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Abstract

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SOUTHEASTERN GEOLOGY

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SEISMIC INVESTIGATION OF THE TYBEE TROUGH AREA GEORGIA/SOUTH CAROLINA

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ABSTRACT

The Tybee Trough is a west-to-east trending topographic anomaly about 18 km in length on an otherwise relatively flat, gently sloping shelf. Interpretation of high resolution seismic data suggests the trough to be the expression of an ancient fluvio-marine feature. Correlation of seismic profiles with data from adjacent areas of the inner continental shelf and coastal plain of Georgia and South Carolina has delineated the stratigraphy of the trough and the relative dates of the processes by which it formed. Variation in subsurface morphology from the head seaward is interpreted to reflect a transition from a fluvial channel to a marine channel modified by estuarine and shelf hydrodynamic processes.

INTRODUCTION

The Tybee Trough is a broad, irregular subsurface feature on the Inner Continental Shelf adjacent to the Georgia-South Carolina border, approximately 14 nautical miles (26 kms) offshore (Figure 1). Water depths range from 8 to 28 m. In this region inner shelf is a broad (60-80 nautical mi., 96-128 kms), gently sloping (0.4m/km) surface with low topographic relief. The trough is an anomaly in this respect, with a maximum relief of 20 m, a width of 2.5 - 10 kms and a length of 18 kilometers. At present, the trough is apparently maintained by the flushing action of a flood tide dominated gyre originating to the north of the study area (Blanton, 1981, pers. comm.). Objectives of this study were to interpret the age of the trough and the strata through which it cuts, and to describe the processes of its formation.

BACKGROUND AND PREVIOUS WORK

The majority of the stratigraphic studies of the Coastal Plain of South Carolina and Georgia since the mid 1960's have involved the refining and correlation of lithologic units (Colquhoun, 1965, 1969a&b; Heron and Johnson, 1966; Duckworth and others, 1968; Furlow, 1969; Akers, 1971; Henry and others, 1973; DuBar, 1975; Herrick, 1976; Hazel, 1976; Gohn and others, 1977; and Huddlestun, 1973, 1981, and in press). In South Carolina, lithostratigraphic information was obtained from the Parris Island water wells (Siple, 1956), Clubhouse Crossroads Corehole (Gohn and others, 1977) and cores and seismic profile from Port Royal Sound (Duncan, 1972, and Colquhoun, 1972, respectively). Additional information was obtained from borings and seismic

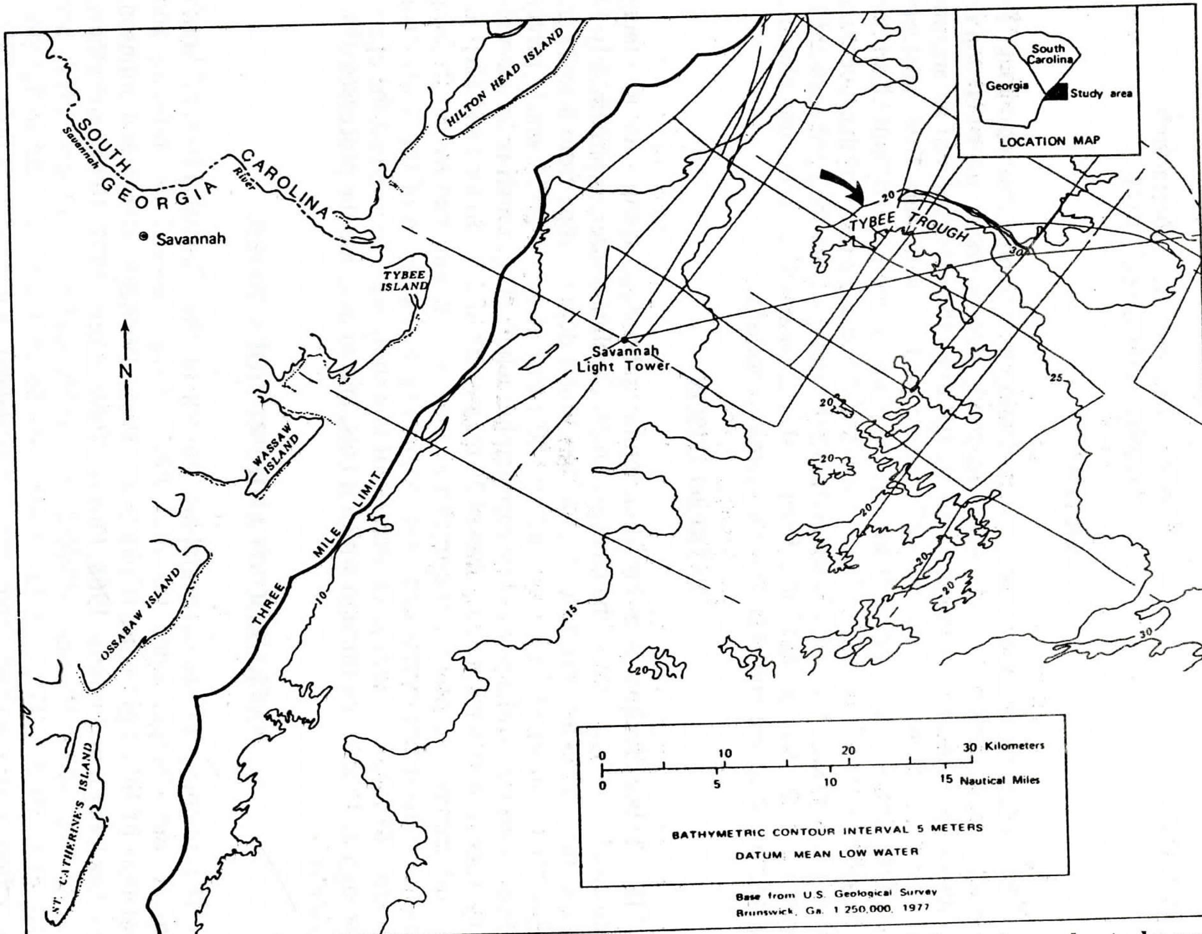


Figure 1. Location of study area. Screened lines represent seismic survey lines through study area.

profiles in the immediate vicinity of the Savannah Light Tower (see Figure 1) (Porter and Associates, 1962; McCollum and Herrick, 1964) and logs from deep cores from the 1965 J.O.I.D.E.S. expedition (Bunce and others, 1965). Figure 1 depicts seismic transects from the study area. Other data concerning offshore stratigraphy were obtained from AMCOR 6002 (Hathaway and others, 1976), COST GE-1 (Scholle, 1979), the Georgia Geologic Survey well 3426, as reported by Martinez (1981), and studies using seismic transects and cores on the inner shelf (Woolsey, 1977; Foley, 1981).

These data reveal a regional continuity of offshore and onshore sequences, and establish a basis for the correlation of the Tybee Trough area with stratigraphic nomenclature of the Coastal Plain used by the Georgia Geologic Survey (Huddlestun, in press). (Table 1).

Table 1. Seismic stratigraphy of the Tybee Trough as inferred by seismic correlation with regional stratigraphy.

SYSTEM	SERIES		STRATIGRAPHIC NOMENCLATURE (After Huddlestun, 1984)			SEISMIC CHARACTERISTICS
Quaternary	Pleistocene to Recent		Undifferentiated			Thin blanket with weak internal reflectors, discontinuous bedding and shallow buried channels
Tertiary	Pliocene		Duplin Marl equivalent			Complex channel fill, discontinuous sheet, lenses with few weak internal reflectors
	Miocene	middle	HAWTHORNE GROUP	Coosawhatchie Formation	Ebenezer Member: Charlton Member; Berryville Clay Member; Tybee Phos. Member	Strongly banded, prograding foresets distinguish upper unit, conformable strong reflector separates units
		lower			Marks Head Formation Parachucla Formation	Weak to moderate discontinuous banding, prograding foresets in Marks Head Formation
	Oligocene		Undifferentiated			Few, weak, discontinuous subparallel internal reflectors generally seismically transparent
	Eocene		Ocala Formation/ Santee Formation equivalent			Seismically transparent, with very few internal reflectors visible

In order to substantiate the interpretation of this feature, its morphology was compared with modern inlet systems. Recent work on the geomorphology of tidal inlets and fluvio-marine channel systems includes Colquhoun (1969a- b), Swift (1973), Oertel (1975), Swift and others (1980), and Vincent and Corson (1980). Studies by Oertel (1975) and Vincent and Corson (1980) deal with the configuration of tidal inlet systems and their topographic expression. Comparison of seismic profiles of the ancient trough with profiles of modern troughs reveals a similarity. Colquhoun (1969a-b), Swift (1973), and Swift and others (1980) describe the processes responsible for the morphology of fluvio-marine channel systems.

METHODS

The data used in this study were obtained as a part of the investigations of the United States southeastern inner continental shelf, conducted for the United States

Geological Survey and the Minerals Management Service by the Marine Geology Program of the University of Georgia (Henry, 1983). Cruises covering the study area took place between May 1979 and September 1980. Tracklines were following along a preplotted course using LORAN"C" navigational equipment. Trackline "shotpoints" were recorded at 10 to 15 minute intervals with simultaneous marks made on each of the analog records. Seismic profiling equipment used included a Model 225 EG&G system, a one cubic inch Bolt airgun, and a 3.5 kHz tuned transducer.

Profiles were analyzed to determine key reflectors related to lithologic changes and/or formational contacts. These reflectors were traced from borehole control points in or near the study area. Stratigraphic sequences were determined from interpretation of seismic profiles and correlated with previous studies in adjacent areas on the Continental Shelf and Coastal Plain of Georgia and South Carolina. Examples of Neogene and Paleogene units, as seen on seismic records, are shown in Figure 2. Characteristic seismic features of the delineated units are summarized in Table 1. Prominent reflectors were identified and transferred to line drawings. An approximate acoustic velocity of 1500 meters per second was used for time-to-depth conversion. Structure-contour and isopach maps were prepared representing, respectively, major erosional/nondepositional surfaces, and thickness and areal extent of the stratigraphic units defined.

Stratigraphy and Depositional History of the Tybee Trough Stratigraphy

Middle Miocene deposits comprise the greatest volume of Neogene sediments on the inner shelf and are the most areally widespread. Seismic signatures of these sediments are distinctive and are traceable on a regional scale due to the characteristic strong reflectivity of the thin banded, parallel reflectors. These strata are assigned to the Coosawhatchie Formation of the Hawthorne Group, as defined in Huddlestun (in press). Seismic records clearly show that the ancestral Tybee Trough has been cut into these strata (Figure 2).

This lithology onshore in Chatham County, Georgia, is described as dark green to brown, abundantly phosphatic, fossiliferous, micaceous marl, sandy clay, and clayey sand with an increasing clastic to carbonate ratio upward in the section (Furlow, 1969). Offshore on the shelf, the clay lithology predominates in the form of the Berryville Clay Member which grades landward into the more sandy Ebenezer Member. The middle Miocene on the inner shelf is locally divisible into two members. The two are separated by a prominent reflector apparently representing a compositional variation (Foley, 1981) with the carbonate (predominantly dolomitic) Charleton Member overlying the Berryville Clay Member (Huddlestun, in press).

The Tybee Phosphorite Member is a newly defined member of the Coosawhatchie Formation (Huddlestun, in press). Where present, it is a basal unit of the Coosawhatchie Formation, consisting primarily of massive bedded quartz sand and phosphorite, with minor clayey and dolomite.

The Tybee Phosphorite member has been seen in cores from Tybee Island and Cumberland Island, Georgia, where its thickness is 11 m and 3 m, respectively. It extends seaward at least as far as the Savannah Light Tower (McCollum and Herrick, 1964).

Depositional History

The ancestral Tybee Trough appears to have originated during a falling stage of sea level in late middle Miocene to early Pliocene (Vail and others 1977), during which time relatively large and deep channels were cut into these middle Miocene strata. It is possible that the subsequent infilling of these channels contains significant economic quantities of phosphatic lag material occurring as fill material. Extensive channeling, influenced by topography established during previous lowstands, is apparently responsible for the genesis of the ancient trough. In addition to the main channel, numerous secondary channels are seen on the seismic records (Figure 2). Due to the wide spacing of the tracklines, it is not possible at present to trace the exact nature of the drainage pattern in the trough.

The axis of the ancestral trough meanders gently and trends to the east-southeast of the seaward end. Axial length of the trough is approximately 18 kms. The width of the trough varies, from about 2.5 km at the western extremity, to 10 kms at the eastern extremity.

The head of the trough is a single, narrow channel with a shallow, steep walled configuration and a 5 m thalweg depth. The northern bank at this point resembles the cut- bank of a meander, being steeper than the more gently sloping southern "fill-bank" (Figure 3, line I). The western-most profiles, lines G, H, and I in Figure 3, reveal a fill pattern comprised of complex fill and cut-and-fill crossbeds characteristic of fluvial sedimentation. Seaward of line G the trough banks gradually steepen and the trough deepens, showing a moderately V shaped cross section on seismic profiles. As shown by lines C, E, and F, the trough becomes wider and somewhat flatter, resembling a broad, channel (Figure 3, lines E and F), interrupted at the bend by a deep scour hole (Figure 3, line D). This scour hole, the deepest section of the trough, cuts 25 m through middle Miocene and into the lower Miocene strata. Immediately seaward, the trough shallows and broadens rapidly to a thalweg depth of 7 m and a width of 4 kms. The fill structures in this portion of the trough resemble the longitudinal progradation of a southward-migrating tidal inlet sound system. Centered around this southeastern portion of the trough is a complex of smaller channels, in a band transverse to the trend of the main channel. These features are clearly seen on the seismic records in Figure 2 (C and D) and are delineated on the middle Miocene structure-contour map as a cluster of negative relief features, and on the Pliocene isopach map as thickened areas (Figure 4, A and B). These channels exhibit characteristics of an estuarine system, in that many show complex fill structures and cross bedding.

