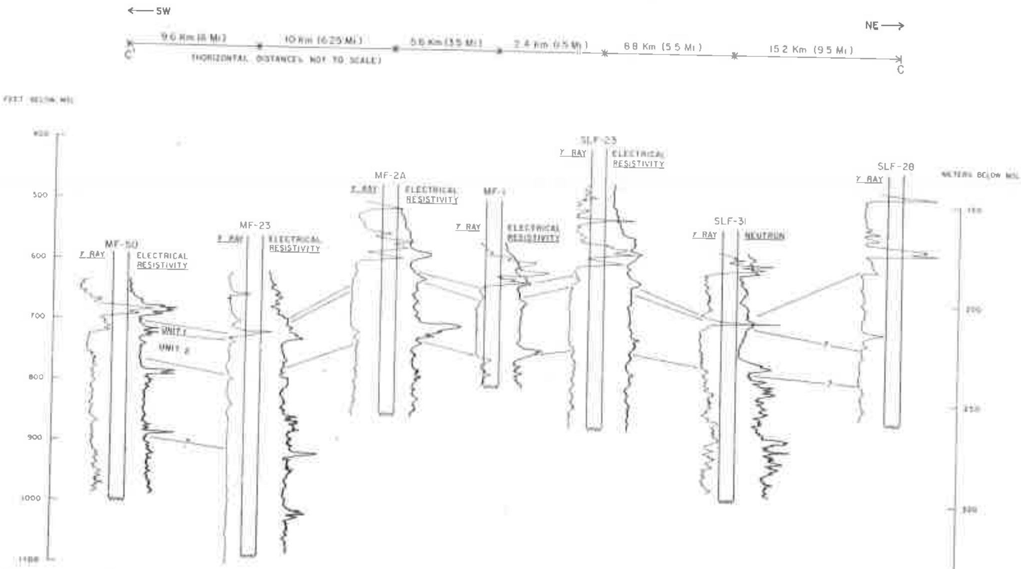

Figure 10. Stratigraphic cross-section C-C'.

may indicate slight warping in the area. No faulting is deemed necessary to produce the structures shown on the cross-sections.

As previously mentioned, slight warping is also seen in the structure contour maps (Figures 3, 4 and 5) and these warps may be the features interpreted by earlier workers as faults parallel to the coastline in this area. It should be noted that in section C-C' (Figure 10), the Ocala and Avon Park limestones trend deeper from well SLF-31 to well SLF-28. This line of section crosses two postulated faults, both of which predict that the formation depths should become deeper to the east (on the downthrown blocks). This deepening of the Ocala and Avon Park in well SLF-28 is accompanied by a dramatic thickening of the unnamed calcilutite. Here it is some 168 feet thick. The configuration and character of the unnamed càlcilutite in this area is the subject of a follow up study in progress by Mr. J.R. Armstrong.

CORRELATION OF GEOLOGIC AND HYDROLOGIC DATA

Comparison of the geologic data presented above with water quality and hydrologic characteristic data gathered in companion studies leads to some interesting correlations and conclusions. Figure 11 shows concentrations of total dissolved solids (TDS) of composite water samples collected in September, 1977, from the South Florida Water Management District's Florida aquifer system monitoring network. In regards to these data, Brown and Reece (1979) have shown that the relatively high TDS concentrations of Floridan aquifer system waters in Martin and St. Lucie Counties (due to the presence of connate water) vary areally but not temporally for measurements taken from 1976 to 1978. In Figure 11 a "ridge" of high TDS concentrations (2500 mg/l) trends northwest-southeast with concentrations decreasing to 1500 and 500 mg/l along the east and west flanks of this feature respectively. The pattern of TDS concentrations correlates with the isopach map pattern for the Ocala Limestone (Figure 7). A comparison of Figures 7 and 12 shows that the "ridge" of poorer quality water correlates with the area where the thickest section of Ocala Limestone exists. To the east and west the decrease in TDS concentrations correlates with the areas where the Ocala Limestone is thinnest, less than 50 feet thick. Thus the Ocala Limestone could be the primary source of the poor quality water in this area.

Brown (1979) documented aquifer recovery test data and analyses for the Floridan aquifer system in St. Lucie and Martin Counties. Figure 12 shows results of this study as ranges of composite transmissivity values. Of interest, the area of relatively high transmissi·ities correlates with the area where the Ocala Limestone is thickest and

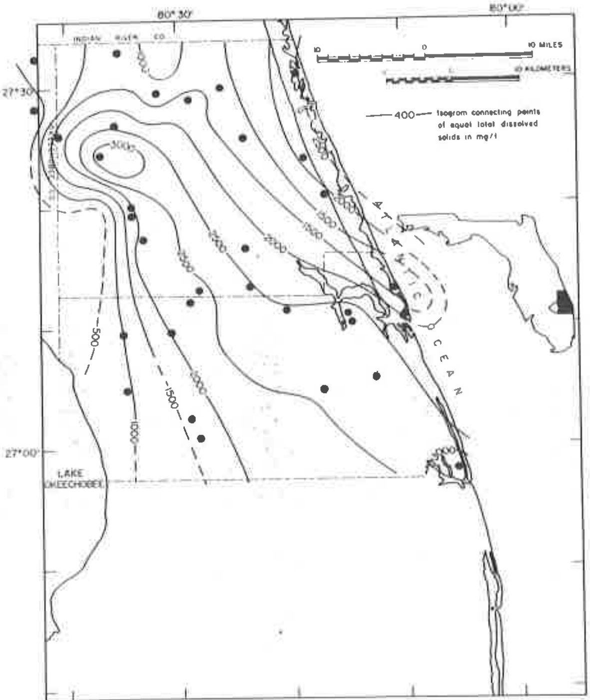

Figure 11. Total dissolved solids content of Floridan aquifer system waters for September, 1977 (replotted from Brown and Reece, 1979).

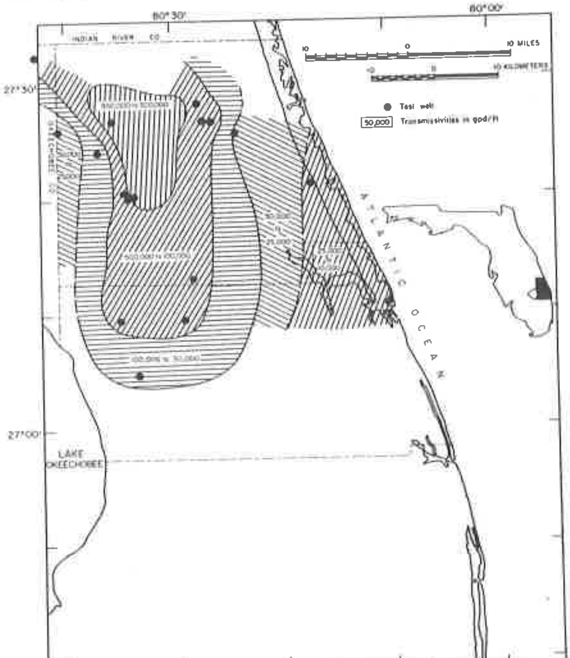

Figure 12. Composite transmissivity of the Floridan Aquifer System (replotted from Brown, 1979).

TDS concentrations are highest. The general north-south trend of high transmissivity values, although not precisely aligned with the axis of thick Ocala Limestone, is thought to be associated with this feature. The lack of data points in southwestern Martin County possibly biassed the contouring to cause the apparent north-south trend in high transmissivity values. The preliminary data, therefore, suggest that the Ocala Limestone could have the highest permeability within the aquifer system, although additional packer-type pumping tests are necessary to confirm this.

CONCLUSIONS

1. Subsurface stratigraphic study has delineated an unnamed post-Ocala limestone which may represent a new stratigraphic unit. If so, it should be formally named, described and mapped.
2. Correlations by gamma-ray and neutron borehole geophysical logs, when tied closely to a stratigraphic reference section, can be used for correlation over rather wide areas in southeast Florida.
3. No hard evidence was obtained to support the hypothesis of faulting in the study area although the evidence presented does not preclude such a possibility.
4. A comparison of geologic and hydrologic data show that the highest composite transmissivities but poorest quality water (highest composite total dissolved solids) occur along a trend where the Ocala Limestone is thickest.

ACKNOWLEDGMENTS

This study was condensed from a masters thesis submitted by the senior author to the Department of Geology, Florida State University in partial fulfillment of degree requirements (Mooney, 1979). We are especially grateful to the South Florida Water management District, particularly Mr. Abe Kreitman, Director of the Groundwater Division, for providing financial and logistical support for the project. The senior author is also pleased to acknowledge support of a one year HEW Mining, Minerals and Minerals Fuels Conservation Fellowship.

We thank Mr. Fred Meyer (U.S. Geological Survey, Water Resources Division, Miami) and Drs. William F. Tanner and Ramil C. Wright (FSU) for helpful discussion and advice. Mr. Meyer and the Florida Bureau of Geology also contributed much needed data or samples. We are also most grateful to Dr. James A. Miller (U.S. Geological Survey, water Resources Division, Atlanta) for his thorough and conscientious critical review of the manuscript.

Kathy Mooney, Rosemaarie Raymond, LaVerne Lamb, Dennis S. Cassidy, Tom Fellers, and John R. Armstrong (all FSU) provided valuable assistance in preparing the manuscript.

REFERENCES CITED

Agassiz, L., 1852, Florida Reefs, Keys and Coast: U.S. Coast Survey Rept. for 1851, app. 10, p. 145-160.

Anonymous, 1975a, PSAR St. Lucie unit 2 (rev. 26 - 1/8/75), Docket no. 50-335: unpublished engineering report, Law Engineering Testing Company, p. 2.5-1 thru 28. (unpublished)

__________, 1975b, Drilling and testing of deep disposal and monitoring wells, the city of Stuart, Martin County, Florida, Project no. 307-74-52: unpublished engineering report, Black, Crow and Eidsness, Inc., Gainesville, Florida, 51 p. (unpublished)

Applin, P.L. and Applin, E.R., 1944, Regional subsurface stratigraphy and structure of Florida and southern Georgia: Amer. Assoc. Petroleum Geologists Bull., v. 28, p. 1673-1753.

Applin, E.R. and Jordan, L., 1945, Diagnostic foraminifera from subsurface formations in Florida: Jour. Paleontol. v. 19, p. 129-148.

Brown, M.P., 1979, Aquifer recovery test data and analyses for the Floridan aquifer system in the Upper East Coast Planning Area: South Florida Water Management District: South Florida Water Management District Pub. 80-1, 52 p.

Brown, M.P. and Reece, D.R., 1979, Total dissolved solids of Floridan aquifer system waters for September, 1977, Upper East Coast Planning Area: South Florida Water Management District: South Florida Water Management District Hydrogeologic Map Atlas, Map Series 79-1, Plate 4.

Chen, C.S., 1965, The regional lithostratigraphic analysis of Paleocene and Eocene rocks of Florida: Florida Geol. Suv. Bull. 45, 105 p.

Cooke, C.W., 1915, The age of the Ocala Limestone: U.S. Geol. Surv. Prof. Paper 95, p. 107-120.

________, 1945, Geology of Florida: Florida Geol. Surv. Bull. 29, 342 p.

Cooke, C.W. and Mansfield, W.C., 1936, Suwannee Limestone of Florida: Geol. Soc. Amer. Proceed. for 1935, p. 71-72.

Dall, W.H., 1903, Tertiary fauna of Florida: Wagner Free Institute of Science Transactions, v. 3, pt. 6, p. 1541-1620.

Dall, W.H. and Harris, G.D., 1892, Correlation Paper: Neogene of North America: U.S. Geol. Surv. Bull. 84, 100 p.

Heilprin, A., 1887, Exploration on the West Coast of Florida and in the Okeechobee Wilderness: Wagner Free Institute of Science Transactions, v. 1, p. 1-134.

King, K.C. and Wright, R.C., 1979, Revision of the Tampa Formation West Central Florida: Gulf Coast Assoc. Geol. Soc. Transactions, vol. 29 (in press).

LeCount, J., 1857, On the agency of the Gulf Stream in the formation of the peninsula and keys of Florida: Amer. Jour. Sci., 2nd series, v. 23, p. 46-61.

________, 1878, Elements of Geology: D. Appleton and Co., New York, 588 p.

Lichtler, W.F., 1960, Geology and groundwater resources of Martin County, Florida: Florida Geological Survey Report of Investigations 23, 149 p.

Loeblich, A.R., Jr. and Tappan, H., 1964, (Sarcodina) Part C, Protista 2, vol. 1 and 2: in Moore, R.C., 1964, Treatise on Invertebrate Paleontology, Geological Society of America and Kansas University Press, 900 p.

Matson, G.C., and Clapp, F.G., 1909, Preliminary report on the geology of Florida: Florida Geol. Surv. Second Annual Report, 299 p.

Matson, G.C., and Sanford, S., 1913, Geology and ground-waters of Florida: U.S. Geol. Surv. Water-Supply Paper 319, p. 70-85.

Mooney, R.T., III, 1979, The stratigraphy of the Floridan aquifer System east and northeast of Lake Okeechobee, Florida: Masters Thesis, Department of Geology, Florida State University, Tallahassee, 61 p. (unpublished)

Parker, G.G., Ferguson, G.E. and Love, S.K., 1955, Water resources of southeastern Florida: U.S. Geol. Surv. Water-Supply Paper 1255, 965 p.

Puri, H.S., and Vernon, R.O., 1964, Summary of the geology of Florida and a guidebook to the classic exposures: Florida Geol. Surv. Spec. Pub. 5 (revised), 312 p.

Schroeder, M.C., Milliken, D.L. and Lowe, S.K., 1954, Water resources of Palm Beach County, Florida: Florida Geol. Surv. Report of Investigations 13, 63 p.

Shipley, T.H., Buffler, R.T. and Watkins, J.S., 1978, Seismic stratigraphy and geologic history of Blake Plateau and adjacent western Atlantic continental margin: Amer. Assoc. Petrol. Geol. Bull., v. 62, p. 792-812.