In this eastern portion, a second scour hole is present. In filling of this scour in the form of prograding sets, indicates migration of the feature and suggests that its actual size at any one time was much less than its total width of slightly less than 1 km (Figure 4, line B). This feature resembles a "tidal jet" scour hole such as those reported to be present in the throats of tidal inlets (Oertel, 1975; Vincent and Corson, 1980). Seaward (line A) the trough is a broad and shallow structure infilled by divergent fill indicative of passive sedimentation processes such as those occurring on a shallow shelf.

The variations in morphology represent, as defined by Colquhoun (1969a-b), the terrestrial and marine components of a river drainage system. According to Colquhoun's study, the steeper more V-shaped portion of a trough would

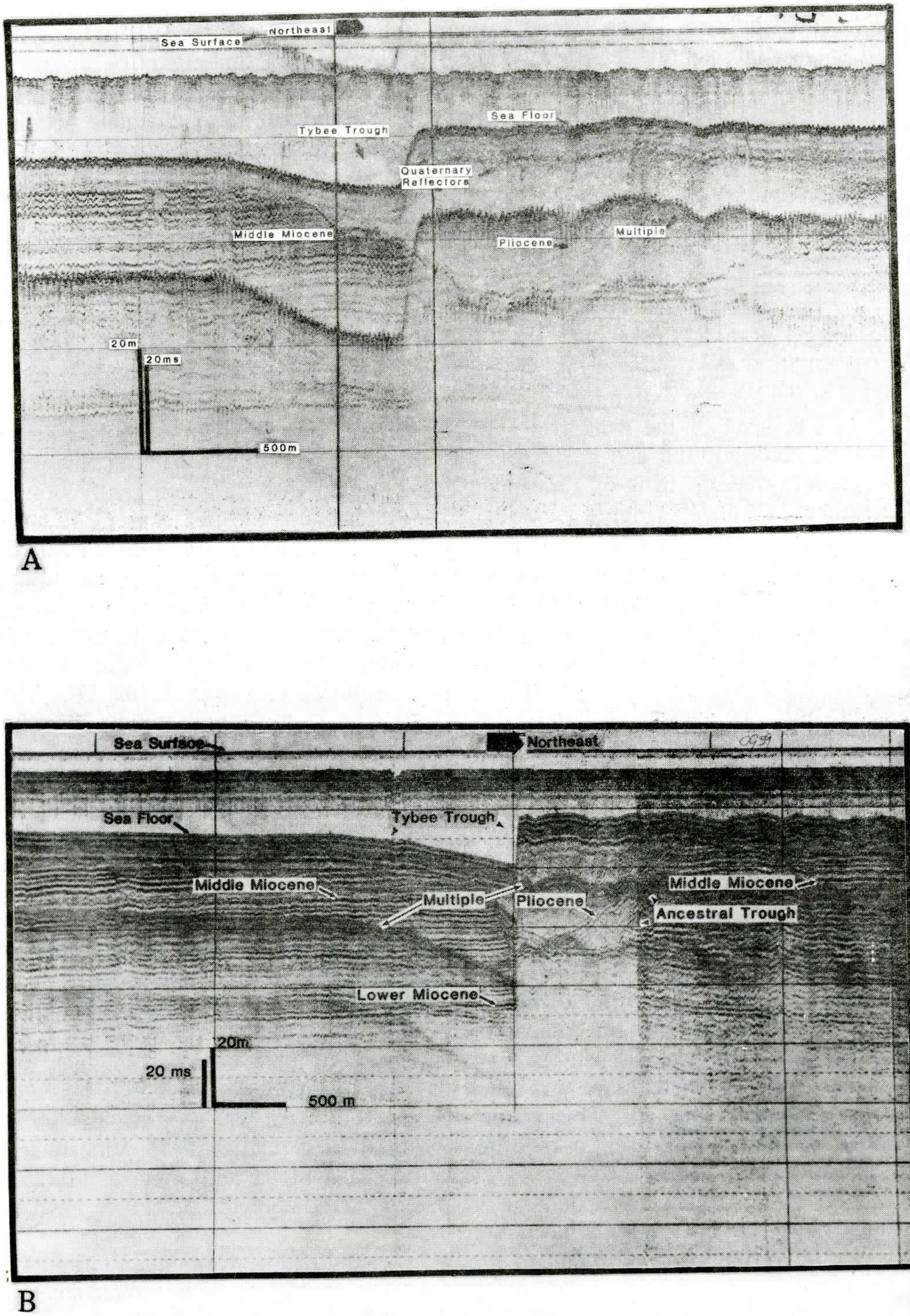
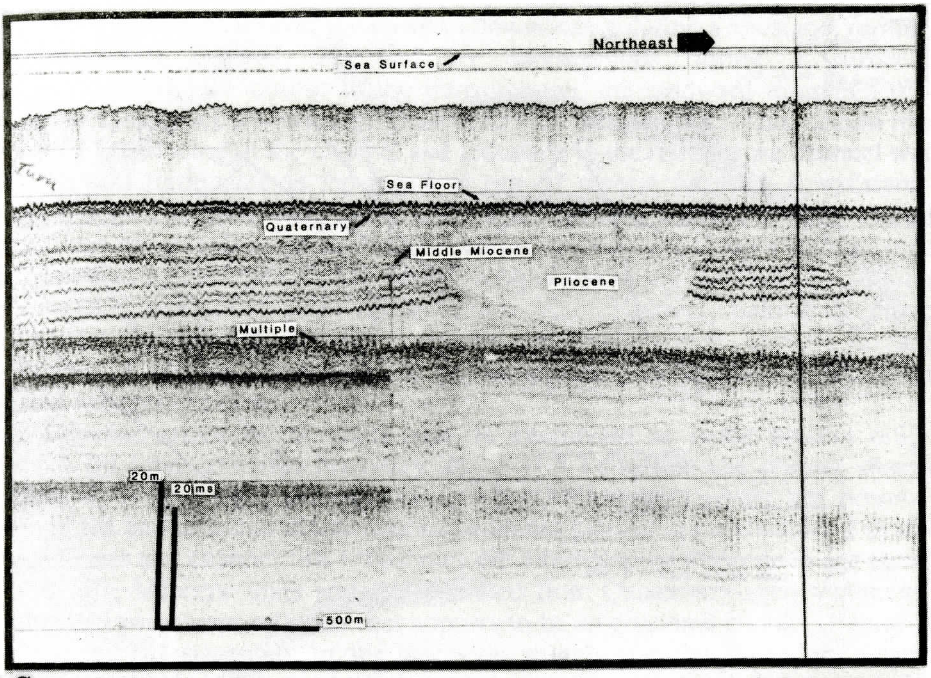


Figure 2. A&B. Seismic profiles (EG&G UNIBOOM) of the Tybee Trough area, present trough and underlying ancestral trough.



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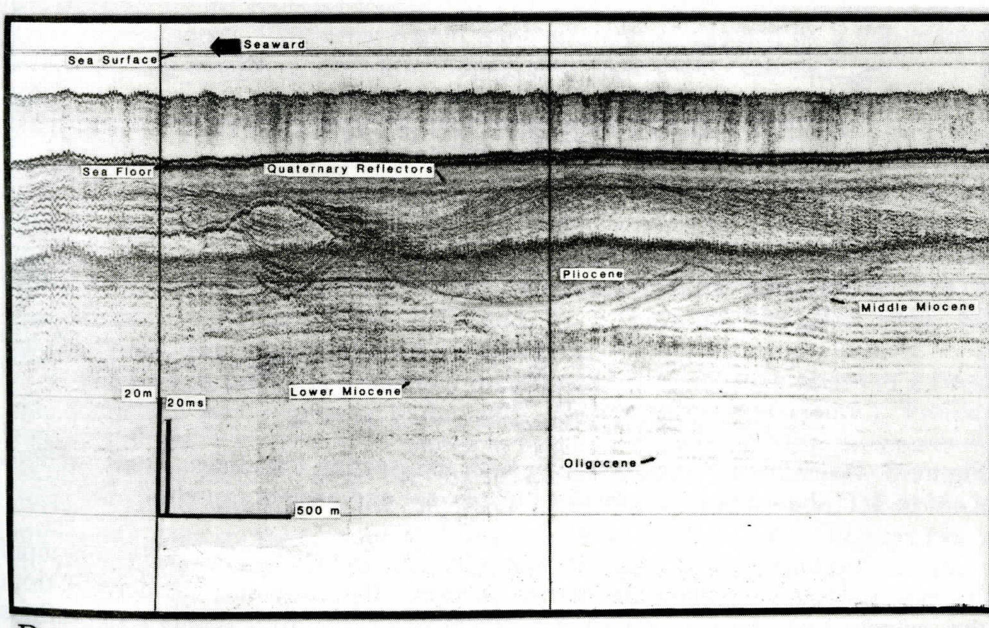

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Figure 2. C&D. Examples of secondary channels.

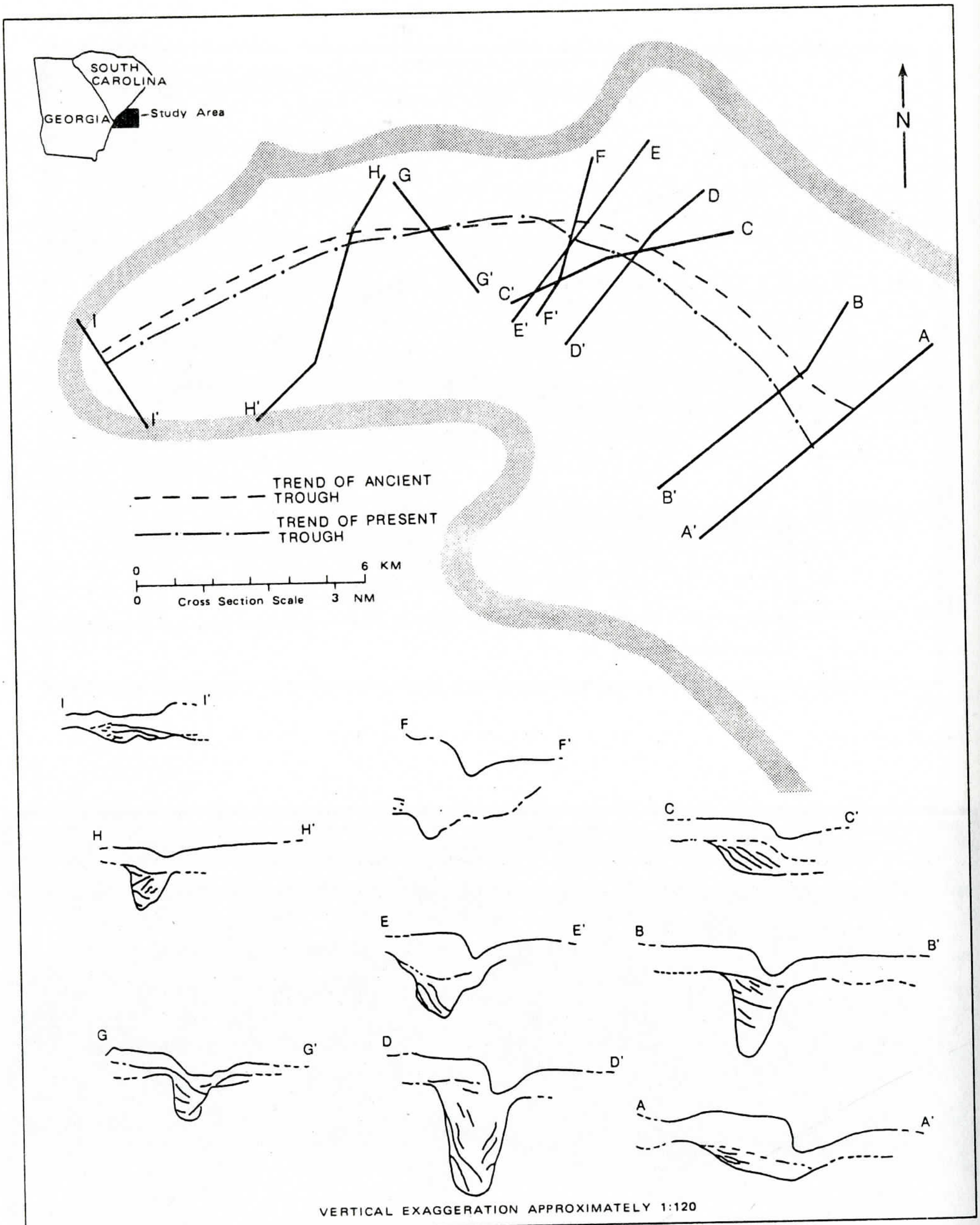


Figure 3. Plainview of Tybee Trough with intersecting tracklines. Cross sections A-A' to 1-1' show seaward variations of present and ancient troughs.

represent the terrestrial (fluvial) portion while the broader more gently contoured component would represent the marine portion. The cross sections derived from the seismic lines bear a remarkable resemblence to cross sections of various present-day tidal inlet systems detailed by Vincent and Corson (1980).

Swift (1973) and Swift and others (1980) suggest that the morphology of a submarine river channel involves a more complex process than simply downcutting of a channel followed by submergence, infilling, and burial. The bipartate morphology of the channel is the result of downcutting during a lowstand, forming a typical fluvial channel, followed by submergence during a succeeding transgression. In the course of the transgression, the landward migration of the estuarine system modifies the original nature of the channel. This occurs in the form of a broadening of the channel and reduction in the relief of the channel walls. This process terminates as the result of one of several factors. A stillstand or sealevel regression will halt in the channel's landward migration. The estuarine system can become decoupled from the channel, proceeding along a path other than that provided by the channel. Or, finally, a rapid rise in sea level could result in the overstepping and submergence of the estuary, also effectively halting the process. The latter process is a more complex sequence which appears to more satisfactorily account for the nature of the ancestral trough buried beneath the present topographic low.

Pliocene strata in the study area is interpreted as being present only as channel fill landward of a boundary 28 to 31.5 kms from the present shoreline (Figure 4, B and C). Seaward of this line, Pliocene deposits are present as channel fill and as a thin discontinuous sediment blanket approximately 2 m thick. These lenses and channel fills lithosomes are inferred to be Pliocene in age because they overlie Miocene units and are separated from Quaternary-aged sediments by a relatively prominent reflector that appears to be an erosional contact (Figure 2C). the cut-and-fill structures in the ancestral trough and the secondary channels probably formed during a relatively sustained stillstand, based on their size and internal complexity. Although, the lack of lithologic data prohibits a definitive age-dating for the channel fill, Pliocene age for the infilling is implied, in that they are overlain by the reflector interpreted, from seismic correlation, as the Pliocene-Quaternary contact. Quaternary-aged sediments are represented by a thin blanket overlying and inheriting a modified topography from older features.

TOPOGRAPHY OF THE PRESENT TYBEE TROUGH

The trend of the present trough appears to be inherited from the ancient trough. At the headward extremity, the trough overlies its ancestor. Eastward, the trough has migrated southward (Figure 3). At the seaward extent, it is offset about 2 kms from the underlying feature. An example of this offset can be seen on the seismic record in Figure 2 (a and b). These profiles also show the asymmetrical form typical of most of its length.

The present trough is apparently maintained by the flushing action of a flood tide dominated gyre originating to the north (Blanton, 1981, pers. comm.). Profiles recorded by the 3.5 kHz system shown in Figure 5, clearly showing an apparently locally derived sediment cloud moving southward over the trough during the flood cycle. Shallow reflectors show onlapping of sediment indicating the direction of migration. This southerly migration, influenced by tidal processes, is responsible for the offset of the trough from its ancestor.

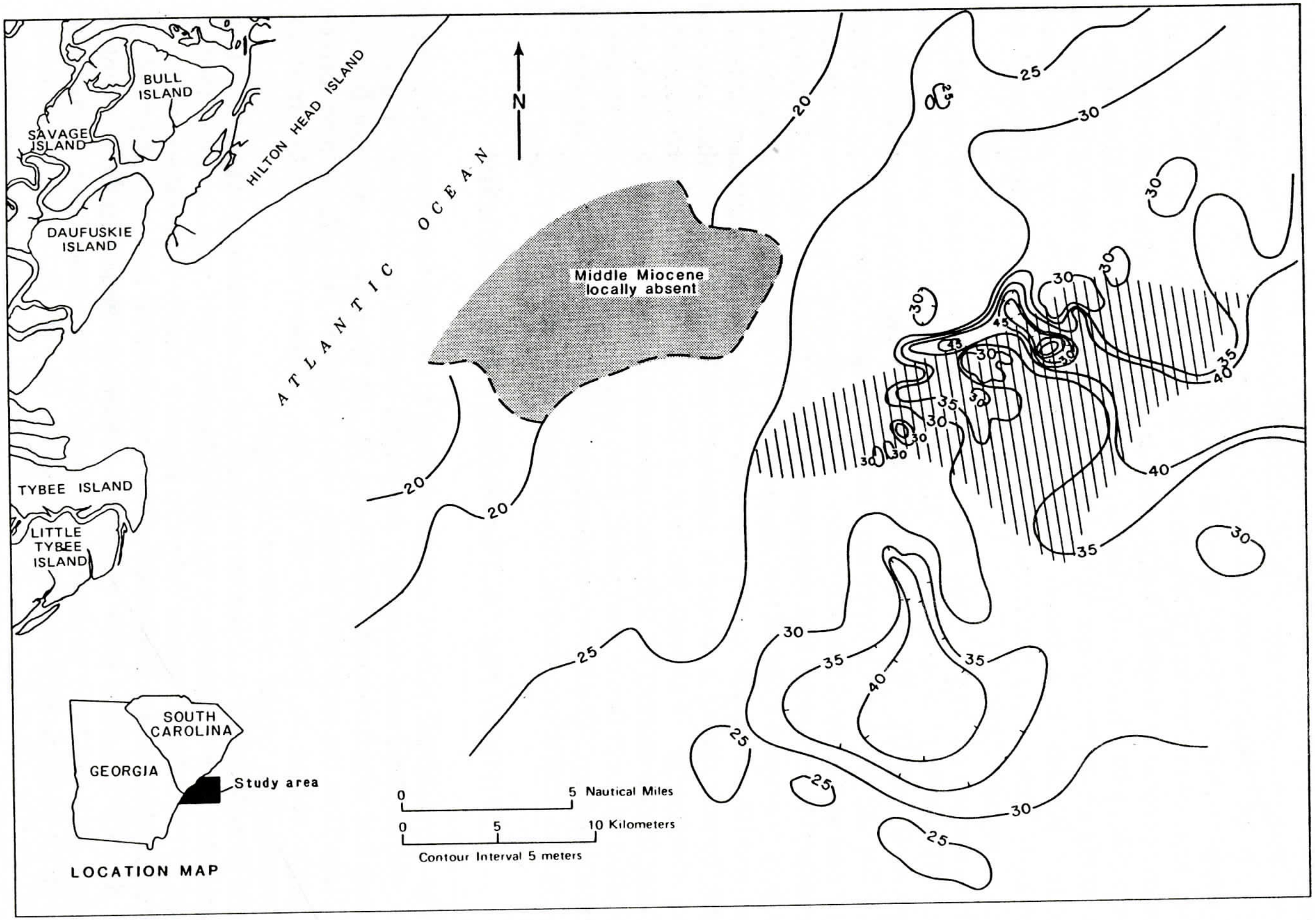


Figure 4a. Structure contour of the erosion surface of the Middle Miocene-aged sediments. Crosshatched area represents approximate location of present Tybee Trough.

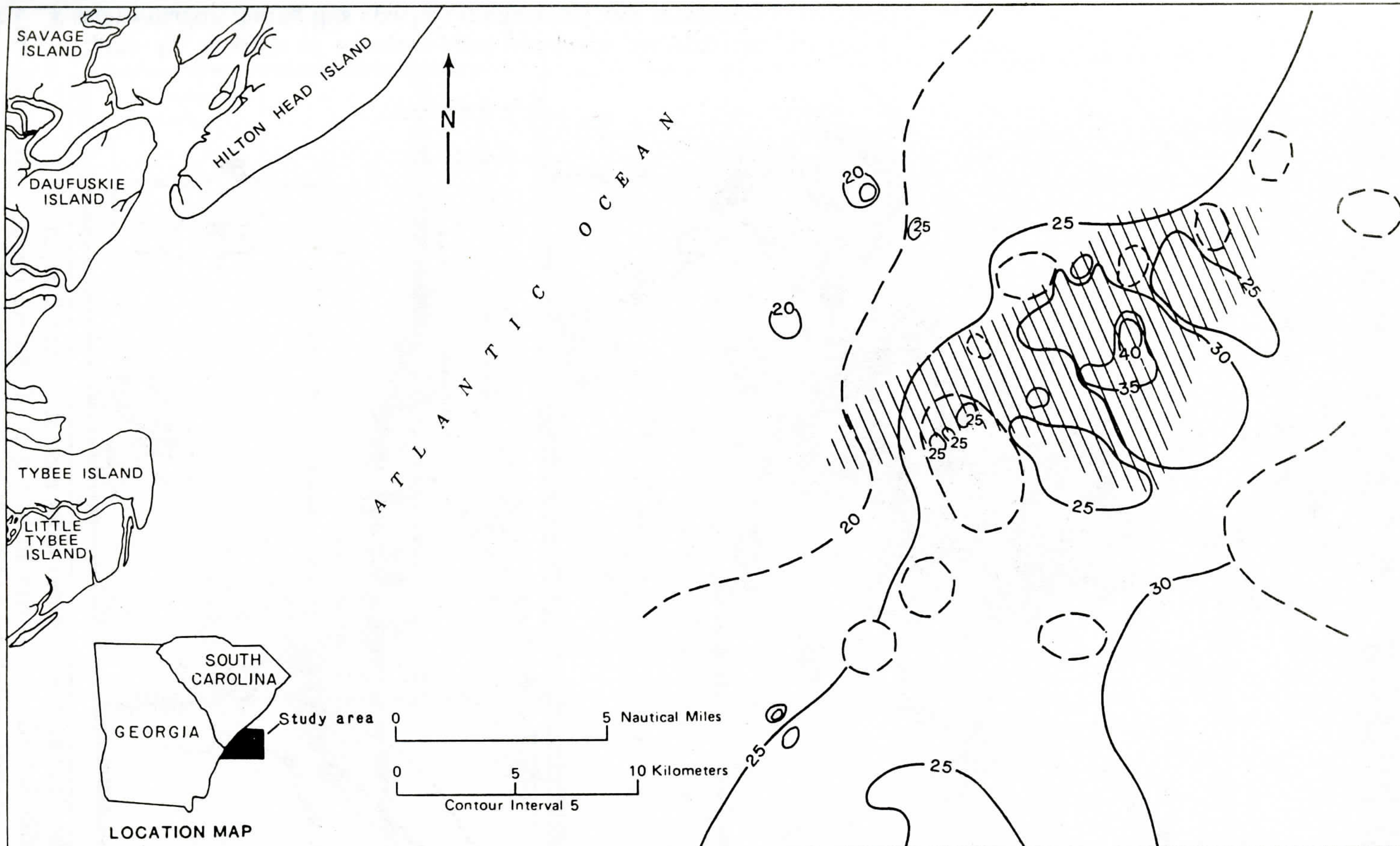


Figure 4b. Structure-contour of the top of Pliocene-aged sediments. Crosshatched area represents approximate location of present Tybee Trough.

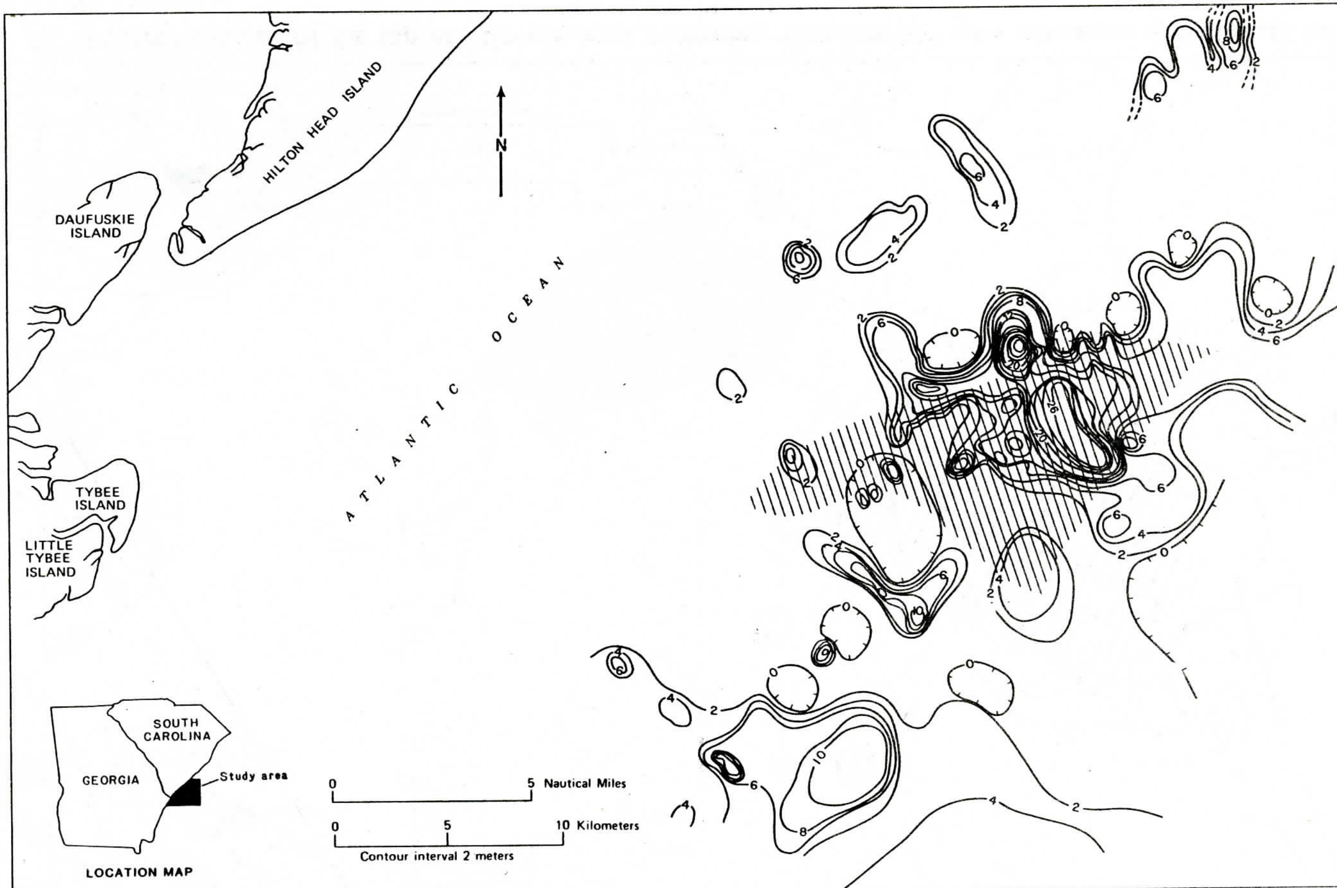
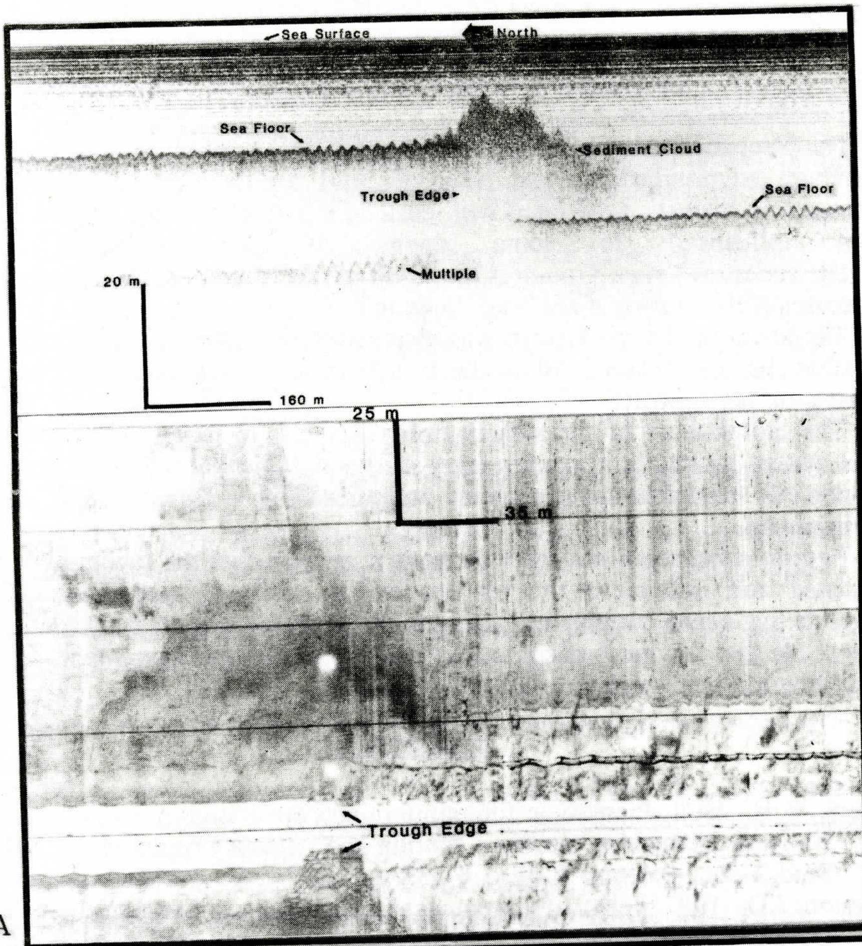


Figure 4c. Isopach of Pliocene-aged sediments showing the relation of Pliocene channel fill to the overall sedimentation during this epoch. Crosshatched area represents approximate location of present Tybee Trough.