Smith, E.A., 1881, On the geology of Florida: Amer. Jour. Sci., 3rd series, v. 21, p. 292-309.

Stringfield, V.T., 1933, Groundwater in the Lake Okeechobee area, Florida: Florida Geol. Surv. Report of Investigations 2, 31 p.

________, 1936, Artesian water in the Florida peninsula: U.S. Geol. Surv. Water-Supply Paper 773-C, p. 115-195.

Vail, P.R., Mitchum, R.M., Jr., and Thompson, S., III, 1978, Relative changes of sea level from coastal onlap, in Payton, C., ed., Stratigraphic interpretation of seismic data: Am. Assoc. of Petrol. Geol. Mem. 26, p. 63-81.

Vernon, R.O., 1970, The beneficial use of zones of high transmissivities in the Florida subsurface for water storage and waste disposal: Florida Geol. Surv. Information Circular 70, 39 p.

LATE DEVONIAN CHRONOSTRATIGRAPHIC CORRELATIONS BETWEEN THE CENTRAL APPALACHIAN ALLEGHENY FRONT AND CENTRAL AND WESTERN NEW YORK

By

George R. McGhee, Jr.
Department of Geological Sciences
Wright Geological Laboratory
Rutgers University
New Brunswick, New Jersey 08903

and

John M. Dennison
Department of Geology
University of North Carolina
Chapel Hill, North Carolina 27514

ABSTRACT

Considerably refined time correlations are now available for Late Devonian clastic strata of the central Appalachian region using a model which combines classical paleontological methods with more recently developed eustatic sea-level variation curves for the Appalachian basin. Preliminary biostratigraphic control for the boundaries of the Greenland Gap Group of Maryland, West Virginia, and Virginia is obtained by utilizing the time ranges of the brachiopods *Cyrtospirifer sulcifer* (Hall), *Athyris angelica* Hall, *Nervostrophia nervosa* (Hall); the bivalve *"Cornellites" chemungensis* (Conrad); and parts of the collective ranges of the brachiopod families Cyrtospiriferidae and Atrypidae. Within this framework, specific correlations from the central Appalachians to New York are developed by matching times of sea-level fluctuations as preserved in the stratigraphic column of the eastern United States. Eustatic sea-level rises preserved winnowed sands on the eastern side of the Appalachian basin, synchronous with presently methane-enriched, black-shale marker zones which accumulated as a result of slow deposition west of the basin axis. Brownish-gray and non-marine redbed pulses into the basin are produced by sea-level drops.

The Red Lick Member of the Foreknobs Formation (Greenland Gap Group) is equivalent to the Caneadea Formation (in New York). The Pound Sandstone, Blizzard, and Briery Gap Sandstone Members of the Foreknobs Formation are believed to be time equivalents to the Hume Shale, Wiscoy Formation, and Pipe Creek Shale, respectively, in New York. The Mallow Member is equivalent to the Nunda and the upper part of the Gardeau Formations of New York. The Scherr Formation (of the Greenland Gap Group) is equivalent to the remaining lower portion of the West Falls Group of New York.

INTRODUCTION

The age of the Greenland Gap Group strata (previously called the "Chemung Formation") along the Allegheny Front in Maryland, West Virginia, and Virginia (Figure 1) has been of interest to Devonian workers for some time. These strata (Figure 2) have been extensively studied from stratigraphic (Dennison, 1970; McGhee and Dennison, 1976), petrological (Kirchgessner, 1973), and paleontological (McGhee, 1976) points of view. Recent attempts to resolve the problem of time relations within the Foreknobs and Scherr Formations (Greenland Gap Group) include the works of Dennison (1971), Curry (1975), and McGhee (1977).

The object of this paper is to propose a new chronostratigraphic correlation between the Late Devonian Greenland Gap Group of the central Appalachians and the New York State Devonian reference section. We feel that this correlation is the most accurate time correlation possible given current stratigraphic and paleontological data.

PALEONTOLOGICAL CONTROL

The following represents a brief review and synopsis of the paleontological contributions toward the resolution of time relations within the Greenland Gap Group, using classical biostratigraphic techniques. Concerning the geologic age of the Greenland Gap Group, Dennison (1970, p. 64) stated: "The basal beds of the Scherr Formation contain *Cyrtospirifer chemungensis* (Conrad), and *Cornellites chemungensis* (Conrad), which indicate age assignment to the lower Cohocton Stage." Elsewhere, concerning the upper limit of the Greenland Gap Group, he stated: "Along the Allegheny Front in the area studied no fossils in the uppermost member of the Foreknobs Formation appear younger than Cohocton Stage" (Dennison, 1970, p. 71). However, in view of the thickness of the Red Lick Member of the Foreknobs Formation at the Route 250 section, Dennison (1971, p. 1186) later considered that the Foreknobs Formation might extend upward into the post-Cohocton toward the southern end of the study area.

Palynologic zonation of the Greenland Gap Group was attempted by Curry (1975), though without marked success due to the largely unfossiliferous nature of the outcrops with respect to miospores. Curry (1975, p. 129) tentatively assigned a Middle to Late Frasnian age to the Greenland Gap Group, after noting that almost all of his described samples came from the Mallow Member of the Foreknobs Formation at the Corriganville section.

McAlester (1962) criticized the use of *"Cornellites" chemungensis* (Conrad) as a guide for the Chemung (Cohocton) Stage, noting that its present usage is based solely on its empirically determined distribution in the New York Devonian to date, and not upon its evolutionary position, which remains uncertain. However, co-occurrence with the brachiopod *Nervostrophia nervosa* (Hall) in the Mallow Member of the Foreknobs Formation (McGhee, 1976) supports the view that the Scherr-Foreknobs contact lies within the middle Cohocton Stage, based on the range of that species in New York and as documented by Rickard (1964).

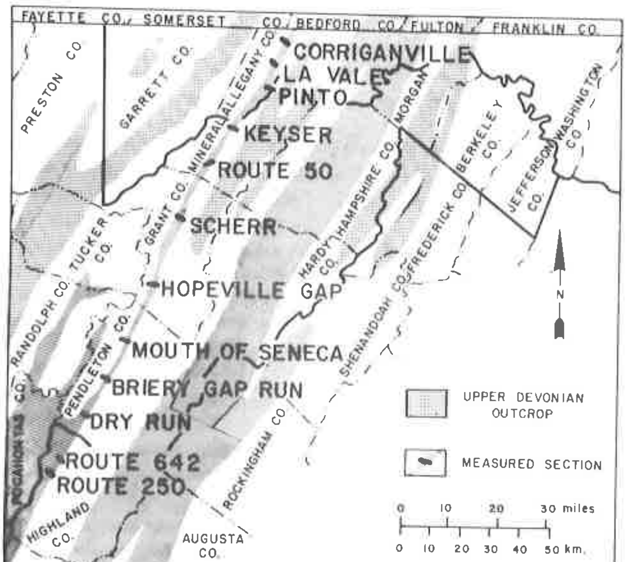

Figure 1. Location of sections studied along Allegheny Front (after Dennison, 1971).

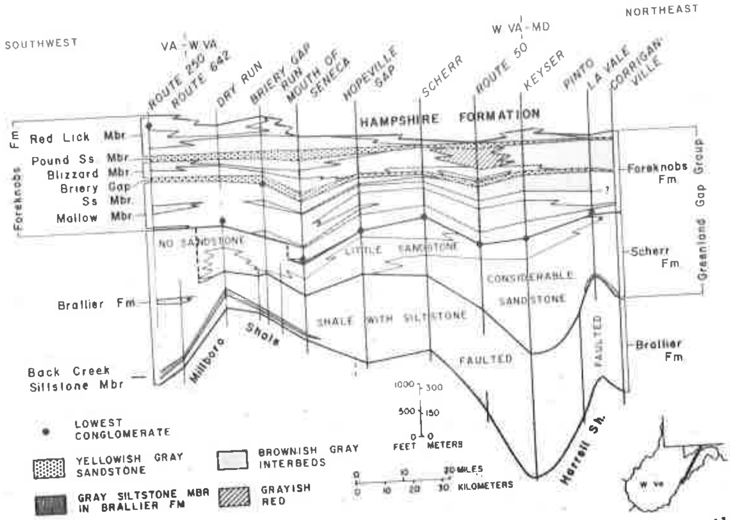

Figure 2. Stratigraphic cross-section of Greenland Gap Group along the Allegheny Front. Datum used is the top of the Pound Sandstone (after Avary and Dennison, modified from McGhee and Dennison, 1976).

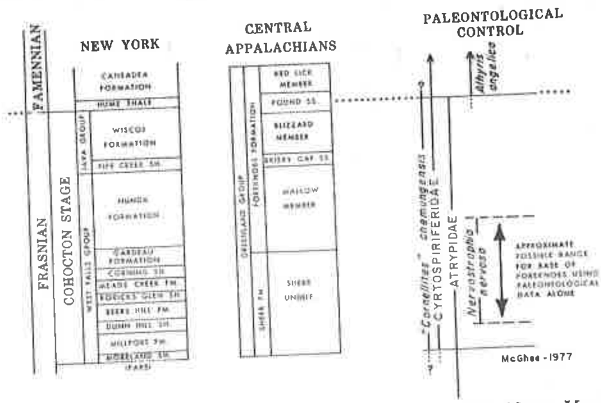

Figure 3. Late Devonian paleontologic correlations between New York and the central Appalachians.

The Frasnian-Famennian Series boundary was determined by McGhee (1977) to lie within the Pound Sandstone Member in Maryland and West Virginia. This determination was based on the first occurrence of two articulate brachiopods, *Athyris angelica* Hall and *Cyrtospirifer sulcifer* (Hall), and the last observed occurrence of members of the Atrypidae. Previous Devonian workers are in agreement that the first appearance of *Athyris angelica* Hall marks the beginning of the Canadaway (Cassadaga) Stage (Chadwick, 1935, p. 323; Willard, 1939, p. 251; Cooper and others, 1942; McAlester, 1962, p. 6). Likewise, *Cyrospirifer sulcifer* is presently considered to extend no lower than the beginning of the Cassadaga Stage (McGhee, 1977). Highest observed occurrences of members of the Atrypidae along the Allegheny Front, usually the species *Spinatrypa hystrix* (Hall), were in the lower Pound Sandstone. This entire family of articulate brachiopods are believed to have become extinct by the close of the Frasnian Epoch (Boucot and others, 1965; Copper, 1973, text-fig. 1).

In summary, from paleontological data alone the age of the base of the Greenland Gap Group, the position of the Frasnian-Famennian boundary within the Group, and a time-range for the base of the Foreknobs Formation can be obtained. These relationships are outlined in Figure 3.

THE EUSTATIC MODEL

Eustatic sea-level changes leave sedimentary imprints which should be traceable for long distances in shallow marine sediments. Thus, when basin-wide sea-level fluctuations occur, they should provide ideal correlation indicators which can be preserved in the stratigraphic column. Difficulty arises, however, in trying to distinguish those lithologic markers which have originated from eustatic fluctuations and those records of submergence and emergence which have arisen from localized tectonic activity on the basin margin, local shoreline shifts related to local delta construction or destruction, and sediment supply factors.

Dennison (1971) proposed a combination geographic and stratigraphic facies model designed to detect the sedimentary imprint of a basin-wide eustatic sea-level change. This model has been considerably refined and expanded by Dennison and Head (1975, Figure 1), and used to construct a sea-level variation curve for the Appalachian basin Silurian and Devonian. This curve compares favorably with a similar curve constructed by Johnson (1971) for the middle Paleozoic of western North America.

In the main Appalachian basin from Virginia north to New York the record is sufficiently preserved that information is available in the stratigraphic column concerning the history of tectonic uplift, subsidence areas, marine and non-marine clastic wedges, and lithofacies and paleoecologic patterns. A key concept in the recognition of eustatic sea-level fluctuations is the discernment of certain sets of contemporaneous events which can be explained by water depth changes and which can be recognized on two or more sides of the basin margin or, in some cases, recognized in at least two separate delta lobes of the Catskill delta complex. These events can be interpreted as regional, basin-wide, sea-level changes as distinguished from local shoreline shifts. A regional change is defined as any water depth change that appears to have affected the entire basin.

In a depositional basin with gentle offshore slopes and bounded by a series of delta lobes, a sea-level drop would immediately lower the base level. This would cause the simultaneous rapid progradation of all the delta lobes, and could result in a series of contemporaneous redbed tongues of non-marine sediments extending into the basin from each lobe. Reddish strata may be swept offshore into marine waters, extending a faint reddish (brownish gray) coloration for some distance offshore, out to the limit where marine reduction of reddish coloration occurs in oxygen-deficient sediments. An abrupt sea-level rise would cause the simultaneous shift of the shoreline back toward the basin margin, flooding the previous deltas and stream channels, and resulting in the deposition of fluvial sediments only in nearshore drowned valleys and floodplains. With the onset of floodplain alluviation, quiet water silts and shales would accumulate over the former marine and non-marine delta constructs.

Likewise, in a basin with steeper offshore slopes and only moderate sediment supplies, wave action energy would be concentrated into narrower longshore zones and

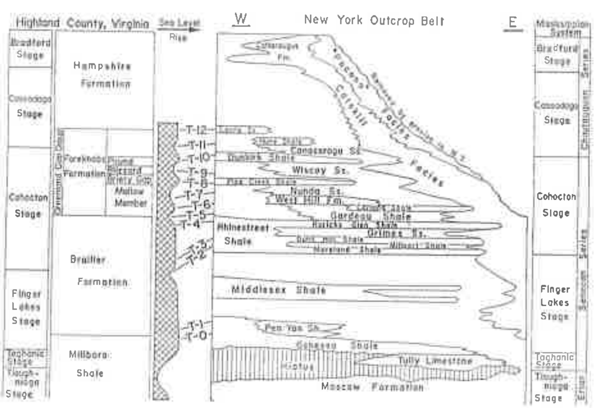

Figure 4. Tentative eustatic correlation of New York Devonian Strata with the Route 250 section in Virginia (after Dennison, 1971).

would not be overwhelmed by clastic influx. This would result in the formation of a series of winnowed submerged or possibly emergent sand bars or shallow delta-front sand deposits. Sea-level drops or rises would not cause the shoreline to migrate as extensively as in the case of a basin with gentle offshore slopes, and could result in the deposition of shallow water sheet sands extending either into or out of the basin, produced by the steady shifting of winnowed sand deposits created by the migration of the wave action zone. Non-marine redbed pulses into the basin would also be expected from sea-level drops, with some faint reddish (brownish gray) coloration extending into near-shore marine waters, offshore to where complete reduction occurs in oxygen-deficient sediments.