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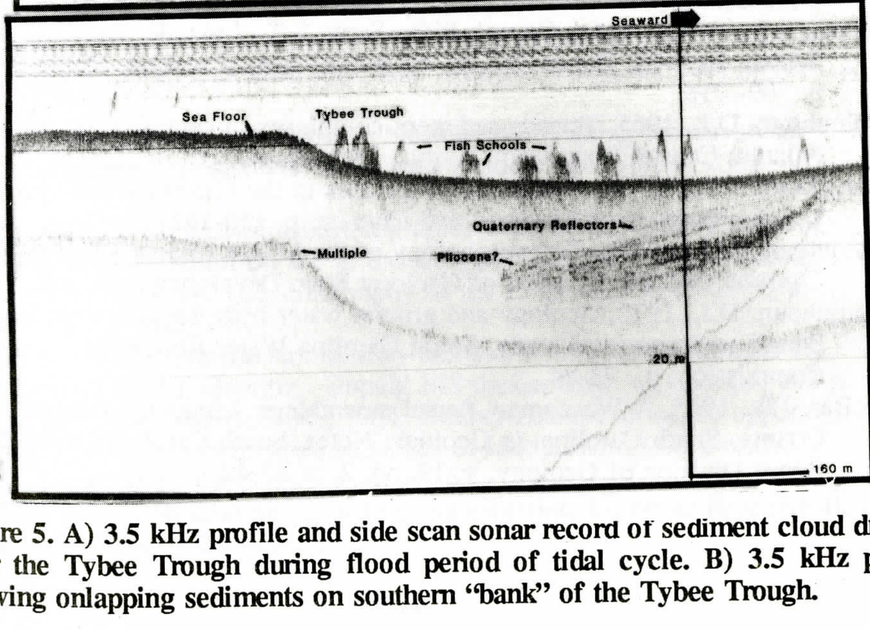

B

Figure 5. A) 3.5 kHz profile and side scan sonar record of sediment cloud drifting over the Tybee Trough during flood period of tidal cycle. B) 3.5 kHz profile showing onlapping sediments on southern "bank" of the Tybee Trough.

CONCLUSIONS

Good correlation of the stratigraphy of the continental shelf in the Trough area with that of the adjacent continental shelf and coastal plain was obtained using high resolution seismic data.

1. The Neogene stratigraphy consists of sediments deposited in shallow shelf and nearshore environments. The Neogene units are bounded by unconformities representing hiatuses associated with eustatic sea level changes (which may have been influenced to some degree by local tectonic activity).
2. The ancestral Tybee Trough is the result of late middle Miocene/early Pliocene processes active during a sea level lowstand.
3. The ancestral Tybee Trough was channelled into phosphate- rich strata. It is possible that the channel fill in the trough contains reworked and concentrated phosphatic lag deposits.
4. The morphology of the ancient trough appears to represent a tidal inlet and river system formed during a sea level stillstand. Channel floor configuration and cut-and-fill structures are interpreted as representing a fluvio-marine transitional environment.
5. Despite subsequent sea level changes it appears that the topographic low has persisted and influenced the topography of overlying sediments. Evidence suggests that ocean currents are acting to maintain the low on the present shelf. It is possible that this mechanism has served to maintain the topographic low during preceding sea level high stands.

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TROLLEITE AND SOME ASSOCIATED MINERALS IN KYANITE QUARTZITE, WILLIS MOUNTAIN, VIRGINIA

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ABSTRACT

Rare trolleite occurs in the Willis Mountain kyanite mine in Buckingham County, Virginia. Chemical and X-ray diffraction data are given for the mineral which apparently is a silician variety (SiO_2 to 5.32%). Data for an associated iron-poor, moderate yellow lazulite are also presented. Other phosphates include a woodhouseite-like species as well as apatite and variscite. All of these minerals are closely associated with paragonite, rutile, and pyrite in a kyanite-rich quartzite. Also listed are the other minerals that have been reported in this amphibolite-facies metamorphic deposit.

INTRODUCTION

Trolleite, $Al_4(PO_4)_3(OH)_3$, and several associated phosphate minerals, have recently been found in kyanite quartzite that makes up Willis Mountain, Buckingham County, Virginia (Giannini and others, 1986). Trolleite is a very rare mineral having been reported from only three other localities: Westanå near Kristianstad in Skane, Sweden (Blomstrand, 1868); White Mountains, Mono County, California (Gross and Parwel, 1968; Wise, 1977); and Rwanda, Africa (von Knorring, 1972). This paper describes the Virginia trolleite as well as some of the phosphate minerals associated with it.

Willis Mountain stands above the surrounding landscape 0.5 mile east of U. S. Highway 15, about 3.9 miles south of Dillwyn in Buckingham County, Virginia. The kyanite-rich nature of the rocks that compose this mountain was first observed by W. B. Rogers in 1835 (Rogers, 1884). The Kyanite Mining Corporation, the largest kyanite mining company in the world today, has been operating open pit mines on the top of the mountain since 1957 (Dietrich, 1962; Dixon, 1980).

TROLLEITE

Although trolleite is not abundant here several pieces have been found embedded in milky to translucent pale gray quartz. This is from the western face of the quarry, at a depth of approximately 50 feet below the original surface. The trolleite is pale green and occurs as anhedral masses to 3 cm across. It is very hard (8.5) and has a distinct conchoidal fracture. In many respects it resembles common beryl. The mineral was identified by the X-ray diffraction powder method. The averaged indexed data from four films are presented in Table 1. Using these data and the least squares method for unit cell refinement of Appleman

Table 1. X-ray data for trolleite, Willis Mountain, Virginia (nickel filtered copper radiation; camera diameter 114.6 mm).

hkl	d calculated (Å)	d observed (Å)	I obs.
110	6.68	6.67	m
011	5.02	5.03	m
400	4.65	4.65	m
211	4.20	4.21	vw
$\bar{4}11$	3.65	3.65	w
$\bar{2}02$	3.51	3.50	vw
220	3.34	3.35	w
$\bar{1}12$, 411	3.21	3.21	vs
600, 121	3.10	3.10	s
$\bar{5}21$	2.525	2.518	ms
$\bar{4}22$	2.330	2.331	vw
$\bar{2}31$	2.225	2.218	w
422	2.101	2.101	vw
910, 431, $\bar{1}32$	1.985	1.989	m-
$\bar{3}23$	1.960	1.962	m-
802	1.798	1.797	w+
532, 114, 10,11	1.677	1.677	w
822	1.605	1.605	w+
242, 12,00	1.552	1.554	m
$\bar{8}04$, $\bar{9}23$	1.545	1.545	m
134, $\overline{12}$,22	1.399	1.399	w
923	1.328	1.328	vw
$\overline{10}$,42	1.261	1.260	m

and Evans (1973), the following monoclinic cell dimensions for space group I2/c were determined: a = 18.89_5 Å, b = 7.16_4 Å, c = 7.17_2 Å, β = 99.99 . These values compare favorably with data reported for synthetic trolleite by Bass and Sclar (1979) in which a = 18.89 Å, b = 7.162 Å, c = 7.142 Å, β = 99.7 . A chemical analysis by X-ray fluorescence is compared with data for the ideal composition in Table 2. The silica content of the Willis Mountain trolleite appears to be caused by the ionic substitution of Si for P in the crystal structure. The ionic radius of Si^{4+} is 0.39 Å and that of P^{5+} is 0.35 Å so replacement of P by Si can be expected (Rankama and Sahama, 1950). This therefore appears to be a new silician variety of trolleite with the formula $Al_4(PO_4)_{3-x}(SiO_4)_x(OH)_{3-x}$ where silicate tetrahedra replace phosphate tetrahedra and there is a corresponding loss of hydroxyl in order to balance the charges. For the Willis Mountain trolleite, x was calculated to be 0.39, and the resulting theoretical analysis for this stoichiometry is also shown in Table 2. The chemical data indicate that the Si^{4+} is not apparently incorporated into this mineral by coupled substitution for both P^{5+} and Al^{3+}. The possibility that silica might have come from admixed quartz or kyanite was considered, but no evidence for this was found.

Table 2. Trolleite chemical data, theoretical and actual.

	1	2	3
Al_2O_3	45.94	47.78	46.76
P_2O_5	47.97	41.53	42.48
SiO_2		5.32	5.37
H_2O	6.09	5.20 (LOI)	5.39
Rem.		0.17	
Total	100.00	100.00	100.00

1. Theoretical composition of $Al_4(PO_4)_3(OH)_3$.
2. Willis Mountain trolleite, X-ray fluorescence chemical analysis, and loss on ignition (LOI). Rem. is CaO 0.02, Fe_2O_3 0.08, TiO_2 0.07; undetected, K_2O, MgO, MnO, Na_2O.
3. Theoretical composition of $Al_4(PO_4)_{3-x}(SiO_4)_x(OH)_{3-x}$ where x = 0.39.

LAZULITE

Amber yellow lazulite occurs in the same part of the quarry in rocks that also have considerable white kyanite and paragonite. This lazulite is of special interest because of its color and composition. It is anhedral and occurs as vitreous masses that measure to 0.5 cm by 1 cm. X-ray diffraction data, based on the averaged data of four films, are essentially identical to those reported by Blanchard and Abernathy (1981) for an Austrian lazulite. Using the least squares method for the refinement of the monoclinic unit cell with space group $P2_1/c$, the Willis Mountain lazulite was shown to have a = 7.13_8 Å, b = 7.26_9 Å, c = 7.23_9 Å, β = 120.47°. The Austrian lazulite has a = 7.152 Å, b = 7.278 Å, c = 7.233 Å, and β = 120.54°. In this study the initial high angle X-ray data indicated a lazulite with low FeO percentage, according to a method outlined by Pecora and Fahey (1950). With further work, using the more recent method of Abernathy and Blanchard (1982) based upon diffractogram charts and the relative intensities of selected reflections, it was shown that these specimens contain a mole content of 90% lazulite, $MgAl_2(PO_4)_2(OH)_2$. Thus it is much poorer in iron than most known lazulites, which perhaps accounts for its unusual moderate yellow, rather than blue, color. It is possible that within the deposit the chemical nature of the mineral is somewhat variable because another specimen brought to the attention of one of the authors (RSM) about 15 years ago, from earlier mining activities, was greenish.

OTHER PHOSPHATE MINERALS

Other specimens associated with the above phosphates show X-ray data close to the beudantite and crandallite mineral groups. Although these are quite variable in their specific compositions, some specimens are close to woodhouseite, $CaAl_3(PO_4)(SO_4)(OH)_6$, in the beudantite group. Although several X-ray fluorescence analyses indicated that these consistently contain the cations required for woodhouseite, there is considerable variability from specimen to specimen. The Ca position also may contain minor Sr, Ce, La, the Al position, minor Fe, and the P position, minor As, suggesting a variability that extends toward kemmlitzite (Sr, Ce)$Al_3(AsO_4)(SO_4)(OH)_6$. The X-ray powder data closely match both of these minerals as well as certain other members of the beudantite and crandallite groups. The hexagonal unit-cell size of one specimen, based on the averaged data from four films, using the least-squares method, is a = 7.02_6 Å, c = 16.27_3 Å. The space group is either R3m or R$\bar{3}$m. Specimens of these minerals consist of tiny (fraction of a mm) grayish orange, sparkling crystals in small cellular cavities (to 1 cm across) in paragonite-rich kyanite quartzite, usually in the same specimens as the iron-poor lazulite.

Other phosphate minerals also have been collected from the deposit. Relatively large masses, to 5 cm across, of light greenish yellow apatite are closely associated with some of the rarer phosphates described above. This apatite exhibits a strong buff to cream-colored fluorescence under shortwave ultraviolet radiation. Under longwave radiation, the colors are weaker. Kyanite associated with the apatite exhibits a deep reddish-orange fluorescence of moderate intensity under only longwave ultraviolet radiation. Tiny clear glassy apatite crystals also occur in the cellular cavities rich in the woodhouseite-like mineral.

Espenshade and Potter (1960) reported variscite as thin stringers in a drill core from this deposit. A small amount of this mineral, as tiny grains in the quartzite and as tabular crystals lining cavities, was observed and verified by X-ray diffraction work in the present study. Dietrich (1970) reported that D. J. Milton found chalcosiderite at Willis Mountain in the 1960's, but recent studies have not encountered it.

It is noteworthy that phosphate minerals similar to those found at Willis Mountain seem to be characteristic of metamorphic aluminum silicate (kyanite, andalusite, pyrophyllite, sillimanite) deposits. Two of the other three known trolleite occurrences are either with kyanite or andalusite (the third, Rwanda, Africa, is a pegmatite). Important lazulite occurrences also are with kyanite, andalusite, sillimanite, and pyrophyllite, for example, Clubbs Mountain and Graves Mountain, Georgia, the White Mountains, Mono County, California, and Westanå, Sweden. Likewise the type locality for woodhouseite is the same White Mountains deposit. Minerals related to woodhouseite, for example, crandallite, goyazite, and svanbergite, also have been observed in these aluminum silicate environments. Variscite too is frequently found in aluminum-rich metamorphic rocks.

OTHER ASSOCIATED MINERALS

The kyanite quartzite at Willis Mountain, in which these phosphates occur, is typically a medium- to coarse-grained, white to light-gray, gneissoid rock in which quartz and kyanite are somewhat segregated into lamellae. Paragonite is common and the rock also contains small disseminated pyrite and rutile crystals. Sporadic amounts of the following also have been noted at the deposit: native sulfur (Giannini and others, 1986), barite, corundum, diaspore, garnet, kaolinite, sphalerite, spinel, topaz, zircon (Espenshade and Potter, 1960), chlorite, chromian muscovite (fuchsite), goethite (iridescent), oligoclase-andesine, staurolite (Dietrich, 1970), chalcanthite, pyrophyllite (Duke, 1983), hematite, magnetite, muscovite (Rogers, 1884), and dickite.

GEOLOGIC SETTING

Conley and Marr (1980) and Marr (1981) agreed with earlier students of the deposit that the kyanite quartzite at Willis Mountain is a product of regional metamorphism of aluminous sandstones. They observed features that were interpreted as resulting from relict cross-bedding which further supported the sandstone protolith for the deposit. However, Duke (1983) disagreed with the clastic sedimentary origin of this deposit. He contended that this rock is part of a lithofacies that was derived from a late Chopawamsic volcanic episode related to the intrusion of voluminous granodiorite magma along the Chopawamsic belt. He suggested that the sedimentary structures noted by Conley and Marr (1980) and Marr (1981) are products of local winnowing and redeposition of pyroclastic volcanics prior to lithification. Duke (1983) assigned the kyanite quartzite to the Upper Member of the Chopawamsic Formation (late Precambrian or Cambrian), while Conley and Marr (1980) considered it to be part of the Arvonia Formation (Middle or Late Ordovician). The genesis and age of the quartzite is still a debated topic. Apparently the formation was folded during the Taconic orogeny, an event producing widespread metamorphism in the central Virginia Piedmont. This event produced amphibolite facies rocks in the vicinity of Willis Mountain. One should note that the petrology of the rock as well as work on the stability of trolleite by Bass and Sclar (1979) also indicate amphibolite facies metamorphism.

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ALLUVIAL FAN ORIGIN FOR TERRACE DEPOSITS OF THE SOUTHEAST PRENTISS QUADRANGLE, NEAR OTTO, NORTH CAROLINA

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ABSTRACT

Lithologies and structures in unconsolidated terrace gravels, sands, and silts formerly exposed along U.S. Highway 23-441 near Otto, North Carolina suggest deposition on an alluvial fan. Deposits at one no-longer-extant exposure included laterally extensive sheets of massive, poorly sorted, matrix-supported conglomerate interbedded with channel-filling cross-bedded sands, and tabular sheets of homogenous or poorly plane-laminated fine sands and silts. One prominent fine- sand interval exhibited a red oxidized lower portion grading upward into a white horizon. The white interval had a distinctive vertical prismatic structure strongly reminiscent of natric or salic horizons of arid alluvial soils. These deposits are interpreted as debris flow, alluvial channel, and overbank deposits (with some arid paleosol development) on local alluvial fans. Although this exposure no longer exists, several other exposures in the area contain at least some of these same features. The record of episodic sedimentation and climate change implied by the lithologies, structures, color, and the pedogenic features, is consistent with recent sedimentological, geomorphic, and paleoclimatological studies elsewhere in the southern Blue Ridge, which have suggested that climate may vary significantly on time scales approaching the mean recurrence interval of sedimentation episodes on the fan. This record also has general implications for the climatic interpretation of alluvial fan sequences. The southeastern Prentiss quadrangle is a promising area for investigating the Pleistocene-Holocene geomorphic history of the southern Blue Ridge.

INTRODUCTION

The Quaternary geomorphology and paleoclimate of the southern Appalachians are not well understood. As Clark (1979) noted, "the Appalachian Quaternary record is characterized by its occurrence in discontinuous, widely-separated depositional sites." A variety of sedimentary deposits have been utilized in paleogeomorphic and paleoclimatological studies. Lake, pond, bog, and marsh sediments have yielded valuable evidence of Quaternary vegetational and climatic change (e.g., H. Delcourt, 1979, 1985; P. Delcourt, 1985; Delcourt and Delcourt, 1985). Geomorphic evidence for and against hypotheses of alpine glaciation in the southern Appalachians also has been a topic of considerable recent debate (e.g., Raymond, 1977, 1979; Raymond and Cadwell, 1980; Haselton, 1973, 1976, 1979a,b, 1980). Some types of sedimentary deposits, however, have not yet been fully exploited for their potential record of Appalachian geomorphology and paleoclimate. Clark (1979) suggests: "Sites that may have promise for future investigation include blockfields and blockstreams and the blocky/bouldery alluvial/colluvial sediments of piedmont coves." Some such deposits have been the subjects of recent reports (e.g., Mills, 1981, 1982a,b, 1983), but studies of the paleoclimatic and geomorphological significance of southern Appalachian fluvial and alluvial deposits have not yet begun in earnest.

In the course of field studies of bedrock weathering in the southern Blue Ridge in the early 1980's, the author encountered several sedimentologically complex exposures of terraces bordering the Little Tennessee River. One of these exposures contained a particularly rich variety of sedimentary and pedogenic features, and was photographed for future reference. This exposure exhibited a range of features with potential geomorphic and paleoclimatolgical significance; some of these features are similar to features at other recently described localities in the southern Appalachians. Unfortunately, the sedimentary material has since been removed (presumably for fill). The purpose of this note is to present the existing record of the Asbury terrace exposure as a datum point for regional studies of alluvial fan and paleopedogenic features, and as a reference point for more detailed local studies of nearby similar (and possibly correlative) exposures which are still available for study.

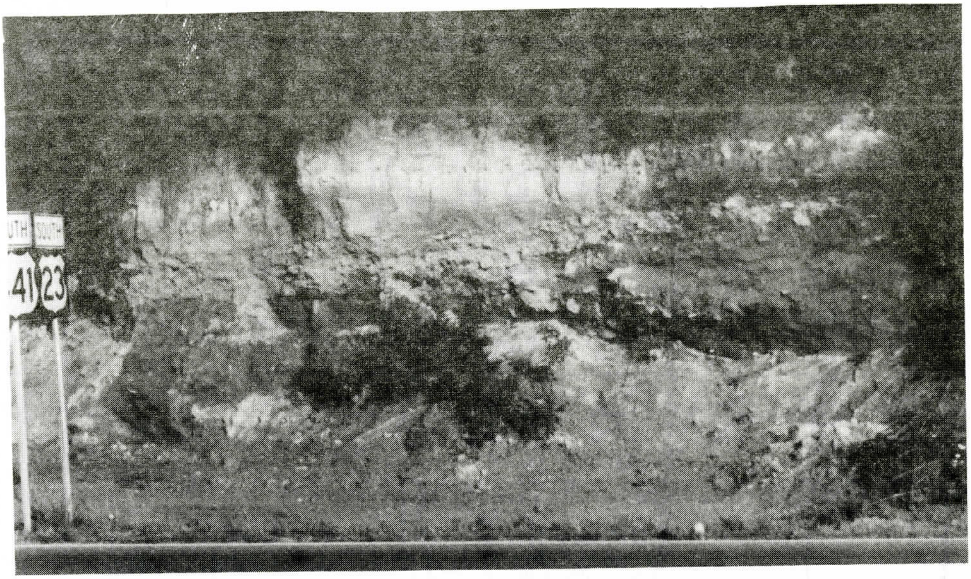
FACIES DESCRIPTION

The terrace deposit described here was formerly exposed on the west side of U.S. Highway 23-441, less than a kilometer north of Asbury Cemetary, near Otto, North Carolina. Although the terrace was a number of meters above the present level of the Little Tennessee River, topographic maps of the area show a broad low terrace at the foot of the nearby hills, with no topographic expression of fan morphology on the scale of the 20 foot topographic contour interval. The roadcut in the photographs of Figures 1-4 was a north-south face; the viewer is facing west, towards the hills which form the western boundary of the Little Tennessee River valley. This view is transverse to the inferred depositional axis of the alluvial fan deposits; flow was towards the viewer.

From bottom to top, the outcrop consisted of six units, each consisting of one or two lithofacies. These units are here classified using a facies classification terminology similar to that of Miall (1978, 1985) for units of sand and finer grain size, and Shultz (1984) for diamictites.

Unit I. Buried saprolite. This *in situ* residuum of weathered of crystalline rock is the "basement" on which the terrace was deposited. The saprolite was predominantly dark yellow-orange, and preserved clearly the schistosity and compositional banding of its parent material. There is no evidence indicating whether the alteration of the crystalline rock to saprolite occurred before or after deposition of the overlying sediment.

Unit II. Facies Dmm - Massive matrix-supported conglomerate. This unit rests unconformably upon the underlying saprolite, with a sharp, broadly irregular contact. Figures 1-3 suggest that the large-scale shape of the contact is a channel cut near the north (right) end of the outcrop. This unit is 30-40 cm thick, and appears to line, rather than completely fill, the channel in the underlying material. The conglomerate is poorly sorted; clast size ranges from sand to coarse cobble size, and clasts are subangular to subrounded. There is no discernable internal fabric, structure, or textural grading. This unit is distinguishable from its superjacent counterpart (unit IV, below) by the fact that the cobbles of unit II are



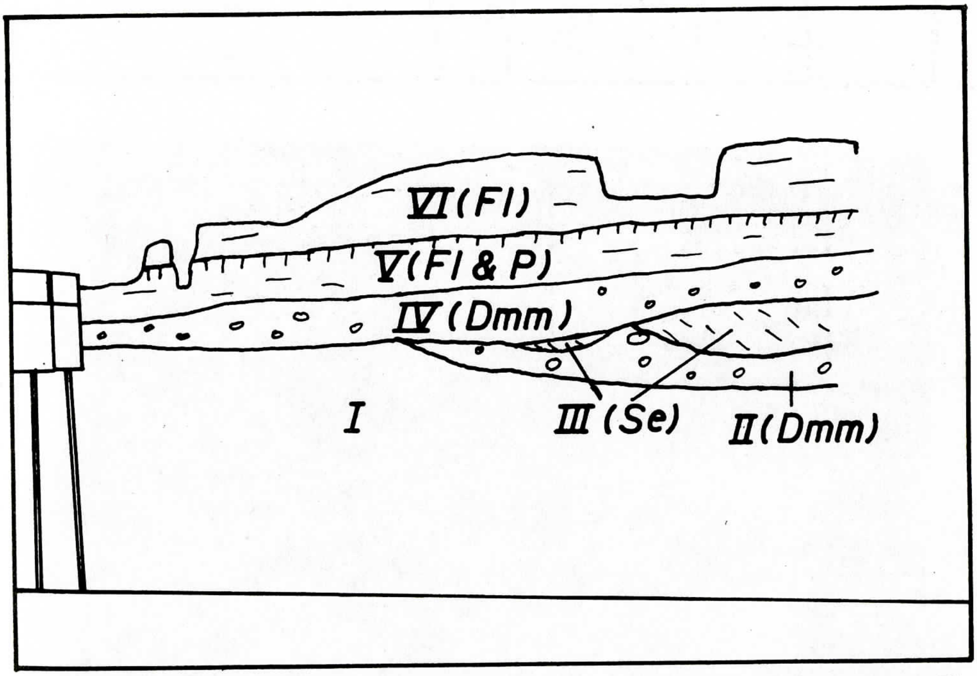


Figure 1. A. Photograph and B. interpretive sketch of Asbury terrace outcrop. Unit numbers are roman numerals; facies classification (following Miall, 1977, 1985, and Shultz, 1984) for each unit is in parantheses. Unit I - Buried saprolite (mostly covered). Unit II (Dmm) - Massive matrix-supported conglomerate. Unit III (Se) - Weakly cross-stratified sand with concave-up lower contact. Unit IV (Dmm) - Massive matrix-supported conglomerate. Unit V (Fl & P) - Red homogenous or poorly plane-laminated fine sands, with bleaching and prismatic structure at top. Unit VI (Fl) - Homogenous or poorly plane-laminated fine sands.