As previously pointed out by Dennison and Head (1975), basin-wide sea-level variations may reflect changes in the world volume of ocean water, vertical or lateral movement of the crust, shifts in rate of seafloor spreading, or even general tilting of portions of the crust on such a large scale that the relatively small Appalachian basin experienced a general rise or drop in sea-level.

COMBINATION MODEL FOR CHRONOSTRATIGRAPHIC CORRELATION

Classically, time correlations and the recognition of time surfaces across extensive geographic distances has been accomplished through the utilization of paleontological comparisons, often with only widely spaced stratigraphic sections as controls. Though the introduction of radiometric dating has considerably refined time correlations and allowed the stratigraphic column to be scaled in terms of "measured time" (rather than the "relative time" of evolutionary paleontology), the fact still remains that the most useful methods of regional and intercontinental time correlation are the techniques of biostratigraphy. Yet there are limits to the amount of subdivision that can be imposed upon the evolutionary spectrum discernable from the fossil record. The problems of ecophenotypic species variations and species habitat preferences ("facies fossils") are well known, and thus it is often difficult to construct an evolutionary phylogeny for a given series of taxa in anything less than the broadest generalizations.

Increasingly, stratigraphers are realizing that physical correlation of closely spaced sections may provide significantly refined resolutions of time relations. One type of such physical correlation may be the matching of bentonite ash beds, another may be that of recognizing eustatic sea-level pulses. However, an essential factor in the accuracy of such correlations is the spacing of the sections compared. Correlation of two widely spaced sections on physical criteria alone can be very risky, for while the respective physical records may be preserved in each section, it is often difficult to demonstrate which portion of the respective records were caused by the same event in time. The problem is compounded when one deals with numerous such events occurring in each section.

Dennison (1971, Figure 7) outlined a preliminary study of "time lines" within the Devonian delta strata along the Allegheny Front in a series of closely spaced sections, each of which has been examined in bed-by-bed detail. A tentative correlation between the Allegheny Front to New York State was then attempted, based primarily on the eustatic sea-level model. That correlation is reproduced in Figure 4, which represents the finest time resolution he could then obtain from a sequencing of physical events.

Though further work has supported the general framework given in Figure 4, recent studies have provided new data which allow substantial revision of the correlations. These studies include the revision of New York Devonian stratigraphy (Rickard, 1975), the paleontological work of McGhee (1976, 1977), the work of Dennison and Head (1975) with Appalachian basin eustatic sea-level variation curves, and a re-examination of the original stratigraphic sections (Figure 2).

Utilizing the paleontological control given in Figure 3, the Pound Sandstone Member of the Foreknobs Formation can be correlated with the Hume Shale of New York, and the base of the Scherr Formation (Greenland Gap Group) can be shown to fall within the New York Moreland Shale. Within this biostratigraphically obtained time-framework, specific physical correlations from the central Appalachians to New York can be developed by matching times of sea-level fluctuations. The Appalachian basin was roughly elliptical in outline, or rowboat-shaped, during most of the middle and late Paleozoic, with the present day New York area being located towards the northeastern end, or stern, of the basin. Sea-level rises produced widespread black shale tongues in this part of the basin, black shale tongues which penetrated eastward into the coarser facies of the Catskill delta complex as documented by Sutton (1963) and Sutton, Bowen, and McAlester (1970). Offshore slopes were steeper along the southeastern side of the basin in the area of the present day central Appalachians, and minor sea-level rises did not produce widespread black shale tongues in the southeastern part of the basin. More pronounced sea-level rises did, however, produce wave-winnowed sheet sands from the migration of offshore bar deposits, while drops in sea-level resulted in mostly marine, brownish gray, "redbed" tongues extending into the deeper water marine facies. Our proposed revision of chronostratigraphic correlations between these two parts of the basin for the Foreknobs Formation is given in Figure 5 (cf. Figure 4). Beside each stratigraphic column is given the pertinent facies and interpreted event producing that facies, and the proposed series of correlations between those events are indicated. The paleontologically obtained correlation between the Pound Sandstone and Hume Shale is taken as the base line against which the other correlations are then sequenced. A eustatic sea-level variation curve is given for this segment of the eastern United States Late Devonian. Sea-level rises are indicated to have been more abrupt than sea-level falls, analogous with the record of the Würm glacio-eustatic event. As previously noted, however, the actual causal mechanisms for Late Devonian eustatic events are by no means clear (Dennison and Head, 1975), glacio-eustatic events being only one of several possible explanations.

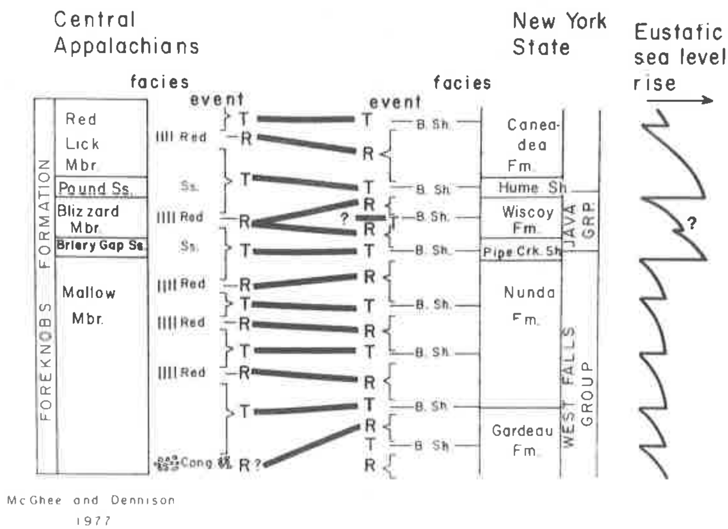

Figure 5. Proposed Devonian chronostratigraphic correlations between the Allegheny Front and Upstate New York. Event explanation: T = marine transgression, R = marine regression. Facies explanation: Red = brownish-gray siltstone tongues, B. Sh. = black shale tongues, Ss. = sheet sands.

The relative magnitudes of sea-level fluctuations are difficult to estimate from the rock record. The Pound Sandstone-Hume Shale and Briery Gap Sandstone-Pipe Creek Shale events appear to have been more pronounced, as is indicated by the formation of sheet sands in the central Appalachians and extensive black shales in New York. In addition, the base of the Foreknobs Formation, which is marked by a clastic pulse into the basin recorded over two separate delta lobes, probably records a basin-wide shallowing event. One anomaly in the fit of the eustatic event sequencing is the presence of a restricted shale tongue indicated by Rickard (1975) to occur within the Wiscoy Formation. If in fact this records a minor sea-level rise in New York, equivalent indications have not yet been recognized in the central Appalachian strata.

A comparison of our revision with previously proposed chronostratigraphic

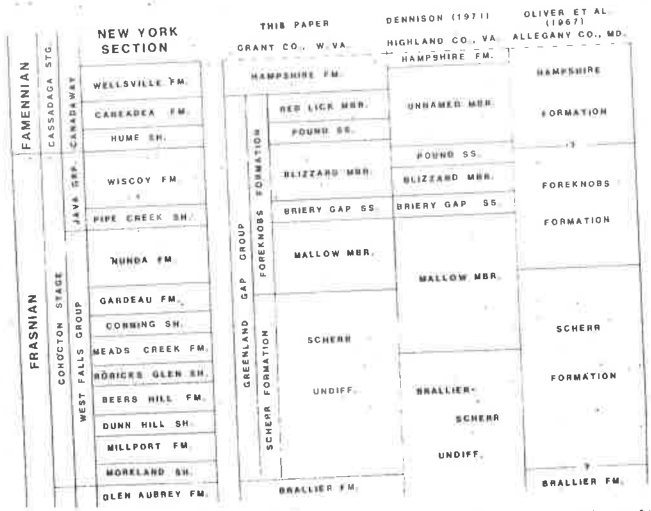

Figure 6. Comparison of proposed correlations (this paper) with the previous chronostratigraphic correlations of Dennison (1971) and Oliver and others (1967).

correlations is provided in Figure 6. It is proposed that the Red Lick Member of the Foreknobs Formation (Greenland Gap Group) is equivalent to the Caneadea Formation (in New York). The Pound Sandstone, Blizzard, and Briery Gap Sandstone Members of the Foreknobs Formation are proposed to be time equivalents to the Hume Shale, Wiscoy Formation, and Pipe Creek Shale, respectively, in New York. The Mallow Member is equivalent to the Nunda and the upper part of the Gardeau Formations of New York. The Scherr Formation is equivalent to the remaining lower portion of the West Falls Group.

ACKNOWLEDGMENTS

We thank A.J. Boucot and W.F. Koch for their constructive review of the manuscript. Much of our work in the central Appalachian region has been supported by the West Virginia Geological Survey, for which support we are grateful. Acknowledgment is made to the donors of the Petroleum Research Fund, administered by the American Chemical Society, for partial support of this research (ACS-PRF 11739-G2 to G.R. McG.), and to Rutgers University. An earlier version of this paper was presented at the Geological Society of America Southeastern Sectional Meeting in Winston-Salem, N.C. (McGhee and Dennison, 1977). We are indebted to Sondra Dennison for her time and patience in drafting Figures 3 and 5.

REFERENCES CITED

Avary, K.L. and Dennison, J.M., 1980, Back Creek Siltstone Member of Devonian Brallier Formation in Virginia and West Virginia: Southeastern Geology, in press.

Boucot, A.J., Johnson, J.G., Pitrat, C.W., and Staton, R.D., 1965, Spiriferida: In R. C. Moore (ed.), Treatise on Invertebrate Paleontology, Part H, Brachiopoda, p. H632-H728, Geol. Soc. America and Univ. Kansas Press, Lawrence, Kansas.

Chadwick, G.H., 1935, Faunal differentiation in the Upper Devonian: Geol. Soc. Amer. Bull., v. 46, p. 305-342.

Cooper, G.A. (Chairman), and others, 1942, Correlation of the Devonian sedimentary formations of North America: Geol. Soc. Amer. Bull., v. 53, p. 1729-1794.

Copper, P., 1973, New Siluro-Devonian atrypoid brachiopods: Jour. Paleo., v. 47, p. 484-500.

Curry, R.P., 1975, Miospores from the Upper Devonian (Frasnian) Greenland Gap Group, Allegheny Front, Maryland, West Virginia and Virginia, U.S.A.: Review of Palaeobotany and Palynology, v. 20, p. 119-131.

Dennison, J.M., 1970, Stratigraphic divisions of the Upper Devonian Greenland Gap Group ("Chemung Formation") along Allegheny Front in West Virginia, Maryland, and Highland County, Virginia: Southeastern Geology, v. 12, p. 53-82.

_______, 1971, Petroleum related to Middle and Upper Devonian deltaic facies in central Appalachians: Amer. Assoc. Petrol. Geol. Bull., v. 55, p. 1179-1193.

Dennison, J.M., and Head, J.W., 1975, Sealevel variations interpreted from the Appalachian basin Silurian and Devonian: Amer. Jour. Sci., v. 275, p. 1089-1120.

Johnson, J.G., 1971, Timing and coordination of orogenic, epeirogenic, and eustatic events: Geol. Soc. Amer. Bull., v. 82, p. 3263-3298.

Kirchgessner, D.A., 1973, Sedimentology and petrology of Upper Devonian Greenland Gap Group along the Allegheny Front, Virginia, West Virginia and Maryland: Unpublished Ph.D. dissertation: 93 p., Univ. of North Carolina, Chapel Hill, North Carolina.

McAlester, A.L., 1962, Upper Devonian pelecypods of the New York Chemung Stage: Peabody Mus. Nat. Hist. Bull., v. 16, 88 p., 32 pls.

McGhee, G.R., Jr., 1976, Late Devonian benthic marine communities of the central Appalachian Allegheny Front: Lethaia, v. 9, p. 111-136.

, 1977, The Frasnian-Famennian (Late Devonian) boundary within the Foreknobs Formation, Maryland and West Virginia: Geol. Soc. Amer. Bull., v.88, p. 806-808.

McGhee, G.R., Jr., and Dennison, J.M., 1976, The Red Lick Member, a new subdivision of the Foreknobs Formation (Upper Devonian) in Virginia, West Virginia, and Maryland: Southeastern Geology, v. 18, p. 49-57.

, 1977, Chrono-stratigraphic correlation of the Late Devonian Greenland Gap Group, central Appalachians, with the New York Devonian Reference Section: Geol. Soc. Amer. Abstr., v. 9, p. 166-167.

Oliver, W.A., Jr. and others, 1967, Devonian of the Appalachian basin United States. In International Symposium on the Devonian System: Alberta Soc. Petroleum Geologists, v. 1, p. 1001-1040.

Rickard, L.V., 1964, Correlation of the Devonian rocks in New York State: New York State Mus. and Sci. Serv., Geol. Survey, Map and Chart Series No. 4.

, 1975, Correlation of the Silurian and Devonian rocks in New York State: New York State Mus. and Sci. Serv., Map and Chart Series No. 24.

Sutton, R.G., 1963, Correlation of Upper Devonian strata in south-central New York: Pennsylvania Topog. and Geol. Survey Gen. Geol. Rept. G39, p. 87-101.

Sutton, R.G., Bowen, Z.P., and McAlester, A.L., 1970, Marine shelf environments of the Upper Devonian Sonyea Group of New York: Geol. Soc. Amer. Bull., v. 81, p. 2975-2992.