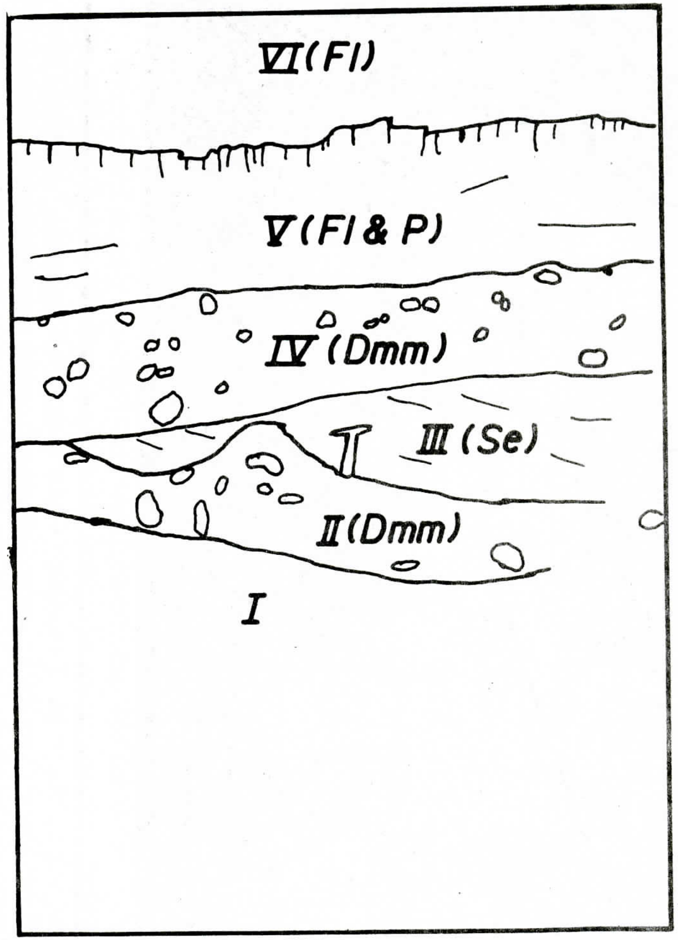


Figure 2. A. Photograph and B. interpretive sketch: close-up of right-center portion of Figure 1. See Figure 1 caption for unit and facies definitions.

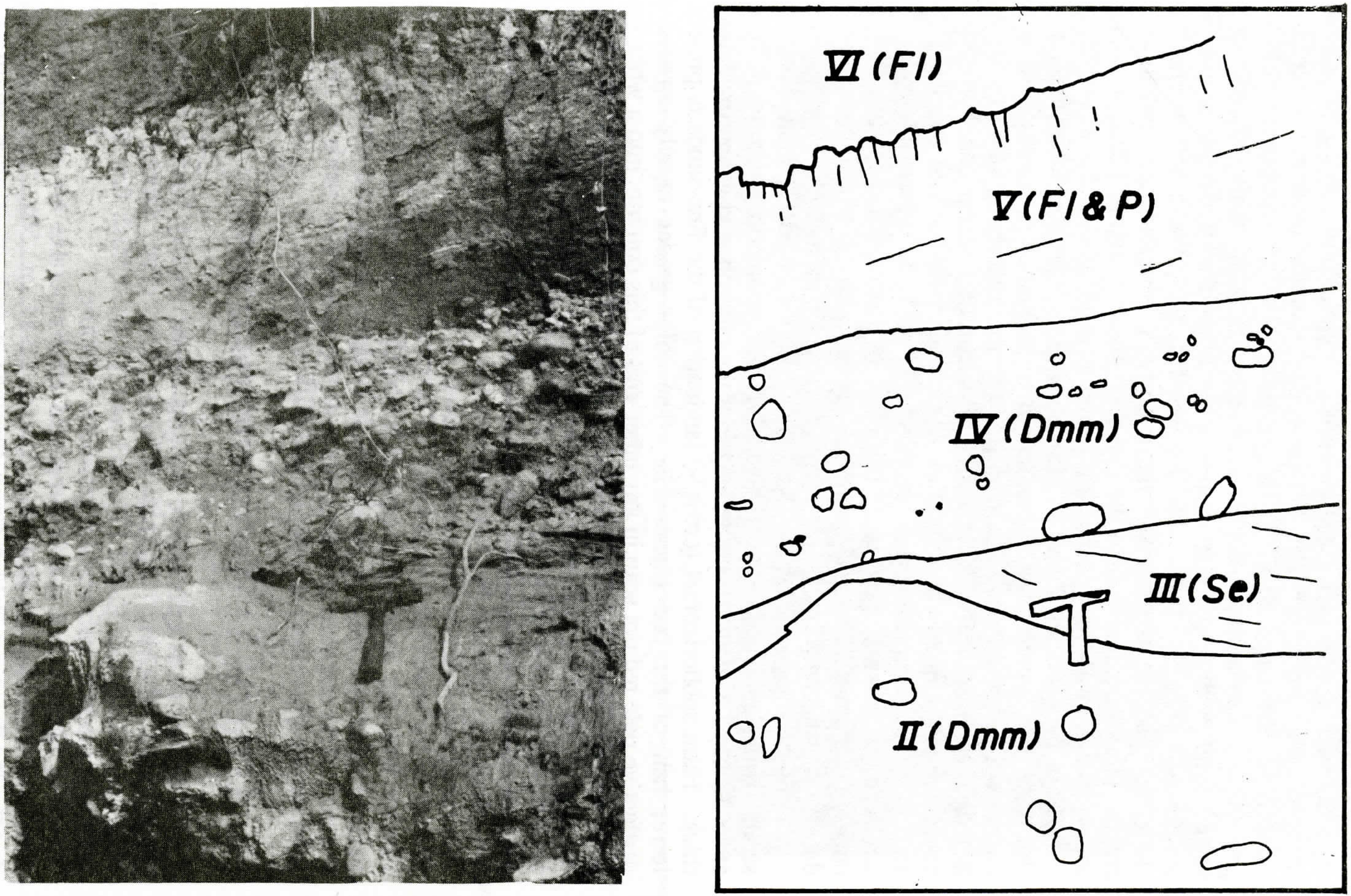


Figure 3. A. Photograph and B. interpretive sketch: close-up of upper five units (II - VI). See Figure 1 caption for unit and facies definitions.

stained, with a manganese-oxide- or desert-varnish-like stain.

Unit III. Facies Se - Weakly cross-stratified sand with concave-up lower contact (interpreted as erosional scour-fill). A lenticular body of poorly cross-stratified, well-sorted medium to coarse sand thickens towards the north, that is, toward the presumed axis of the channel. The channel axis itself was not preserved in the outcrop. A second, smaller channel or scour-fill containing well-sorted sand with weakly-developed cross-stratification occurs at the southern margin of the larger channel-fill (Figures 1 & 4). Where not covered by loose slope material, the contact with the underlying conglomerate (and its matrix) is sharp. The scoured surface of the underlying unit is stained black; in contrast, the scour- and channel-filling sands are yellow-orange.

Unit IV. Facies Dmm - Massive matrix-supported conglomerate. This sheet-like bed extends the entire length of the outcrop, maintaining a uniform thickness of about 65 cm. The contact with the underlying sands (unit III, facies Se) or conglomerate (unit II, also facies Dmm) is sharp and horizontal. The conglomerate is poorly sorted; clast size ranges from sand to coarse cobble size, and clasts are subangular to subrounded. There is no discernable internal fabric, structure, or textural grading. This bed is distinguishable from its subjacent counterpart (unit II, facies Dmm) by the absence of any manganese-oxide- or desert-varnish-like stain.

Unit V. Facies Fl & P - Red, homogenous or poorly plane-laminated fine sands, with (paleopedogenic?) bleaching and prismatic structure at top. This sheet-like bed extends the entire length of the outcrop, and is approximately 75 cm thick. Faint subhorizontal (cross?-) lamination of the fine sands is present in the lower half of the bed (facies Fl). The color grades rapidly upward from a distinctive pale red not seen in any other unit of this outcrop, into a white fine sand with well-developed vertical prismatic structure (facies P). The color change may be due to bleaching *via* remobilization of iron pigments, or it may be due to masking of the red color by white fine-grained precipitates (such as evaporite mineral efflorescences, which are common in certain types of arid soils).

Unit VI. Facies Fl - Homogenous or poorly plane-laminated fine sands. The remaining meter of the outcrop consists of a tabular bed of fine sand, similar in texture to the underlying sand (unit V) bed, but with a yellow-orange color. The contact with the underlying unit is probably non-erosive; the thickness of the vertical prismatic interval in unit V is uniform across the entire length of the outcrop. This suggests that deposition of the yellow fine sands of unit VI did not truncate the previously developed prismatic structure.

FACIES INTERPRETATIONS

The two most informative features of the Asbury terrace exposure are the unorganized (lacking internal structure or clast fabric) matrix-supported conglomerates (units II and IV), and unit V, the reddish fine sand capped by the whitened interval with vertical prismatic structure.

The unorganized matrix-supported conglomerates are interpreted as debris flows. According to Rust and Koster (1984, p. 59), "commonly preserved features" of ancient debris flows "include a lack of internal stratification and imbrication, and a sheet-like form, in contrast with the common channel forms of

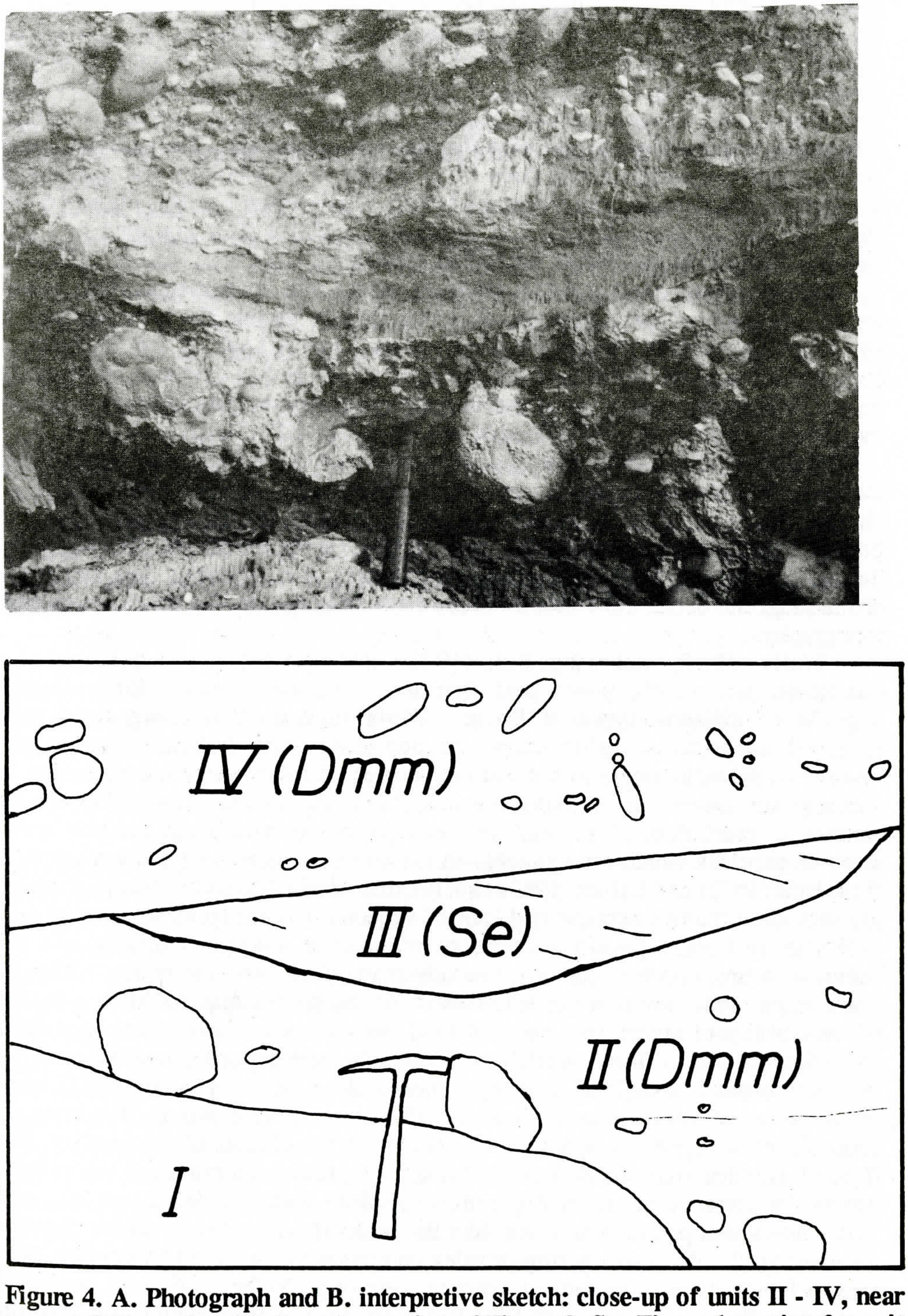


Figure 4. A. Photograph and B. interpretive sketch: close-up of units II - IV, near center of Figure 1 and left-center portion of Figure 2. See Figure 1 caption for unit and facies definitions.

water-laid deposits." Uniformity of thickness, and sheet-like form, which characterize unit IV, are widely but not universally accepted as characteristics of debris flow deposits. Bull (1972) notes that "because the bulk of fan deposits are deposited as sheets and lobes, uniform thickness for a given bed is common in most outcrops, particularly for debris-flow deposits." Similarly, Wasson (1977) finds that debris flow deposits consist of thin, laterally extensive sheets, parallel or subparallel with each other, and with the depositional surface. The channel-lining, matrix-supported conglomerate (unit II) is similar in its lithology and distribution to a debris flow deposit from Wasson's (1977) Windy Point Fan (his Figures 4 and 10). The Windy Point deposit was apparently a debris flow confined to a pre-existing channel on the Windy Point Fan surface.

Gloppen and Steel (1981) interpret clast- and matrix-supported, poorly sorted, unorganized conglomerates as debris flows, "despite the common lack of any large amounts of clay or silt-sized matrix." (p.51) Lack of erosion between beds, a feature shared by the Asbury terrace fan, is cited as evidence favoring the debris flow interpretation. The Asbury terrace conglomerates strongly resemble the subaerial debris flows of Gloppen and Steel's (1981) classification.

Shultz (1984) infers that massive matrix-supported conglomerate (facies Dmm in his classification) is deposited from plastic debris flows, which are characterized by high yield strength, laminar flow, and viscous (rather than grain-grain collision) interactions. Like Wasson (1977), Shultz (1984) also notes that diamictites associated with channeling do not imply that the debris flow caused the channeling; the debris flow might simply have followed and/or filled pre-existing topography.

Wells (1984), following Bull (1972), distinguishes between sheetfloods, mudflows, and debris flows, and discusses diagnostic criteria for evaluating deposits of different types of flows. Wells suggests that unorganized mud-(matrix-) supported conglomerates are deposited from turbulent unchanneled mudflows, whereas unorganized clast-supported conglomerates are deposited by sheet debris flows. It is likely, however, that the mechanism of the support (matrix *vs.* clast) depends not only on the depositional mechanism, but also on the sizes of particles which are available in the source materials of the gravity flow. Both Bull (1972) and Nilsen (1982) observe that debris flows form where (i) slopes are unstable (due to steepness and lack of vegetation), (ii) abundant water is made available over short periods of time at irregular or seasonal intervals, and (iii) sediment sources provide abundant muddy material to form the matrix. This last point in particular stresses the significance of the source material in determining the availablity of matrix-size material (and, presumably, coarse clasts as well) to the sediment gravity flow. Wells' conglomerates were deposited under a cool, wet maritime climate with local perennial snowbanks. Such an environment favors preservation of coarse clastic material, allowing coarse clasts to dominate, for example, clast-supported debris flows. However, the climate of the southern Blue Ridge facilitates vigorous chemical weathering of rocks and minerals; this is borne out by the presence of weathering profiles so thick and well developed that, even under humid-temperate conditions (like the present) which favor intense chemical weathering, the deep weathering profiles on nearby slopes must have taken many tens or hundreds of thousands of years to form (e.g., Velbel, 1985). It is possible that debris flows in the southern Blue Ridge are derived from source materials that do not contain sufficient coarse clasts to form clast-supported debris flow deposits like those discussed by Wells. If so, the conglomerates of the Asbury terrace exposure may be sheet debris flow deposits, despite the fact that they are matrix-, rather that clast-, supported. DeCelles and others (1987), Wells and Harvey (1987), and Blair (1987) have all recently stressed the importance of source materials (especially the availability of fine-grained materials) in determining the mechanisms and styles of alluvial fan sedimentation, particularly the occurrence of debris flows.

The fine sand bed (unit V, Facies Fl & P), with the red lower portion and the white upper portion with a vertical prismatic structure, is interpreted as an arid paleosol developed on a fine-grained fluvial deposit. Similar bleaching of shallow soil horizons *could* result from other processes, such as kaolinization of feldspar, or development of a podzol-like A horizon. However, both of these alternatives are inconsistent with the dehydrated state of the associated iron minerals, which is implied by the color of unit V, facies Fl (discussed below). Silts in beds 20 cm to 1 m thick, interstratified with sheet gravels, occur in stratigraphically similar sequences in Quaternary deposits of Spain (Harvey, 1984); the Spanish silts are partly cemented, reddish, and contain only occasional weak horizontal bedding. Harvey (1984) interprets these sediments to be either flood deposits, or clast-free mudflows. The prismatic structure is a diagnostic property of natric soil horizons (Soil Survey Staff, 1975, p. 28 & Plate 4B). Such structures are commonly associated with soils from arid and semi-arid zones, and in regions with a pronounced dry season (Buringh, 1970; Birkeland, 1984). Because the outcrop no longer exists, and there are no samples, it is not possible to analytically characterize these horizons. However, the fact that the bright red color of the parent fluvial material (facies Fl) is lost in facies P, possibly due to masking by evaporite mineral efflorescences, supports (although it cannot prove) the natric character of the horizon in question.

The reddish color of the parent material (unit V, facies Fl) supports the hypothesis that the white prismatic horizon formed under the influence of an arid, semi-arid, or seasonally desiccated climate. Following Schwertmann (1985, p. 176-179), the reddish color suggests that hematite rather than goethite dominates the iron-oxide mineralogy of unit V. The dominance of hematite, in turn, favors the hypothesis of a relatively warm and/or dry climate during the formation of unit V. Furthermore, the fact that only unit V exhibits red color, whereas the saprolite and all other fine clastic materials are yellow-orange, suggests that unit V was the only unit in the sequence deposited under such warm and/or dry conditions. Unlike other "redbeds" (e.g., Walker, 1967), age of the iron-oxide material is probably not the controlling factor in this example, because units both above and below the red interval are yellow-orange. Residual and transported materials with both red and yellow-orange colors are widely exposed in the immediate vicinity of the Asbury terrace exposure. Therefore, unlike questions of the character and origin of the "natric" horizon, which cannot be tested unless another occurrence is discovered, hypotheses regarding the relationship between color and iron oxide mineralogy, and the relationship between iron oxide mineralogy and controlling factors like climate, can be tested. This work is in progress.

DEPOSITIONAL SYNTHESIS

The sediments exposed in the Asbury terrace exposure were deposited on an alluvial fan. Many of the sedimentary and pedogenic features of these deposits are widely recognized in alluvial fan sequences. Among these features are the unorganized matrix-supported conglomerates (units II and IV), and unit V, the reddish fine sand capped by the whitened interval with vertical prismatic structure. Other features that the Asbury exposure shares with alluvial fans include; proximity to source (the Asbury terrace locality is less than one-half kilometer from the crests of the nearest hills to the west of the Little Tennessee River Valley), major vertical and lateral facies variations, colors characteristic of oxidized sediments, channels, depositional bodies containing soil profiles, and possible salts (criteria of Nilsen, 1982). The channeled, cross-stratified sands (unit III) and the finer sands of units V and VI are amenable to a variety of interpretations, but are consistent with alluvial fan deposition.

The dominance of debris-flow deposits in the lower part of the section (units II-IV) suggests a proximal-fan origin for these units; Bull (1972) states that "the proportion of debris flow deposits decreases downfan from the apex in those fans where both water-laid and debris-flow deposits are present." (p. 81). According to Gloppen and Steel (1981), mass-flows indicate deposition on proximal portions of the fan. Similarly, Shultz (1984) finds that his facies Dmm conglomerates are common in proximal-fan settings. The presence of cut-and-fill structures (unit III) also suggests a proximal origin for the lower part of the section (Bull, 1972). The red fine sand layer with its paleosol (unit V) cannot be easily placed into a proximal-distal framework, but its lithology, color, and paleosol features are again consistent with the alluvial fan model.

DISCUSSION

Most studies of alluvial fans and their deposits are based on examples from tectonically active, arid geologic settings (e.g., western North America; Bull, 1972; Nilsen, 1982; Rust and Koster, 1984). Recently, however, alluvial fans and their deposits have been discovered and studied in other geologic and/or climatic settings, including Quaternary and ancient cold-humid and periglacial settings (e.g., Wasson, 1977; Harvey, 1984; Wells, 1984; Wells and Harvey, 1987; Blair, 1987). Other recent studies have found, described, and interpreted alluvial fan deposits in the tectonically stable, humid-temperate setting of the southern Appalachians (e.g., Gryta and Bartholomew, 1977).

The deposits of the Asbury terrace deposit, and the sedimentological interpretation of those deposits, are similar to other alluvial fan deposits found elsewhere in the southern Appalachians. For example, Mills (1981, 1982a,b, 1983) reported numerous examples in the Great Smoky Mountains of fan-like features consisting largely of matrix-supported diamicton, which he interprets to be debris flows. The debris flows are thought by Mills to have originated *via* relatively recent (post-glacial?) high-magnitude episodic (catastrophic) erosion/sedimentation events. Acknowledging that definitive evidence is lacking, Mills suggests that episodic erosion/sedimentation in the Great Smoky Mountains may have been caused by climate change associated with Holocene post-glacial climatic evolution, although physiographic and autocyclic factors may play an even more significant role.

Kochel and Johnson (1984) recognized that the tectonic and climatic setting from which the generic alluvial fan facies models were developed are not universal. They defined "The Virginia Humid-Temperate Alluvial Fan Model" based on their studies of fan deposits in Nelson County, Virginia. These authors found that humid-temperate fans experience periods of depositional activity at intervals on the order of thousands of years. Many fan surfaces are presently forested, which suggests an absence of recent depositional activity. Some portions of fan surfaces, however, were reactivated by sedimentation and transport caused by the intense precipitation which accompanied Hurricane Camille in 1969. Nelson County fan sequences are less than 110 m thick; most are about 20 m thick. Even the most resistant rock types making up conglomerate clasts are subangular to rounded. Thin (<2 m thick) matrix-supported boulder facies are most common in proximal portions of the fans; Kochel and Johnson (1984) interpret these to be debris flow deposits; in fact, they conclude (p. 119) that "the major depositional process on the Nelson County alluvial fans appears to be debris flow and debris avalanches triggered by intense rainfall." Channel-fills consisting of cross-bedded sand also occur locally. The Nelson County alluvial fan deposits are lithologically, sedimentologically, and stratigraphically similar to the Asbury terrace deposit.

Kochel and Johnson (1984) dated several of their depositional units by radiocarbon, and found three discrete episodes of high-magnitude sedimentation in less than four meters of sediment, representing approximately ten thousand years. On the basis of their radiocarbon data, Kochel and Johnson (1984) estimate a recurrence interval for fan sedimentation events on the order of three to six thousand years. Deposits of the Camille flood (1969) are clearly distinguishable from older deposits by their unweathered nature. Discrete sediment units within individual stratigraphic columns of the Nelson County deposits differ from one another in their weathering characteristics, a finding similar to that of Mills (1981, 1982b) in the Great Smoky Mountains. Kochel and Johnson's (1984) radiocarbon analysis supports Mills' (1982b) inferences that debris-flow dominated alluvial fan deposits of the southern Appalachians originated *via* relatively recent (post-glacial) high-magnitude episodic erosion/sedimentation events, possibly influenced by climate change associated with Holocene post-glacial climatic evolution.

The similarities between the fan deposits discussed above and the Asbury terrace deposit are striking. All share evidence of brief episodic sedimentation at long, irregular time intervals, although only Kochel and Johnson's (1984) study permits quantification of the time intervals between sedimentation events.

The Asbury terrace study illustrates the need for further careful and multi-faceted studies of alluvial fan/terrace deposits of the southern Appalachians. The fans of the southern Appalachians are not tectonically controlled like those in more active tectonic settings (e.g., Decelles and others, 1987; Nichols, 1987). Instead, climate is a major controlling factor. Even in other climatic and tectonic settings "Many thin fan deposits may result from minor climatic changes and represent brief depositional intervals during long periods of erosion" (Nilsen, 1982, p. 54). Other workers (e.g., Mills, 1981, 1982a,b, 1983; Kochel and Johnson, 1984) in the southern Appalachians have noted the importance of debris flows in the geomorphic evolution of the region, and the possible role of climate in driving the erosive/sedimentational episodes. Debris avalanching is increasingly recognized as an important geomorphic process in the southern Appalachians (Grant, 1983; Velbel, 1985; Neary and others, 1986).

Recent work has suggested, however, that even climate may not play the fundamental role in determining the style of alluvial fan sedimentation. Mills (1983) has suggested that intrinsic geomorphic thresholds within fan settings have a greater influence than climatic triggering on sedimentation. Wells and Harvey (1987, p. 197) note that "the humid fans of the Howgill Fells (their study area) display characteristics which occur in arid-land fans and which have been attributed to climatic fluctuations during the Quaternary." Like Blair (1987), they suggest that this similarity in sedimentary deposits reflects *primarily* the similarity of response to intrinsic geomorphic thresholds, related to physiographic properties of source areas and fan surfaces. Climatic fluctuations exert only secondary comtrols over the major sedimentary facies on any individual fan. The work of Mills (1983), Wells and Harvey (1987) and Blair (1987) suggests that source and fan physiography and lithology, and autocyclic (intrinsic) factors, can combine even in humid settings to give sedimentary facies assemblages which are usually attributed to arid fan processes. Their work suggests that such factors may be as important as climate in controlling the nature of alluvial fan sedimentation.

To say that facies architecture may not be as strongly dependent on climate as was previously thought does not, however, mean that these sediments are poor recorders of climate. The stratigraphically controlled color changes in the Asbury terrace deposit strongly suggest that the hydration state of the pigment-causing iron oxide minerals was determined approximately at the time of deposition, and has not been modified by later diagenesis of the deposits. This clearly demonstrates that color does *not* uniformly redden with age, as is widely assumed in relative age dating of surficial deposits. It also suggests that the paleosol features, and the chemistry and mineralogy of the sediment pigments, may provide a record of Holocene climatic and geomorphic history in the southern Appalachians. (Framework components of terrace deposits may also contain such information; see Grantham and Velbel (in press) for a discussion of the climatic and source-area geomorphic influences on composition of modern sands in the area.) The stratabound distribution of iron oxide/oxyhydroxide pigment in the Asbury terrace sediments indicates that unit V was affected by greater dehydration than any of the other units. Color is a useful index of the hydration state of iron oxide/oxyhydroxides in soils; reddish colors indicate the dehydrated form, hematite, which in turn indicates drier conditions (Schwertmann, 1985). The fact that subjacent and superjacent units were not similarly subjected to dehydration suggests that the time interval between the deposition of units V and VI represents a temporary but significant excursion to drier climatic conditions. The yellow-orange color of the bulk of the sediments suggests that most of the deposition on the Asbury terrace took place under conditions of temperature and humidity more like those of the present (conditions under which yellow-orange goethite weathering products are forming; Velbel, 1984, 1985) rather than under warmer or drier conditions. Irrespective of the primary control, alluvial fan terrace deposits may therefore provide an important source of information on magnitude/frequency properties, geomorphic history, and possibly the climatic history, of the southern

Blue Ridge landscape and its episodic sediment gravity flows.

The Asbury terrace deposit thus illustrates another important point in the application of alluvial fan facies models. Rust and Koster (1984) are correct in pointing out that "Red colouration and evaporitic paleosols occur in several ancient alluvial fan deposits, and point to a semi-arid paleoclimate" (p. 59). However, as discussed above, it can take many thousands of years to deposit a few meters of sediment (Kochel and Johnson, 1984). This interval of time almost certainly encompasses climatic changes, possibly significant ones. It is therefore dangerous to attribute a specific paleoclimate to an entire stratigraphic interval of alluvial fan deposits, because climate may vary significantly on time scales approaching the mean recurrence interval of sedimentation episodes on the fan.

Although the Asbury terrace deposit no longer exists, there are several other known examples of terrace deposits which contain at least some of the features displayed by the Asbury exposure, and which may be correlative with it and with one another (Velbel and Grantham, 1985). If the Asbury deposit is any indication, other alluvial terrace deposits in the southern Blue Ridge may contain valuable and retrievable information about the climatic and geomorphic history of the southern Blue Ridge Mountains.