Willard, B., 1939, The Devonian of Pennsylvania: Pa. Geol. Survey, 4th Ser., Bull. G19, 481 p.

EXPRESSION OF A CROSS-STRIKE STRUCTURAL DISCONTINUITY IN PENNSYLVANIAN ROCKS OF THE EASTERN PLATEAU PROVINCE

By

John A. LaCaze, Jr.[1]

and

Russell L. Wheeler
Department of Geology and Geography
West Virginia University
Morgantown, West Virginia 26506

ABSTRACT

Major, cross-strike structural discontinuities appear to be large volumes of intensely fractured rock in overthrust belts, and may be useful guides in exploration for fractured gas reservoirs. To determine whether the Petersburg lineament of the western Valley and Ridge province of West Virginia is expressed in the Pennsylvanian rocks of the easternmost Plateau province, LaCaze mapped about 60 square km (23 square mi) of late Paleozoic molasse atop Allegheny Front, in an area straddling the lineament's westward projection. Eight outcrops of Homewood sandstone strung across the projection contain longitudinal, transverse, and an east-striking set of diagonal joints. Joint intensity (surface area per unit volume of rock) in the projection is about twice that outside. Structure contours show that a map-scale syncline is disrupted immediately west of the high-intensity zone. Thus the lineament is expressed in the Pennsylvanian rocks of the easternmost Plateau province as a zone of increased fracturing at least 1.3 km (0.81 mi) wide, with associated broader disruption of map-scale structure.

INTRODUCTION

Purpose

Cross-strike structural discontinuities (CSD's) are "map-scale structural lineaments or alignments, at high angles to regional strikes, and best recognized as disruptions in strike-parallel structural or geomorphic patterns" (Wheeler, Winslow, and others, 1979). CSD's are abundant in the central and southern Appalachians and in other overthrust belts. Wheeler, Winslow, and others summarize characteristics of many mapped CSD's, and much of the following discussion of CSD's is based on work cited by them.

Present evidence is that CSD's are several km (mi) wide and each may contain on the order of 1000 cubic km (240 cubic mi) of intensely fractured rock (Wheeler, 1978a, 1978b, and 1978c). The Devonian clastic sequence of the central and southern Appalachians contains gas in fractured reservoirs, and many workers are trying to improve exploration and stimulation methods applicable to such reservoirs (Shumaker and Overbey, 1976; Wheeler and others, 1976; Schott and others, 1977; Anonymous, 1978). CSD's and their associated fracture systems may provide a method for extending that exploration effort into the little tested eastern Plateau province of West Virginia and adjacent states, where the Devonian clastic sequence is detached.

One such CSD is the Petersburg lineament in the Valley and Ridge province of West Virginia (Figure 1; Sites, 1978 and 1979). This paper has two purposes: (1) To determine whether the Petersburg lineament is expressed in Pennsylvanian rocks of the easternmost Plateau province, atop the Allegheny Front, and (2) to test methods for mapping the Petersburg lineament and other CSD's westward, under the Pennsylvanian and Permian cover of the Plateau province.

[1]Present address: Rt. 5, Box 1000, Space 203, Midland, Texas 79701

Setting

In Grant and Hardy Counties, West Virginia, the Petersburg lineament trends about N75°E and is well exposed in the Silurian and Devonian rocks of the western Valley and Ridge province (Sites, 1978 and 1979; other work cited by Wheeler, Winslow, and others, 1979). Immediately west of the Wills Mountain anticline at the western edge of the Valley and Ridge province, the lineament is subtly present in Devonian and Mississippian rocks on Allegheny Front (Wilson and Wheeler, 1974). The westward projection of the lineament into the easternmost Plateau province passes through the Dolly Sods Wilderness Area of Tucker County, West Virginia (Figure 1). There, red, flaggy sandstones, siltstones, and shales of Late Mississippian age are overlain by coals, shales, siltstones, conglomerates, and thick, flaggy or cross-bedded sandstones of Pennsylvanian age. The exposed section is 1000 to 1500 feet thick (305 to 457 m), and sedimentary layers (as distinct from cross beds) have low dips on both limbs of the map-scale Stony River syncline that includes the map area (LaCaze, 1978; Reger, 1921). The map area covers about 23 square mi (60 square km) astride the lineament's projection.

JOINTS

Systematic Joints

Sets. The topographic break of Allegheny Front in easternmost Tucker County exposes thick, cross-bedded, well-cemented, quartz sandstones and conglomerates of the uppermost Lower Pennsylvanian Homewood sandstone (LaCaze, 1978). There, the Homewood contains three sets of systematic joints: transverse, striking about N48°W; longitudinal, striking about N28°E; and diagonal, striking about N75°E. Eight large Homewood outcrops of relatively constant lithology and bed characteristics are scattered for 6.2 mi (9.5 km) across the westward projection of the Petersburg lineament through Dolly Sods, and along the eastern border of the map area. These outcrops and the outline of the map area are shown in Figure 3. In each outcrop, LaCaze measured spacing for those three sets: perpendicular separations of adjacent joints in the same set. The diagonal set was not found in the two outcrops southwest of the lineament's projection. Wheeler later found but did not measure that set in a smaller Homewood outcrop yet further southwest.

Relative Ages. Wheeler attempted to estimate relative ages of the three sets, using criteria successfully applied to finer-grained rocks by Kulander and others (1977 and 1978), Wheeler and Holland (1978), Dixon (1979b), and Wheeler, Holland, and others (1979). Plumose and other delicate structures on joint faces (transient features of Kulander and others, 1977) can record the origin, propagation history, and end of a joint, but the Homewood exposures are too coarse-grained and joint surfaces too weathered to preserve such features recognizably. A young joint will usually abut against or hook into an older one in a T- or J-intersection, and such tendential features (Kulander and others, 1977) are abundantly preserved in the Homewood outcrops. However, evaluation of 105 tendential relations, mostly abuttings, in and near the outcrops that yielded spacing measurements gave no conclusive results. For each of the three possible pairings of joint sets, the tendential relationships are about equally divided on relative ages of the two sets involved. The most conclusive determination of relative ages was that transverse joints predate longitudinal joints, which was also found by Wheeler and Holland (1978) and Dixon (1979a, 1979b) elsewhere. However, that relative age has a significance value of only 0.227 by the binomial test. That is, there is one chance in four or five that the observed relative ages between longitudinal and transverse joints, or a relationship yet more conclusive, could have arisen by chance alone.

There are several possible reasons for the lack of clear-cut relative ages. (1) All three sets may have formed coevally. (2) Each set may have been filled and cemented shut, or its walls may have been pressed together with coarse grains interlocking, in stress-transmitting contact, before the next set formed. Neither of us saw any evidence

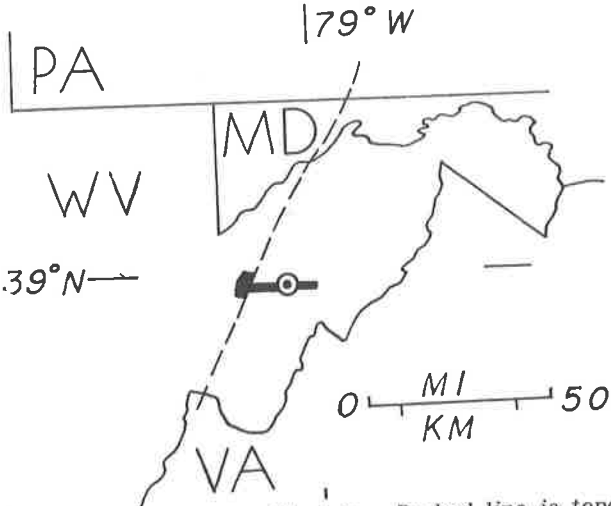

Figure 1. Index map of eastern West Virginia. Dashed line is topographic break of Allegheny Front, with Valley and Ridge province to east and Plateau province to west. Circled dot is Petersburg, and heavy solid line through it is mapped extent of Petersburg lineament (lineament may be several times the length shown: Sites, 1978 and 1979). Black polygon atop Allegheny Front at west end of lineament is Dolly Sods map area (size and shape approximate).

of remaining joint fillings, but joint walls are intensely weathered, stained with iron oxyhydroxides and eroded. (3) In some X-intersections where two joints appear to cross each other, one arm of the X is short and narrow. Also, some joints hook abruptly into other nearby and parallel joints, and in some such occurrences, the hook is much less weathered than the rest of the joints. Those observations are consistent with the possibility that physical and chemical weathering, acting long after jointing ceased, might cause a younger joint to propagate across an older one in some manner, changing a T-intersection into an X-intersection. At present, improving understanding of jointing and modification of joints by weathering seems more important than collecting more data.

Joint Intensity

Calculation. Wheeler (1979), Wheeler and Dixon (1979), and Dixon (1979b) define joint intensity (I) as the amount of joint surface area per unit volume of rock. Following Vialon and others (1976), they calculate the intensity of a single set as the inverse of the set's mean spacing. Then intensity of an exposure is the sum of the intensities of all sets in that exposure. Stearns (1968) and Rousell and Everitt (1978) independently developed an equivalent measure of intensity based on joint frequency. Wheeler and Dixon (1979) argue that intensity is less distorted by a few extremely large or small spacing values if one replaces mean spacing by the trimean S_T (Tukey, 1977). The resulting estimator of I is called I_T. S_T is calculated as

$$S_T = \frac{1}{4} \left(Q_1 + 2\, S_M + Q_3\right)$$

where Q_1 and Q_3 are the first and third quartiles and S_M the median of spacing values.

Evaluation. Values of I_T for the Homewood sandstone are in Figure 2. Intensities are higher in the westward projection of the lineament (outcrops 3 through 6). That occurs for all three sets, and all outcrops are on the topographic break of Allegheny Front, so topography and mass wasting are not responsible for the increased intensity.

Photolineaments are often found or assumed to overlie unusually fractured rock.

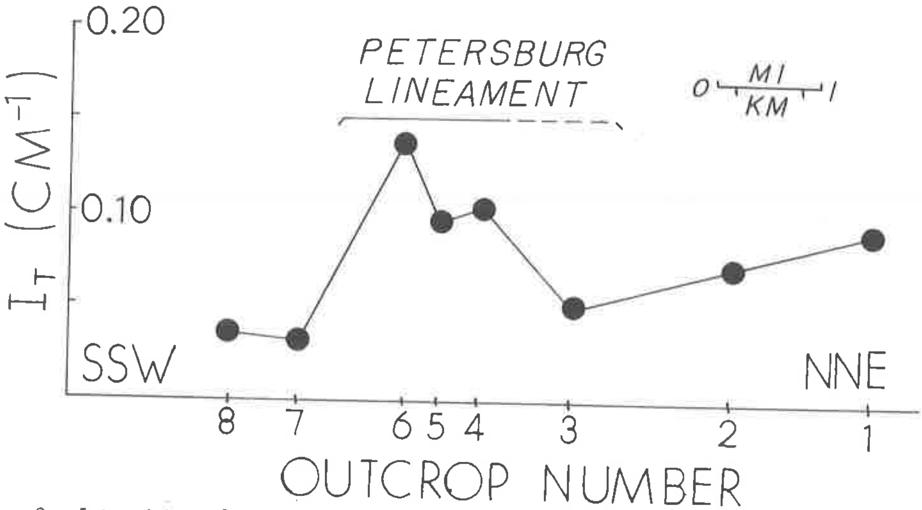

Figure 2. Intensities of systematic joints across Petersburg lineament atop Allegheny Front. Shown is westward projection of lineament, from where it is mapped in more deformed and older rocks to the east (Figure 1). Based on data of LaCaze (1978). Outcrops 1 through 8 are also located in Figure 3.

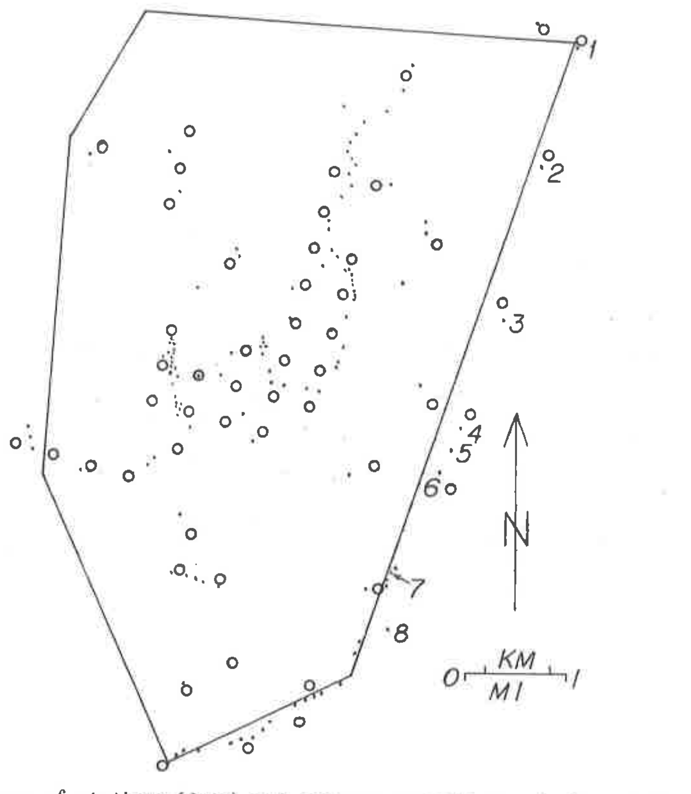

Figure 3. Locations of stations (dots) and centers of cells (circles) in Dolly Sods map area (polygon). See text for explanation of cells. 1 through 8 are outcrops for which joint intensities of Figure 2 were calculated.

Wheeler examined a false-color composite LANDSAT print (taken in October at 1:250,000), a positive LANDSAT print (at about 1:1,000,000) and color infrared images (taken in February, from about 65,000 feet elevation), and found no photolineaments passing through the high-intensity area. Thus the Petersburg lineament does appear in the Pennsylvanian rocks of the easternmost Plateau province as a zone of increased joint intensity.

Interpretation. The lineament narrows as it passes upsection, westward, or both from the Silurian and Devonian rocks of the western Valley and Ridge province to the Pennsylvanian rocks of the Plateau province. Outcrops 3 through 6 are more intensely jointed than are the other four, but only with a significance value of 0.09 (randomization test: Siegel, 1956). However, if outcrop 3 is regarded as lying northeast of the lineament, the significance value drops to 0.01. Thus the lineament at Dolly Sods includes only outcrops 4 through 6, and may be as narrow as 1.3 km (0.81 mi), rather than at least 2.7 km (1.7 mi) wide (Figure 2).

DISRUPTIONS OF SEDIMENTARY LAYERS

In Silurian, Devonian, and Mississippian rocks of the Valley and Ridge and eastern Plateau provinces, the Petersburg and other similar lineaments are expressed as disrupted patterns on strike-line maps and on contour maps of bed dip and bed strike (Trumbo, 1976; Sites, 1978; Wilson, 1979; other sources cited by Wheeler, Winslow, and others, 1979). We applied that approach to the Pennsylvanian and Upper Mississippian rocks of Dolly Sods, to see if such methods are equally effective in molasse with low dips.

Data

Many or most exposures are of cross-bedded sandstones. Using four methods, we measured or constructed orientations, not of cross-beds, but of the top or bottom contacts of cross-bedded depositional units (sedimentary layers: A. C. Donaldson, oral communication, April 1979). (1) LaCaze measured strike and dip of layers directly, with a Brunton compass. (2) Where several measurements on the same exposure were feasible, we use their median. (3) LaCaze also constructed three-point solutions for mapped upper contacts of thick, resistant, continuous sandstones. (4) Wheeler added layer orientations constructed from traces of layers on three faces of the same exposures (see "Layer orientations . . ." below). The four types of estimates of layer orientation were then smoothed by first weighting each estimate according to a qualitative estimate of its uncertainty, and then taking medians of the weighted estimates falling within square cells 2000 feet (610 m) on a side (Figure 3). The number and uncertainty of orientation estimates in each cell were considered in interpreting maps of the smoothed data.

Structure Contours

Figure 4 shows structure contours on three horizons. Intersections of topographic contours with geologic contacts mapped by LaCaze (1978, Plate 1) on a 7-1/2 minute topographic base define contours atop the Upper Mississippian Princeton sandstone and the uppermost Lower Pennsylvanian Homewood sandstone, where those two members are exposed. Reger (1921) contoured the base of the uppermost Middle Pennsylvanian Upper Freeport coal. The three contoured horizons span about 350 m (1150 ft) stratigraphically (LaCaze, 1978, Figure 11). All three horizons show the trough of the Stony River syncline (Figure 4), and the upper two show structural relief of 500 to 700 feet (152 to 213 m). Dips calculated from Figure 4 range from about 10 degrees in the northwest part of the map area to a more representative three degrees along the area's eastern edge. For comparison, median layer dips for direct measurements, three-point solutions, and constructions from layer traces are five, one, and eight degrees, respectively.

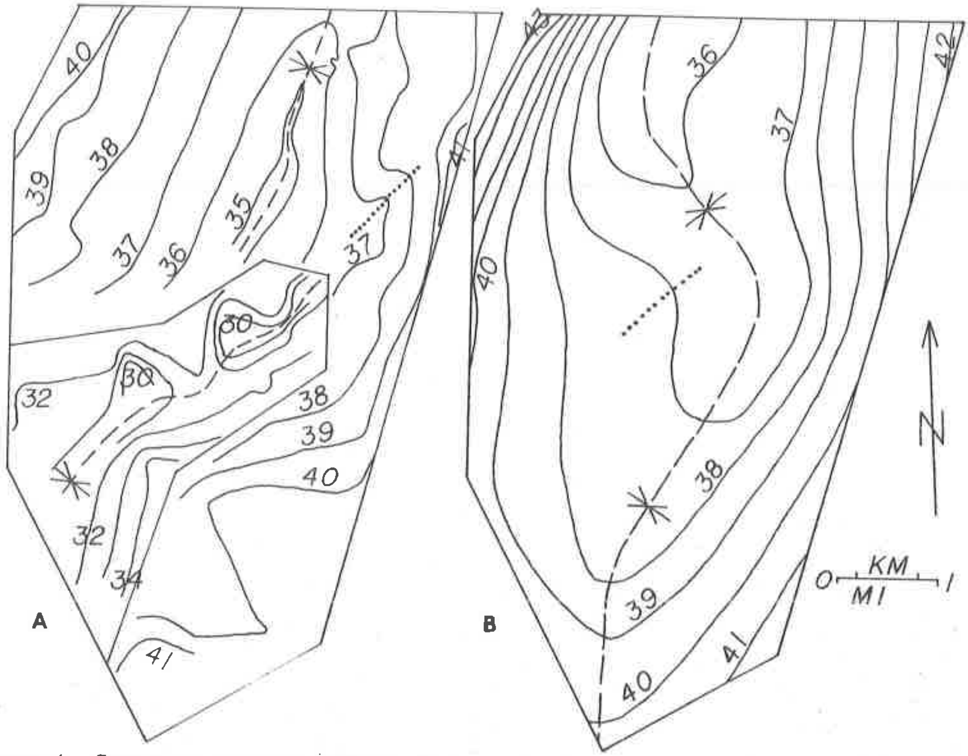

Figure 4. Structure contours in Dolly Sods map area. Contour interval is 100 feet (30 m). Contours are labeled in hundreds of feet. Dashed lines are trough lines of Stony River syncline, inferred from contours. Dotted lines are possible positions of parts of the hypothesized Breathed Mountain anticline, inferred from contours (see text). a) Eight-sided polygon in southwest encloses contours atop Princeton sandstone. Other contours are atop Homewood sandstone. Based on mapping by LaCaze (1978, Plate 1; see text). b) Contours atop Upper Freeport coal, traced from Reger (1921).

In the center of the map area, Figure 4 shows a saddle in the trough at Princeton level, and anomalous bends in the trough line at that and Upper Freeport level. Indeed, Cardwell and others (1968) and LaCaze (1978, Plate 1) show two northeast-trending, en echelon synclines. LaCaze interprets them as separated by a northeast-trending anticline which he names the Breathed Mountain anticline. The structure contours of Figure 4 are consistent with both existence and nonexistence of the anticline. In either case, map-scale structure is disrupted in the center of the map area, in a crudely east-trending zone about three km (2 mi) wide, and including outcrops three through six of the joint intensity work (Figures 2 and 3).

Structure contours atop deeper horizons are not detailed enough to determine the vertical extent of that disruption (Haught, 1968; Cardwell, 1971, 1973, 1976). LaCaze (1978) suggests that the Breathed Mountain anticline has 75 m (246 ft) of closure and speculates that it may be formed over a detachment no deeper than the Upper Devonian.

Other Maps of Layer Orientations

Methods. We drew maps of (1) strike lines, and contours on numerical values of (2) strike, (3) absolute dip, and (4) angular deviation from a pi axis of N35E/02NE fitted by hand to all direct measurements of layer orientation. Those maps drawn by both of us independently are similar, and resemble computer-drawn maps of the same data. Thus our maps are both objective and reproducible, properties that should be demonstrated for hand-drawn contour maps (Dahlberg, 1975, 1979).

Table 1. Summary of Results From Maps and Longitudinal Sections of Layer Orientations.

	Data Type	Map or Section?	Anomaly Type	Anomaly Width	Statistically Significant?
1.	Strike lines	Map	Disrupted zone	1.6 km	No
-	Strike lines	Section	None	-	-
2.	Strike lines	Map	Synclinal nose	5.5 km	Yes
3.	Strike lines	Section	Synclinal nose	3.6 km	No
-	Strike-value contours	Map	None	-	-
4.	Absolute dip contours	Map	High dips	4.2 km	No
5.	Absolute dip contours	Section	High dips	3.6 km	No
6.	Angular deviations	Map	High dispersion	2.6 km	Almost
7.	Angular deviations	Section	High dispersion	3.6 km	No

Results. Results are summarized in Table 1. On the strike-line map, Wheeler chose a disrupted zone 1.6 km (1.0 mi) wide measured in a north-south direction, and a broad synclinal nose 5.5 km (3.4 mi) wide. He chose no disruptions on the map of strike-value contours. On the map of absolute dip he chose a high-dip zone 4.2 km (2.6 mi) wide, and on the map of angular deviations, a high-dispersion zone 2.6 km (1.6 mi) wide. All but the 1.6 km-wide disrupted zone also appear on longitudinal sections made by projecting cell values (Figure 3) onto the east edge of the map area. In the longitudinal sections, all three anomalous zones have widths of 3.6 km (2.2 mi) and overlap but do not coincide. The resulting seven anomalous zones on maps or sections thus vary in width and location, but all are crudely coincident with the Petersburg lineament's westward projection and with the zone of high joint intensity.

Evaluation. However, the seven anomalous zones are subtle and subjective, so tests of statistical significance are necessary to demonstrate their validity. To insure that the probability of obtaining one or more spuriously significant results does not exceed the standard value of 0.05, we select a significance level of 0.05/7 = 0.007 (by the Bonferroni inequality: Miller, 1966; Wheeler and Holland, 1978). The appropriate tests are the Siegel-Tukey and Kruskal-Wallis (Conover, 1971; Siegel, 1956).

The 5.5 km-wide synclinal nose on the strike-line map is clearly not a random pattern: the significance value is 0.0007. The 2.6 km-wide zone of high dispersions on the map of angular deviations approaches but does not achieve significance: the significance value is 0.0233. However, neither result adds much to the information already obtained from the structure contour map (Figure 4). None of the five other anomalies on the maps and longitudinal sections of layer orientations is remotely significant: significance values exceed 0.1.

Summary. For shallowly-dipping Pennsylvanian rocks of the eastern Plateau province, maps of layer orientations do not appear to provide information not more easily extracted from structure contour maps. Structure contour maps either already exist (Reger, 1921, and similar maps), or can be generated relatively fast using aerial photographs, maps of strip mines in known coals, and carefully designed field work to map tops of resistant sandstones and continuous coals. Even form lines, such as those of Shumaker (1974), may suffice to locate map-scale structural disruption characteristic of CSD's like the Petersburg lineament.

LAYER ORIENTATIONS IN CROSS-BEDDED ROCKS

Problem. As a by-product of this work, we found that a field technique in routine use in structural analysis of multiply-folded metamorphic rocks provides a fast, precise way

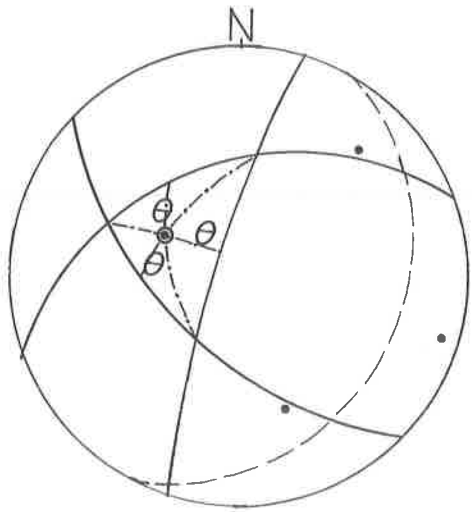

Figure 5. Constructing layer orientation from three traces (see text). Dots are traces of a layer on exposure faces. Solid curves are great circles perpendicular to the traces. Circled dot is center of gravity of the spherical triangle, constructed as the intersection of the dash-dot curves (great circles connecting triangle vertices to bisectors of opposite triangle sides). Dashed curve is estimated orientation of the layer. Precision is mean of the three values of θ, *which are exaggerated here for clarity.*

to estimate orientations of irregular layers in the Pennsylvanian sandstones of the Plateau province. Large cross-beds and coarse grains often make location of a smooth, representative, directly measurable layer surface difficult, even in large exposures or when using an aluminum plate. Low dips make directly-measured strikes uncertain by 15 degrees or more (Woodcock, 1976). Averaging several independent direct measurements may solve the problem, but is often infeasible in small or weathered exposures.

Method. However, our experience is that the eye can usually detect the traces of parallel sedimentary layers on at least two and usually three faces of an exposure. The traces need not be of the same layer, particularly if the three can be measured on faces at moderate to high angles to each other. By closing one eye to eliminate parallax, sight along each trace and measure its trend and plunge. The eye smooths irregularities in traces several meters (yards) long.

Then fit a great circle to the three traces by constructing a beta axis (Figure 5; Ramsay, 1967, p. 12-14). That is, plot the traces in lower-hemisphere spherical projection. Great circles perpendicular to each trace define a spherical triangle. Construct the triangle's center of gravity as the intersection of the three great circles that connect each vertex with the bisector of the opposite side of the triangle. The great circle perpendicular to the center of gravity estimates the orientation of the layer or layers, and the mean of the angles between the center of gravity and the three side bisectors gives the precision of that estimate.

Evaluation. Wheeler estimated layer orientations for nine outcrops of Homewood sandstone along the east border of the map area. The nine precisions have a median of three degrees. That is about the minimum achievable using a Brunton compass and careful hand plotting and construction. Median precision of two or more direct measurements on each of 29 exposures is four degrees. Thus, for the rocks we studied, the beta axis method produces more precise results than does direct measurement. Because layers are irregular and the layer traces are visually averaged over large parts of an exposure, the beta axis method is also probably more accurate than direct measurement.

CONCLUSIONS

1. The Petersburg lineament extends west from the Valley and Ridge province, and appears in the Pennsylvanian rocks atop Allegheny Front as a zone of roughly doubled intensity of systematic joints at least 1.3 km (0.81 mi) wide.

2. In general, the lineament does not occur in the Pennsylvanian rocks of the easternmost Plateau province as a zone of disrupted sedimentary layers that can be efficiently mapped, and so is unlikely to do so further west in still less deformed rocks. Thus it and other CSD's probably can best be extended west using joint intensity, structure contours, subsurface and geophysical data, and aerial and satellite images. The CSD's may have very narrow expressions in Pennsylvanian rocks, so close station spacing and careful planning of fieldwork are advisable.

3. Even in coarse-grained, weathered, cross-bedded Pennsylvanian sandstones, orientations of shallowly-dipping sedimentary layering can be estimated precisely and accurately if constructed from several traces. Even with dips of 10 degrees or less, precise and accurate strikes can be constructed.

ACKNOWLEDGMENTS

The manuscript was improved by suggestions of R. Sites. The Eastern Gas Shales Project of the Department of Energy, Morgantown Energy Technology Center, supported Wheeler through contract EY-77-C-21-8087, Task Order No. 25. The West Virginia Geological and Economic Survey supported, and the United States Forest Service and West Virginia Department of Natural Resources otherwise assisted, LaCaze's fieldwork. Computer work was done with a modification of WYLBUR, a text-editing system developed at the Stanford University Computation Center and the National Institutes of Health, Bethesda, Maryland.

REFERENCES

Anonymous, ed., 1978, Eastern Gas Shales Symposium, 2d, Morgantown, W.Va., Oct. 16-18, 1978, Proceedings, v. I, 454 p., and v. II, 163 p., METC/Sp-78/6; available from National Technical Information Service, Springfield, Va.

Cardwell, D. H., 1971, Structural geologic map of West Virginia; datum: top of Williamsport (Newburg) Sandstone, in Cardwell, D. H., The Newburg of West Virginia: West Virginia Geological and Economic Survey, Bull. 35, 54 p.

Cardwell, D. H., 1973, Deep well and structural geologic map of West Virginia, in Cardwell, D. H., 1974, Oriskany and Huntersville gas fields of West Virginia: West Virginia Geological and Economic Survey, Mineral Resources Series, v. 5, 139 p.

Cardwell, D. H., 1976, Structural geologic map of West Virginia; datum: top of Ordovician, in Cardwell, D. H., 1977, West Virginia gas development in Tuscarora and deeper formations: West Virginia Geological and Economic Survey, Mineral Resources Series, v. 8, 34 p.

Cardwell, D. H., Erwin, R. B., and Woodward, H. P., compilers, 1968, Geologic map of West Virginia: Morgantown, W. Va., West Virginia Geological and Economic Survey. Scale 1:250,000.

Conover, W. J., 1971, Practical nonparametric statistics: New York, Wiley, 462 p.

Dahlberg, E. C., 1975, Relative effectiveness of geologists and computers in mapping potential hydrocarbon exploration targets: International Association for Mathematical Geology, Journal, v. 7, p. 373-394.

Dahlberg, E. C., 1979, Relative effectiveness of geologists and computers in mapping potential hydrocarbon targets (abs.), in Anonymous, ed., Appalachian Petroleum Geology Symposium, 10th, Program and Abstracts, Morgantown, W.Va., March 6-9, 1979: West Virginia Geological and Economic Survey, Circular 15, p. 8.

Dixon, J. M., 1979a, A method to identify zones of intense jointing with application to the Parsons lineament, West Virginia (abs.): Geological Society of America Abstracts with Programs, v. 11, no. 4, p. 177.

______________, 1979b, Techniques and tests for measuring joint intensity: West Virginia University, Ph.D. dissertation, 143 p.; available from University Microfilms, Ann Arbor, Mich.

Haught, O. L., 1968, Structural contour map on the Greenbrier Limestone in West Virginia: Morgantown, West Virginia Geological and Economic Survey. Scale 1:500,000.

Kulander, B. R., Dean, S. L., and Barton, C. C., 1977, Fractographic logging for determination of pre-core and core-induced fractures--Nicholas Combs No. 7239 well, Hazard, Ky.: MERC/CR-77/3, 44 p; available from National Technical Information Service, Springfield, Va.

Kulander, B. R., Barton, C. C., and Dean, S. L., 1978, The application of fractography to core and outcrop fracture investigations: METC/SP----79/3, 174 p.; available from National Technical Information Service, Springfield, Virginia.

LaCaze, J. A., Jr., 1978, Structural analysis of the Petersburg lineament in the eastern Appalachian Plateau province, Tucker County, West Virginia: West Virginia University, M.S. thesis, 69 p.

Miller, R.G., Jr., 1966, Simultaneous statistical inference: New York, McGraw Hill, 272 p.

Ramsay, J. G., 1967, Folding and fracturing of rocks: New York, McGraw-Hill, 568 p.

Reger, D. B., 1921, Map II of Tucker County showing general and economic geology, in Reger, D. B., 1923, Tucker County: Morgantown, W.Va., West Virginia Geological and Economic Survey (county reports), 542 p.

Rousell, D. H., and Everitt, R. A., 1978, Jointing in the Sudbury Basin, Ontario: in O'Leary, D., ed., International Conference on Basement Tectonics, 3d, Durango, Colo., 1979, Proceedings: Denver, Colo., Basement Tectonics Committee (in press).

Schott, G. L., Overbey, W. K., Jr., Hunt, A. E., and Komar, C. A., eds., 1977, Eastern Gas Shales Symposium, 1st, Morgantown, W.Va., Oct. 17-19, 1977, Proceedings, MERC/SP-77/5, 783 p.; available from National Technical Information Service, Springfield, Va.

Shumaker, R. C., 1974, Coal form-line structure: Morgantown, W.Va., West Virginia Geological and Economic Survey.

Shumaker, R. C., and Overbey, W. K., Jr., eds., 1976, Devonian shale--production and potential: Appalachian Petroleum Geology Symposium, 7th, Morgantown, W.Va., Proceedings, MERC/SP-76/2, 271 p.; available from National Technical Information Service, Springfield, Va.

Siegel, S., 1956, Nonparametric statistics for the behavioral sciences: New York, McGraw-Hill, 312 p.

Sites, R. S., 1978, Structural analysis of the Petersburg lineament, central Appalachians: West Virginia University, Ph.D. dissertation, 274 p.; available from University Microfilms, Ann Arbor, Mich.

Sites, R. S., 1979, Structural analysis of the Petersburg lineament, central Appalachians (abs.): Geological Society of America Abstracts with Programs., v. 11, no. 4, p. 212.

Stearns, D. W., 1968, Certain aspects of fracture in naturally deformed rocks: in Riecker, R. E., ed., NSF advanced science seminar in rock mechanics: Bedford, Mass., Terrestrial Sci. Lab., Air Force Cambridge Research Labs., v. 1, p. 97-116.

Trumbo, D. B., 1976, The Parsons lineament, Tucker County, West Virginia: West Virginia University, M.S. thesis, 81 p.

Tukey, J. W., 1977, Exploratory data analysis: Reading, Mass., Addison-Wesley, 688 p.

Vialon, P., Ruhland, M., and Grolier, J., 1976, Elements de tectonique analytique: Paris, Masson, 118 p.

Wheeler, R. L., 1978a, Cross-strike structural discontinuitites: possible exploration tool in detached forelands (abs.): Geological Society of America Abstracts with Programs, v. 10, no. 4, p. 201.

______________, 1978b, Fracture intensity predictions for eastern gas shales: Morgantown, W.Va., Dept. of Energy, Morgantown Energy Technology Center, Eastern Gas Shales Project Open File Rept., Document 137, 78 p.

______________, 1978c, Cross-Strike structural discontinuities: Possible exploration tool for natural gas in Appalachian overthrust belt (in review).

______________, 1979, A numerical estimate of joint intensity, confidence limits on it, and applications to large structural lineaments (abs.): Geological Society of America Abstracts with Programs, v. 11, no. 1, p. 58-59.

Wheeler, R. L., and Dixon, J. M., 1980, Intensity of systematic joints: methods and application: Geology, v. 8, p. 230-233.

Wheeler, R. L., and Holland, S. M., 1978, Style elements of systematic joints: an analytic procedure with a field example, in O'Leary, D., ed., International Conference on Basement Tectonics, 3d, Durango, Colo., 1978, Proceedings: Denver, Colo., Basement Tectonics Committee (in press).

Wheeler, R. L., Holland, S. M., Dixon, J. M., and Kulander, B. R., 1979, Style elements, intensity and relative ages of joints on a detached anticline (abs.): Geological Society of America Abstracts with Programs, v. 11, no. 1, p. 59.

Wheeler, R. L., Shumaker, R. C., Woodfork, L. D., and Overbey, W. K., Jr., 1976, Gas from Devonian shales: Geotimes, v. 21, no. 10, p. 18-19.

Wheeler, R. L., Winslow, M., Horne, R. R., Dean, S., Kulander, B., Drahovzal, J. A., Gold, D. P., Gilbert, O. E., Jr., Werner, E., Sites, R., and Perry, W. J., Jr., 1979, Cross-strike structural discontinuities in thrust belts, mostly Appalachian: Southeastern Geology, v. 20, p. 193-203.

Wilson, T. H., 1979, Bedding orientation contours of Middle Devonian shales in the Middle Mountain syncline, Valley and Ridge province, W.Va. (abs.): Geological Society of America Abstracts with Programs, v. 11, no. 1, p. 60.

Wilson, T. H., and Wheeler, R. L., 1974, Structural geology of the Plateau-Valley and Ridge transition, Grant County, West Virginia: West Virginia Academy of Science, Proceedings, v. 46, p. 204-211.

Woodcock, N. H., 1976, The accuracy of structural field measurements: Journal of Geology, v. 84, p. 350-355.

THE MISSISSIPPIAN-PENNSYLVANIAN SYSTEMIC BOUNDARY IN EASTERN KENTUCKY: DISCUSSION AND REPLY

Discussion

By

Charles L. Rice
U.S. Geological Survey
Reston, Virginia 22092

Haney (1979, p. 54), in describing the Mississippian-Pennsylvanian systemic boundary along the Pottsville escarpment of eastern Kentucky, posed the following question: "Is the boundary unconformable or is it a conformable transition from marine Mississippian to nonmarine Pennsylvanian rocks?" He suggested in his introductory paragraphs that the unconformity has been taken for granted since it was first identified by Campbell in 1898 by stating: "Without exception, United States Geological Survey maps of eastern Kentucky through 1977 show an unconformity" between the Mississippian and Pennsylvanian Systems. This last statement refers to the major cooperative project of the Kentucky Geological Survey and the U.S. Geological Survey for which, between 1960 and 1978, the State of Kentucky was mapped at a scale of 1:24,000. More than 75 geologic maps compiled by approximately 50 geologists cover the outcrop belt of the Pottsville escarpment, where the Mississippian-Pennsylvanian unconformity is poorly to well-exposed. A search of the literature shows, however, that far from accepting the unconformity as a fact, particularly after Englund and Smith (1960) had demonstrated intertonguing between the largely Mississippian Pennington Formation and the largely Pennsylvanian Lee Formation in southeastern Kentucky, some workers speculated that the systemic boundary was "the combination of a gradational and interfingering contact coupled with local disconformable contacts below channels" (Huddle and Englund, 1966, p. 26). Sheppard and Dobrovolny (1963) identified a sandstone in Upper Mississippian strata in northeastern Kentucky that was lithologically similar to sandstone of the Lee Formation and suggested that the systemic boundary there was gradational and intertonguing. Englund and Windolph (1971) later named that sandstone the Carter Caves Sandstone and showed that it was separated from overlying Pennsylvanian rocks by a major unconformity. In fact, as noted above, detailed mapping along the escarpment has shown the systemic boundary to be everywhere an unconformity.

Figure 1 shows the unconformity along the western margin of the Appalachian basin in eastern Kentucky to be a north or northwestward truncation of progressively older Mississippian strata. Of particular significance is the consequent drainage system, suggested by outcrop patterns, that was formed after emergence of the region in Late Mississippian time. About 10 separate drainage areas can be identified in Figure 1; those in the northern part of the outcrop belt drained southeast, whereas those in the southern part drained in a more southerly direction.

The presence of a widespread unconformity in northeastern Kentucky was again questioned by Horne and Ferm (1970, p. 217) who postulated that the "very erratic distribution of underlying limestone and red and green shales and of the overlying orthoquartzite" was related to a laterally migrating environmental interface, or shoreface, between primarily marine rocks below and primarily continental rocks above. Their postulation was later formalized in a model called the "Lee-Newman barrier shoreline model" (Horne and others, 1971, p. 5), which, though attractive in many ways, has been shown by Dever (1973), Rice and others (1979), and Ettensohn (1980) as inapplicable to the rocks in northeastern Kentucky. Even Haney (1979, p. 59) clearly showed in his sections of Rowan and Menifee Counties a pronounced unconformity associated with ancient channeling and paleokarst. Because the "Lee-Newman model" has not been substantiated in northeastern Kentucky (or elsewhere), its use as a premise to question the presence of an unconformity at the Mississippian-

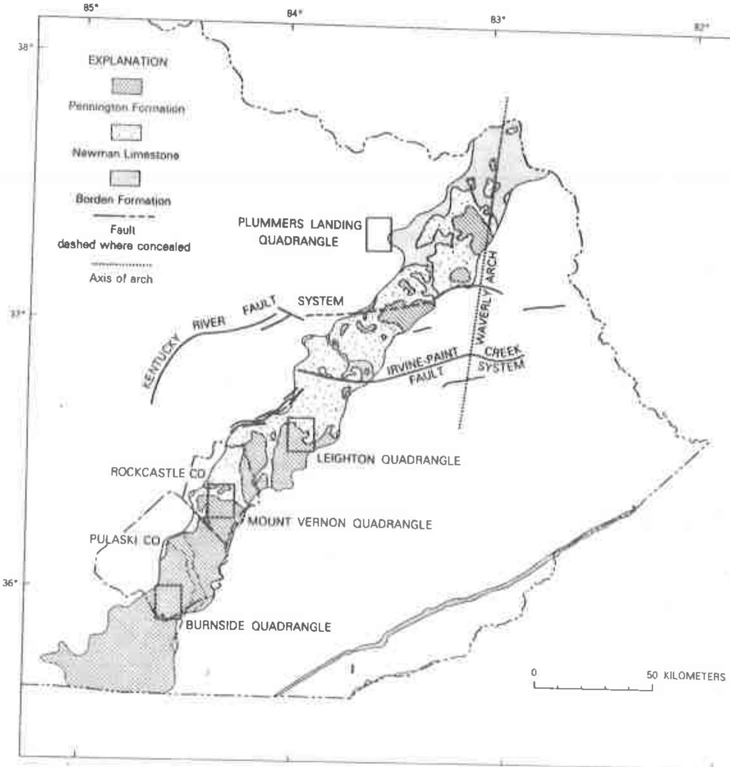

Figure 1. Generalized geologic map of the pre-Pennsylvanian surface in eastern Kentucky compiled from U.S. Geological Survey 7-1/2-minute Geologic Quadrangle maps, showing the major structural features that affected Mississippian and Pennsylvanian deposition in northeastern Kentucky. Depiction of channel just north of Irvine-Paint Creek fault system was enhanced by combining the Newman Limestone where it is thin with the Borden Formation. Outline of Pennsylvanian channel in Pulaski County is shown by dashed lines because the channel is confined within the Pennington Formation.

Pennsylvanian boundary (Haney, 1979, p. 54) is inappropriate.

Other workers questioning a widespread unconformity include Haney and Hester (1976, p. 188), Ettensohn (1977, p. 35), and Short (1978, p. 25). These workers and Haney (1979) suggested that the unconformity is restricted in northeastern Kentucky to areas of tectonism and that elsewhere the systemic boundary lies in a gradational sequence. All described the unconformity as being localized in the vicinities of the Kentucky River fault system, the Waverly arch, or the Irvine-Paint Creek fault system (Fig. 1). Haney (1979, p. 58) was more specific when he stated that "transitional boundaries occur in the areas between or away from the structural features; this statement suggests that the unconformity is localized along the traces of the structures. None attempted to explain the Pennsylvanian sandstone and conglomerate-filled channels shown in the Plummers Landing quadrangle (McDowell and others, 1971) (Fig. 1) that rest on Lower Mississippian middle Osagean strata about 40 km west of the Waverly arch and about 30 km north of the Kentucky River fault system. Also, for example, Haney and I coauthored the Leighton Geologic Quadrangle Map (Haney and Rice, 1978) (Fig. 1) which is about 12 km southeast of an extension of the Irvine-Paint Creek fault; in that report, we showed the unconformity to have a relief of about 60 m.

Haney's (1979, p. 59, 60) strongest argument for a transitional systemic boundary is made with respect to areas "remote" from the listed structural features "particularly in Rockcastle and Pulaski Counties." In those counties (Fig. 1) the uppermost Mississippian unit is the Pennington Formation, which consists primarily of

THE MISSISSIPPIAN-PENNSYLVANIAN SYSTEMIC BOUNDARY IN EASTERN KENTUCKY: DISCUSSION AND REPLY

Discussion

By

Charles L. Rice
U.S. Geological Survey
Reston, Virginia 22092

Haney (1979, p. 54), in describing the Mississippian-Pennsylvanian systemic boundary along the Pottsville escarpment of eastern Kentucky, posed the following question: "Is the boundary unconformable or is it a conformable transition from marine Mississippian to nonmarine Pennsylvanian rocks?" He suggested in his introductory paragraphs that the unconformity has been taken for granted since it was first identified by Campbell in 1898 by stating: "Without exception, United States Geological Survey maps of eastern Kentucky through 1977 show an unconformity" between the Mississippian and Pennsylvanian Systems. This last statement refers to the major cooperative project of the Kentucky Geological Survey and the U.S. Geological Survey for which, between 1960 and 1978, the State of Kentucky was mapped at a scale of 1:24,000. More than 75 geologic maps compiled by approximately 50 geologists cover the outcrop belt of the Pottsville escarpment, where the Mississippian-Pennsylvanian unconformity is poorly to well-exposed. A search of the literature shows, however, that far from accepting the unconformity as a fact, particularly after Englund and Smith (1960) had demonstrated intertonguing between the largely Mississippian Pennington Formation and the largely Pennsylvanian Lee Formation in southeastern Kentucky, some workers speculated that the systemic boundary was "the combination of a gradational and interfingering contact coupled with local disconformable contacts below channels" (Huddle and Englund, 1966, p. 26). Sheppard and Dobrovolny (1963) identified a sandstone in Upper Mississippian strata in northeastern Kentucky that was lithologically similar to sandstone of the Lee Formation and suggested that the systemic boundary there was gradational and intertonguing. Englund and Windolph (1971) later named that sandstone the Carter Caves Sandstone and showed that it was separated from overlying Pennsylvanian rocks by a major unconformity. In fact, as noted above, detailed mapping along the escarpment has shown the systemic boundary to be everywhere an unconformity.

Figure 1 shows the unconformity along the western margin of the Appalachian basin in eastern Kentucky to be a north or northwestward truncation of progressively older Mississippian strata. Of particular significance is the consequent drainage system, suggested by outcrop patterns, that was formed after emergence of the region in Late Mississippian time. About 10 separate drainage areas can be identified in Figure 1; those in the northern part of the outcrop belt drained southeast, whereas those in the southern part drained in a more southerly direction.

The presence of a widespread unconformity in northeastern Kentucky was again questioned by Horne and Ferm (1970, p. 217) who postulated that the "very erratic distribution of underlying limestone and red and green shales and of the overlying orthoquartzite" was related to a laterally migrating environmental interface, or shoreface, between primarily marine rocks below and primarily continental rocks above. Their postulation was later formalized in a model called the "Lee-Newman barrier shoreline model" (Horne and others, 1971, p. 5), which, though attractive in many ways, has been shown by Dever (1973), Rice and others (1979), and Ettensohn (1980) as inapplicable to the rocks in northeastern Kentucky. Even Haney (1979, p. 59) clearly showed in his sections of Rowan and Menifee Counties a pronounced unconformity associated with ancient channeling and paleokarst. Because the "Lee-Newman model" has not been substantiated in northeastern Kentucky (or elsewhere), its use as a premise to question the presence of an unconformity at the Mississippian-

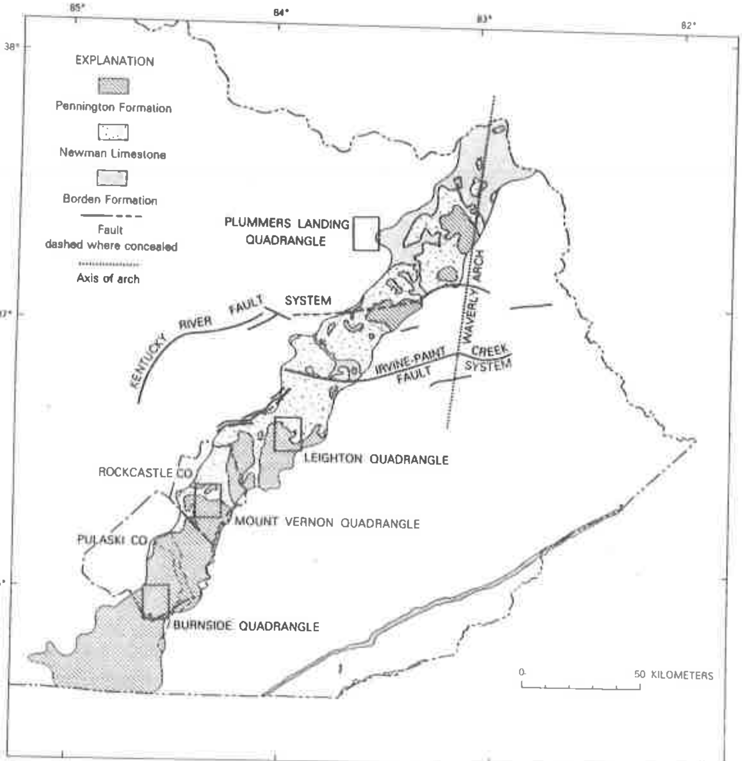

Figure 1. Generalized geologic map of the pre-Pennsylvanian surface in eastern Kentucky compiled from U.S. Geological Survey 7-1/2-minute Geologic Quadrangle maps, showing the major structural features that affected Mississippian and Pennsylvanian deposition in northeastern Kentucky. Depiction of channel just north of Irvine-Paint Creek fault system was enhanced by combining the Newman Limestone where it is thin with the Borden Formation. Outline of Pennsylvanian channel in Pulaski County is shown by dashed lines because the channel is confined within the Pennington Formation.

Pennsylvanian boundary (Haney, 1979, p. 54) is inappropriate.

Other workers questioning a widespread unconformity include Haney and Hester (1976, p. 188), Ettensohn (1977, p. 35), and Short (1978, p. 25). These workers and Haney (1979) suggested that the unconformity is restricted in northeastern Kentucky to areas of tectonism and that elsewhere the systemic boundary lies in a gradational sequence. All described the unconformity as being localized in the vicinities of the Kentucky River fault system, the Waverly arch, or the Irvine-Paint Creek fault system (Fig. 1). Haney (1979, p. 58) was more specific when he stated that "transitional boundaries occur in the areas between or away from the structural features; this statement suggests that the unconformity is localized along the traces of the structures. None attempted to explain the Pennsylvanian sandstone and conglomerate-filled channels shown in the Plummers Landing quadrangle (McDowell and others, 1971) (Fig. 1) that rest on Lower Mississippian middle Osagean strata about 40 km west of the Waverly arch and about 30 km north of the Kentucky River fault system. Also, for example, Haney and I coauthored the Leighton Geologic Quadrangle Map (Haney and Rice, 1978) (Fig. 1) which is about 12 km southeast of an extension of the Irvine-Paint Creek fault; in that report, we showed the unconformity to have a relief of about 60 m.

Haney's (1979, p. 59, 60) strongest argument for a transitional systemic boundary is made with respect to areas "remote" from the listed structural features "particularly in Rockcastle and Pulaski Counties." In those counties (Fig. 1) the uppermost Mississippian unit is the Pennington Formation, which consists primarily of red and green shale containing thin beds of fossiliferous limestone and dolomite. As representative of Rockcastle County, he showed a section measured at or near Mount Vernon, where the Pennington Formation is about 6 m thick. The Mount Vernon Geologic Quadrangle Map (Schlanger and Weir, 1971) (Fig. 1) showed that the thickness of the Pennington Formation ranges from 0 to about 25 m. About 3.5 km west of Mount Vernon, Schlanger and Weir mapped the western edge of the Livingston paleovalley, which is 3 to 5 km wide and at least 60 m deep. The paleovalley is a southward-draining channel filled with as much as 35 m of Pennsylvanian conglomerate and sandstone (Wixted, 1977). The channel truncates the entire Pennington Formation and the upper half of the Newman Limestone.

Haney (1979, p. 59) illustrated as typical for Pulaski County a stratigraphic section exposed in Sloans Valley, where the Pennington Formation is about 10 m thick. Sloans Valley is in the Burnside Quadrangle (Fig. 1). Taylor and others (1975) showed that the thickness of the Pennington ranged from 20 to 50 m in this quadrangle. This range in thickness and the thickness shown in Haney's section suggest that Haney's section was measured at or near the base of an obscure Pennsylvanian paleochannel. Sloans Valley is about 7 km west of another major south-trending channel system containing Pennsylvanian sandstones and conglomerates (Fig. 1).

Similar paleovalleys in the Illinois basin were incised into "an otherwise nearly level plain" (Wanless, 1975, p. 78). Thus, in many places that are some distance from the paleovalleys in Rockcastle and Pulaski Counties, the Mississippian-Pennsylvanian systemic boundary is commonly a paraconformity that may be misinterpreted as a normal bedding plane or facies-related boundary. For example, in those localities where dominantly marine Mississippian red and green shales are paraconformably overlain by a Pennsylvanian coal bed and a bioturbated carbonaceous roof shale, this relationship could be interpreted erroneously as a conformable regressive sequence. A similar erroneous conclusion is also possible in other localities, where the apparent gradational nature of the systemic boundary is even more strongly suggested by red and green shale of the Mississippian Pennington Formation which contains thin beds of coal. Thus, only a knowledge of the size and extent of the hiatus at the base of the Pennsylvanian strata will clarify the nature of the relationship of the underlying and overlying rocks. As has been indicated, the hiatus is both sizable and of regional extent along the Pottsville escarpment in eastern Kentucky.

Certainly, not all the problems concerning the Mississippian-Pennsylvanian systemic boundary have been solved. Because the Pennington Formation is generally poorly exposed, little is known about its stratigraphy. We have little fossil data from rocks near the boundary. Though the unconformity seems to have a maximum relief of about 60 m along the Pottsville escarpment, the total thickness of missing Mississippian strata may be much more. Regional analyses suggest that the size of the unconformity diminishes to the south and southeast in Kentucky; however, a comprehensive picture that includes all of the Appalachian and Illinois basins and that shows the precise nature of the pre-Pennsylvanian surface in eastern United States is yet to be compiled.

Reply

By

Donald C. Haney
Kentucky Geological Survey
Lexington, Kentucky 40506

The basic point addressed by Haney (1979) is that, contrary to the commonly held view which has been expressed once again by Rice (1980), the Mississippian-Pennsylvanian systemic boundary and the Mississippian-Pennsylvanian unconformity in the outcrop belt along the Pottsville Escarpment of eastern Kentucky are not

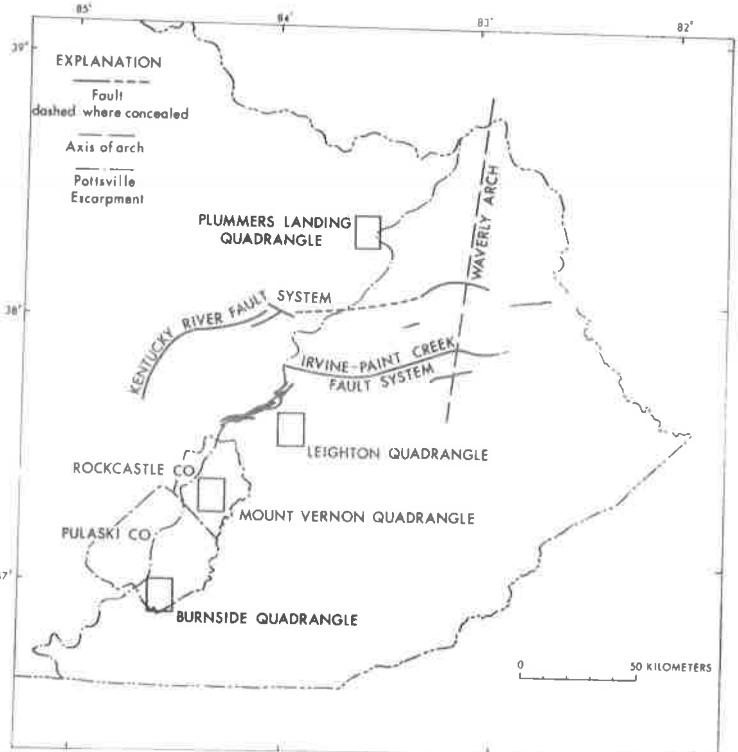

Figure 1. Sketch map of eastern Kentucky showing major structural features and locations of geologic quadrangles pertinent to the interpretation of the nature of the Pennsylvanian-Mississippian boundary.

synonymous (Fig. 1). In areas along or near tectonic elements which were active during the Carboniferous, the boundary between rocks of the Mississippian and Pennsylvanian Systems is coincident with an unconformity. Away from these tectonic elements, and generally in the southern part of the outcrop belt, field evidence indicates that not only does the magnitude of the unconformity decrease (Rice, 1980), but also that deposition was continuous across the systemic boundary in certain parts of the area. Robbs and Lumsden (1979) do not recognize an unconformity at the systemic boundary in the Cumberland Plateau of Tennessee, and Thomas (pers. comm., 1979) does not recognize an unconformity in northern Alabama. Therefore, the areas of unconformity between the Mississippian and Pennsylvanian rocks in the southern outcrop belt noted by Rice (1980) are related to intra-Pennsylvanian erosion (Fig. 2).

Rice states that Englund and Smith (1960), Huddle and Englund (1966), and Sheppard and Dobrovolny (1963) alluded to the possibility of a systemic boundary which was gradational and intertonguing and that I had ignored those authors when I questioned the interpretation of USGS workers. However, my statement was that the USGS *maps* of Kentucky through 1977 show an unconformity at the Mississippian-Pennsylvanian boundary, and no reference is made to specific authors. My statement clearly was in reference to the fifty-one 7-1/2-minute geologic quadrangle maps along the Pottsville Escarpment, and Rice agrees that they all show the "unconformity." Also, I definitely questioned the relationship of an unconformity to the *systemic boundary* and not an unconformity which may have resulted from intra-Pennsylvanian erosion which, in places, cut channels into Mississippian-age rocks (Fig. 2). Such may well be the situation of the channels described by Rice.

Rice states that McDowell and others (1971) show Pennsylvanian sandstone- and conglomerate-filled channels in the Plummers Landing quadrangle (Fig. 1) in Fleming County that " . . . rest on Lower Mississippian middle Osagian strata about 40 km north of the Kentucky River fault system." That is correct; however, the Plummers

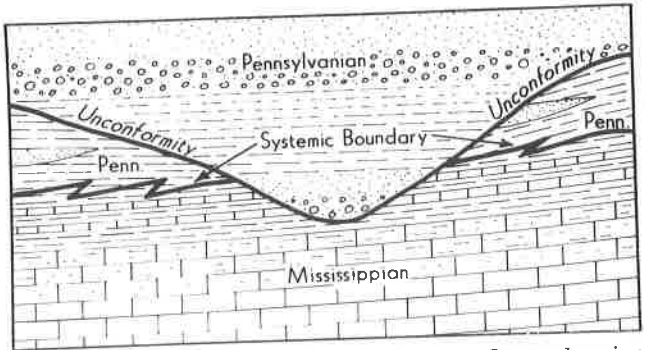

Figure 2. Diagram illustrating the effects of intra-Pennsylvanian erosion which extended into Mississippian rocks.

Landing quadrangle is on the upthrown block of the often reactivated Kentucky River fault system which has been subject to periodic erosion throughout Paleozoic time. Also, a large area of northeastern Kentucky and southeastern Ohio was most definitely high during Late Mississippian and Early Pennsylvanian times; thus channeling would be expected. The unconformity of the Leighton geologic quadrangle map (Haney and Rice, 1978) in Jackson County (Fig. 1), referred to by Rice as being located 12 km southeast of the Irvine-Paint Creek fault system, is in an area where the fault location is poorly defined and where subsurface faults associated with the Irvine-Paint Creek fault may exist. The presence of several small oil fields in the area supports such an interpretation.

Rice also questions the interpretation of a transitional systemic boundary in the Mount Vernon quadrangle in Rockcastle County (Fig. 1) where he states that the Pennington is not present. The Pennington here is absent in the vicinity of a structural feature which trends southeastward and is well demonstrated by the structure contours on the Mount Vernon geologic quadrangle map (Schlanger and Weir, 1971). However, a well-developed Pennington sequence is recognized south of the structural feature. Dever (1973), Haney (1975), and Rice (1972) recognize abrupt thinning and varied distribution of Mississippian carbonate units in the Leighton, Mount Vernon, and Alcorn quadrangles, all of which are near structural features--common features in areas along or near tectonic elements which were active during Carboniferous time.

I would again like to stress that my paper was primarily concerned with the systemic boundary. I do not dispute the presence of a major erosional unconformity over much of the study area; however, the unconformity is not necessarily at the systemic boundary. It, in part, developed during Early Pennsylvanian time, thus not affecting the boundary in certain areas "remote" from structural features.

In conclusion, it appears that east-northeast trending structures in central and eastern Kentucky have significantly affected deposition at various times during the Carboniferous and that the area along and north of the Irvine-Paint Creek and Kentucky River fault systems was most affected (Fig. 1). As acknowledged by Rice, the magnitude of erosion diminishes to the south and southeast in Kentucky.

My earlier conclusions (Haney, 1979) remain valid:

> The nature of the Mississippian-Pennsylvanian systemic boundary depends upon where the observer views the rocks. In eastern Kentucky, an unconformity clearly exists where deposition has been interrupted by tectonism; where the deposition was not affected by tectonic activity, a conformable sequence exists. Thus, in eastern Kentucky the Mississippian-Pennsylvanian systemic boundary is both conformable and unconformable.

COMBINED REFERENCES CITED

Campbell, M.R., 1898, Description of the Richmond quadrangle (Ky.): U.S. Geological Survey Geologic Atlas Folio 46.

Dever, G.R., Jr., 1973, Stratigraphic relationships in the lower and middle Newman Limestone (Mississippian), east-central and northeastern Kentucky (M.S. thesis): Lexington, Univ. Kentucky, 121 p.

Dever, G.R., Jr., McGrain, P., and Moody, J.R., 1979a, St. Louis Limestone of eastern Kentucky, in Ettensohn, F. R., and Dever, G. R., Jr., eds., Carboniferous geology from the Appalachian basin to the Illinois basin through eastern Ohio and Kentucky. Field trip no. 4, Ninth International Congress of Carboniferous Stratigraphy and Geology: Lexington, Univ. Kentucky, p. 171-174.

Dever, G.R., Jr., Hester, N. C., Ettensohn, F.R., and Moody, J. R., 1979b, Newman Limestone (Mississippian) of east-central Kentucky and Lower Pennsylvanin slump structures, in Ettensohn, F. R., and Dever, G. R., Jr., eds., Carboniferous geology from the Appalachian basin to the Illinois basin through eastern Ohio and Kentucky. Field trip no. 4, Ninth International Congress of Carboniferous Stratigraphy and Geology: Lexington, Univ. Kentucky, p. 175-181.

Englund, K. J., and Smith, H. L., 1960, Intertonguing and lateral intergradation between the Pennington and Lee Formations in the Tri-State area of Kentucky, Tennesse, and Virginia (abs.): Geol. Soc. America Bull., v. 71, no. 12, pt. 2, p. 2015.

Englund, K. J., and Windolph, J. F., 1971, Geology of the Carter Caves Sandstone (Mississippian) in northeastern Kentucky; U.S. Geological Survey Professional Paper 750-D, p. D99-D104.

Ettensohn, F.R., 1977, Effects of synsedimentary tectonic activity on the upper Newman Limestone and Pennington Formation, in Dever, G. R., Jr., and others, Stratigraphic evidence for late Paleozoic tectonism in northeastern Kentucky--Field trip, Fifth Annual Meeting, Eastern Section, Am. Assoc. Petrolum Geologists, Lexington, Kentucky, October 9, 1976: Lexington, Kentucky Geological Survey, p. 18-29.

____________, 1980, An alternative to the barrier-shoreline model for deposition of Mississippian and Pennsylvanian rocks on northeastern Kentucky--Summary: Geol. Soc. America Bull., v. 91, pt. 1, no. 3, p. 130-135.

Haney, D.C., 1979, The Mississippian-Pennsylvanian systemic boundary in eastern Kentucky: Southeastern Geology, v. 21, no. 1, p. 53-62.

Haney, D.C., and Hester, Norman, 1976, Unconformity at the Mississippian-Pennsylvanian systemic boundary in eastern Kentucky--true or false? (abs.): Geol. Soc. America Abs. with Programs, v. 8, p. 188.

Haney, D. C., Hester, N. C., and Hoge, H. P., 1975, Evidence for fault-controlled deposition of Pennsylvanian sediments in central-eastern Kentucky: Geol. Soc. America Abs. with Programs, v. 7, no. 4, p. 495-496.

Haney, D.C., and Rice, C. L., 1978, Geologic map of the Leighton quadrangle, east-central Kentucky: U.S. Geological Survey Geologic Quadrangle Map GQ-1495.

Horne, J. C., and Ferm, J. C., 1970, Facies relationships of the Mississippian-Pennsylvanian contact in northeastern Kentucky (abs.): Geol. Soc. America Abs. with Programs, v. 2, no. 3, p. 217.

Horne, J. C., Swinchatt, J. P., and Ferm, J. C., 1971, Lee-Newman barrier shoreline model, in Carboniferous depositional environments in northeastern Kentucky: Geological Society of Kentucky, Guidebook for annual spring field conference, April 1971: Lexington, Kentucky Geological Survey, p. 5-9.

Huddle, J. W., and Englund, K. J., 1966, Geology and coal reserves of the Kermit and Varney area, Kentucky: U.S. Geological Survey Professional Paper 507, 83 p.

McDowell, R. C., Peck, J. H., and Mytton, J.W., 1971, Geologic map of the Plummers Landing quadrangle, Fleming and Rowan Counties, Kentucky: U.S. Geological Survey Geologic Quadrangle Map GQ-964.

Rice, C. L., 1972, Geologic map of the Alcorn quadrangle, east-central Kentucky: U.S. Geological Survey Geologic Quadrangle Map GQ-963.

Rice, C. L., Sable, E. G., Dever,G. R., Jr., and Kehn, T. M., 1979, The Mississippian and Pennsylvanian Systems in the United States--Kentucky: U.S. Geological Survey Professional Paper 1110-F, 32 p.

Rice, C. L., and Weir, G. W., 1980, Distribution and environments of deposition of Lee Sandstones of Pennsylvanian age along the Pottsville escarpment in Kentucky: Geol. Soc. America Abs. with Programs, v. 12, no. 4, p. 206.

Robbs, Edward, and Lumsden, David, 1980, Petrology and paleoenvironments of the Late Mississippian Pennington Formation in Tennessee and Georgia: Geol. Soc. America Abs. with Programs, v. 12, no. 4, p. 206-207.

Schlanger, S. O., and Weir, G. W., 1971, Geologic map of the Mount Vernon quadrangle, Rockcastle County, Kentucky: U.S. Geological Survey Geologic Quadrangle Map GQ-902.

Sheppard, R. A., and Dobrovolny, Ernest, 1963, Mississippian-Pennsylvanian boundary in northeastern Kentucky: U.S. Geological Survey Professional Paper 450-E, p. E45-E47.

Short, M. R., 1978, Petrology of the Pennington and Lee Formations of northeastern Kentucky and the Sharon Conglomerate of southeastern Ohio (Ph.D. thesis): Cincinnati, Univ. Cincinnati, 216 p.

Taylor, A. R., Lewis, R. Q., Sr., and Smith, J. H., 1975, Geologic map of the Burnside quadrangle, south-central Kentucky: U.S. Geological Survey Gelogic Quadrangle Msp GQ-1253.

Wanless, H.R., 1975, Illinois basin region, in Paleotectonic investigations of the Pennsylvanian System in the United States; pt. 1, Introduction and regional analyses of the Pennsylvanian System: U.S. Geological Survey Professional Paper 853-E, p. 71-95.

Wixted, J. B., 1977, Sedimentary aspects of the Livingston Conglomerate--southeastern Kentucky (M.S. thesis): Lexington, Univ. Kentucky, 63 p.

ERRATUM FOR VOLUME 21, NO. 1

Structure and Tectonics of the Appalachian Miogeosyncline near the Junction of Tennessee, Kentucky, and Virginia.

Page 88.

5. Geologic evidence does not require the basement to presently be deeper on the east side of the miogeosyncline, and this suggests a westward tilt of basement during major deformation which would have added gravity to the deforming forces of the orogeny.

ERRATUM FOR VOLUME 21, NO. 2.

Statistical Study of Zircon Populations from Igneous and Metamorphic Rocks as a Method of Determining Mixed Populations.

Page 113.

$$z = \frac{x_0 (a_1 - a_2) + (b_1 - b_2)}{\sqrt{\sigma^2_{a_1} (x_0 - \bar{x}_1)^2 + \sigma^2_{a_2} (x_0 - \bar{x}_2)^2}}$$