CONCLUSIONS

The Asbury terrace deposit consisted of massive matrix-supported conglomerate, weakly cross-stratified scour-filling sand, and several homogenous or poorly laminated fine sands, including a distinctive red interval with white color and prismatic structure at the top. Conglomerates are interpreted to be products of debris flows; both sheet-like forms with uniform thickness, and deposits which fill pre-existing channels occur. The red and white interval capped by prismatic structure is interpreted to be an arid paleosol developed on a fine-grained fluvial deposit. Color differences among different strata are of primary, rather than diagenetic, origin, and are probably caused by differences in the relative abundance of goethite and hematite. Similar color and mineralogical differences in other surviving outcrops may therefore prove useful in elucidating aspects of paleoclimate in the region. The channeled and cross-stratified sand is of fluvial origin.

The sedimentary facies of the Asbury terrace deposit indicate deposition on an alluvial fan, under variable climatic conditions. Debris flows, the yellow-orange (goethitic) color of most units, and the young age and regional paleoclimatic setting suggest that the bulk of the sediments were deposited under humid-temperate conditions much like the present. One sedimentary unit (unit V), however, indicates arid conditions by its red (hematitic?) pigment and the natric paleosol features. The existence of alluvial fan deposits in this instance is more strongly influenced by climate rather than by fault-block or thrust-fault tectonism, but no one climate characterizes the deposition of the entire fan sequence, apparently because climatic variations took place on a time-scale approaching the mean recurrence interval of sedimentation episodes on the fan.

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SUBSURFACE STRATIGRAPHY OF THE UPPER DEVONIAN AND LOWER MISSISSIPPIAN OF NORTHERN WEST VIRGINIA

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ABSTRACT

Formal stratigraphic nomenclature for the Upper Devonian-Lower Mississippian clastics in the outcrop areas of the Appalachian Basin is well established. However, the subsurface stratigraphy has not been formally described in the central Appalachians and confusion exists concerning the correct correlation with surface strata. This report combines detailed subsurface correlations based upon gamma-ray well logs with outcrop study, providing a better understanding of the subsurface stratigraphy of the "Catskill" wedge across northern West Virginia. This study documents the changing position of rock unit boundaries, the areal extent of stratigraphic units, and the facies relationships of units traced east from the subsurface to outcrops located at Rowlesburg, Hannahsville, and Elkins, West Virginia.

Certain revisions and extensions of the present stratigraphy are proposed. 1) The Greenland Gap Group is reduced to formation rank west of the Allegheny Front, replacing the term "Chemung". 2) The Hampshire Formation is elevated to Group, comprised of two new formations, the Cannon Hill and the Rowlesburg. 3) The Cussewago Member of the Price Formation is not recognized in the northern West Virginia subsurface. 4) The Riddlesburg Shale Member of the Price Formation is redefined as the Riddlesburg Member. 5) The Venango, Oswayo, and Riceville formations are extended southward from Pennsylvania and recognized as subsurface units. 6) The Cuyahoga Formation is extended into western West Virginia from Ohio. 7) The Greenland Gap, Cannon Hill and Rowlesburg formations are subdivided into subsurface intervals which correspond closely to the traditional driller's intervals.

INTRODUCTION

In recent years, the Upper Devonian-Lower Mississippian stratigraphy of West Virginia has become better understood as a result of the extensive study of these units in outcrops along the eastern margin of the Allegheny Plateau (Dennison, 1970; McGhee and Dennison, 1976; Kammer and Bjerstedt, 1986). These studies recognize the ambiguity of the earlier stratigraphic terminology such as "Pocono" and "Chemung", and establish more accurate and detailed nomenclature. This new knowledge is herein applied to the subsurface so that an accurate stratigraphic framework can be established for future detailed studies and for basin analysis.

East-west facies changes recognized in the subsurface of West Virginia preclude the extension of outcropping units across the state. This paper describes and correlates these facies changes in the subsurface and proposes a more detailed stratigraphic terminology for the subsurface rocks in northern West Virginia.

The specific study area includes more than 10,000 square kilometers (roughly 4000 square miles) of northern West Virginia, extending from the eastern outcrop belt to the Ohio River (Figure 1). The area encompasses the pinchout of the sandstone-rich part of the "Catskill" clastic wedge. The stratigraphic interval of interest extends from the top of the Pound Member of the Greenland Gap Formation, vertically through approximately 900 meters (3000 feet) of section to the base of the Greenbrier Limestone. Stratigraphic relationships were recognized following the correlation of stratigraphic units recognized in twelve regional subsurface cross-sections and three measured sections from outcrops near Rowlesburg, Elkins, and Hannahsville, West Virginia. Over 400 gamma-ray well-logs were analyzed and correlated in this study (Figure 1).

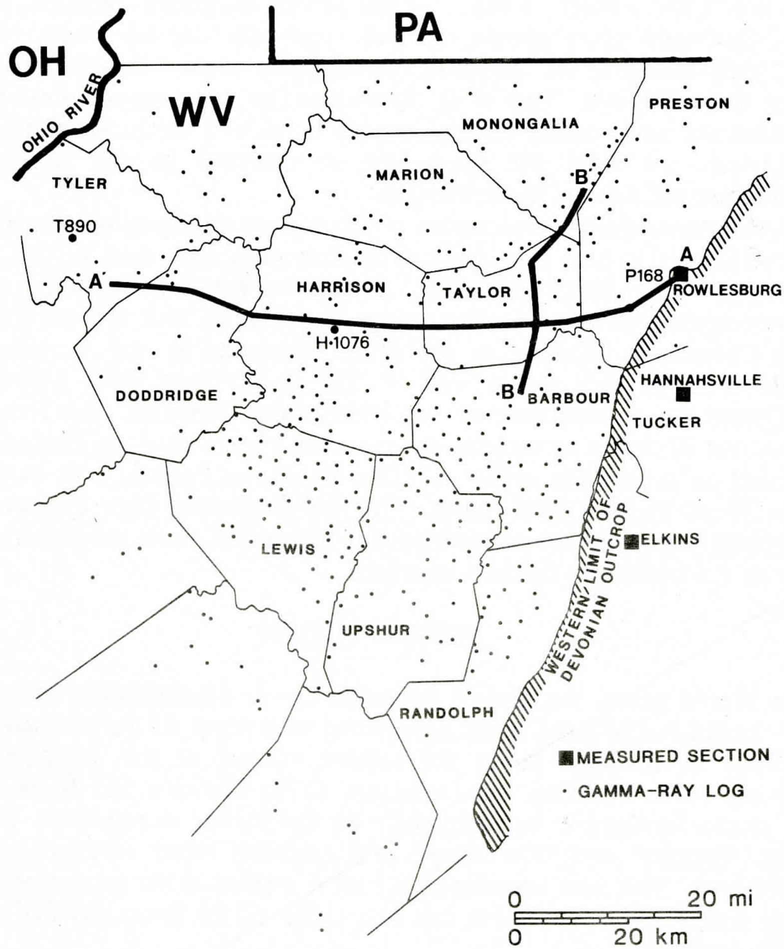

Figure 1. Location of the study area. Line A-A' marks the location of the gamma-ray cross-sections shown in Figures 4 and 5.

The criteria for recognizing units in the subsurface are: precedent, lithologic homogeneity, substantial lithologic contrast with bounding units, correlatability, and mappability. These guidelines led to four necessary changes in the stratigraphic nomenclature. 1) The lithologic contrast between the Foreknobs and Scherr Formations was not sufficient to permit their recognition in the subsurface. Consequently, the Greenland Gap Group is reduced to the rank of formation west of the Allegheny Front, where it replaces "Chemung". 2) The Hampshire Formation is herein elevated to the Hampshire Group. The "Upper Hampshire" of Lewis (1983) is formally designated the Rowlesburg Formation, and is described from an outcrop dominated by red shales and channel-filling sandstones located 1.9 kilometers (1.2 miles) west of Rowlesburg, Preston County, West Virginia (a complete description and location of the Rowlesburg type section is presented in the appendix). A reference section is located 19 kilometers (12 miles) to the south, along route 72 south of Hannahsville in Tucker County. The "Lower Member" of Lewis (1983) is named the Cannon Hill Formation for an exposure of "clean" tabular sandstones and minor red shales at Rowlesburg. A reference section for the Cannon Hill Formation is located along Route 33 east of Elkins, Randolph County. 3) The Cussewago Member of the Price Formation (Kammer and Bjerstedt, 1986) is poorly developed and cannot be confidently identified in West Virginia. 4) The Riddlesburg Shale Member of the Price Formation (Kammer and Bjerstedt, 1986) is found to contain significant sandstone in the study area, and is therefore redefined as the Riddlesburg Member.

The Greenland Gap, Cannon Hill, and Rowlesburg Formations are subdivided into intervals that are particularly useful in subsurface studies. These intervals correspond to the well-known driller's-named intervals and basically are chronostratigraphic units. Subsurface correlation indicates that these divisions were not arbitrarily made by drillers, rather, they mark distinct sandstone packages that were segregated from the larger clastic wedge by cycles of transgression and regression of lesser magnitude than the one that produced the Hampshire Group. The westward time-transgressive nature of the lithostratigraphic formations results in some intervals being recognized in more than one formation. For example, the "Fifth" sandstone interval is contained within the Rowlesburg Formation at Rowlesburg, within the Cannon Hill Formation in the subsurface of Taylor County, and within the Greenland Gap Formation in Harrison County and westward.

CURRENT NOMENCLATURE

The current stratigraphic nomenclature for the West Virginia subsurface and surrounding outcrops is shown in Figure 2. This chart combines the data presented on the COSUNA correlation chart published by the AAPG (Patchen and others, 1985) with recent modifications proposed by Kammer and Bjerstedt (1986). In general, there is more detailed information in the published literature on the correlation of Frasnian and Lower Mississippian rocks than exists for rocks of the Famennian. One reason is that, along the Allegheny Front, much of the Famennian is represented by the Hampshire Formation; a unit that is difficult to subdivide because of its lenticularity and relative lack of fossils. Similarly, there are no Riceville or Venango lithologies exposed in the outcrops of West Virginia.

Without regional subsurface study, the understanding of the stratigraphy of these highly diachronous units has not advanced.

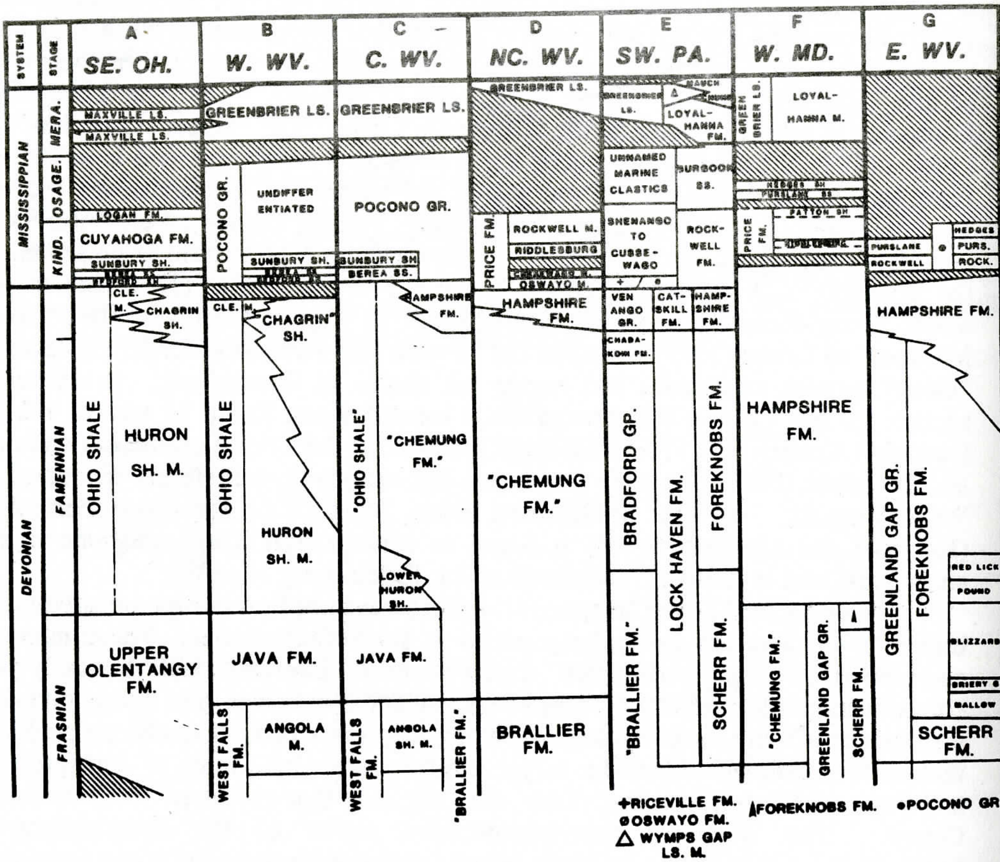

Figure 2. Correlation of Upper Devonian and Lower Mississippian stratigraphic units. This chart is based on the COSUNA correlations (Patchen and others, 1985) with the modifications of Kammer and Bjerstedt (1986). The areas to which the columns refer is shown in the insert of Figure 3.

PROPOSED NOMENCLATURE

The proposed system of nomenclature for the subsurface of West Virginia is illustrated in Figure 3. The interpreted stratigraphy across northern West Virginia is shown in two abbreviated gamma-ray cross-sections. The correlation of subsurface units from the "Warren" interval to the Greenbrier Limestone with the rocks exposed at Rowlesburg is given in Figure 4. The correlation of lower units is shown in Figure 5.

The proposed nomenclature shown in Figure 3 gives a more accurate portrayal of the position of diachronous unit boundaries across the state. A systematic description of each unit as it occurs in the subsurface is presented. Each unit will be referred to one of the three typical gamma-ray logs (Figures 6, 7 and 8) in which it is best represented.

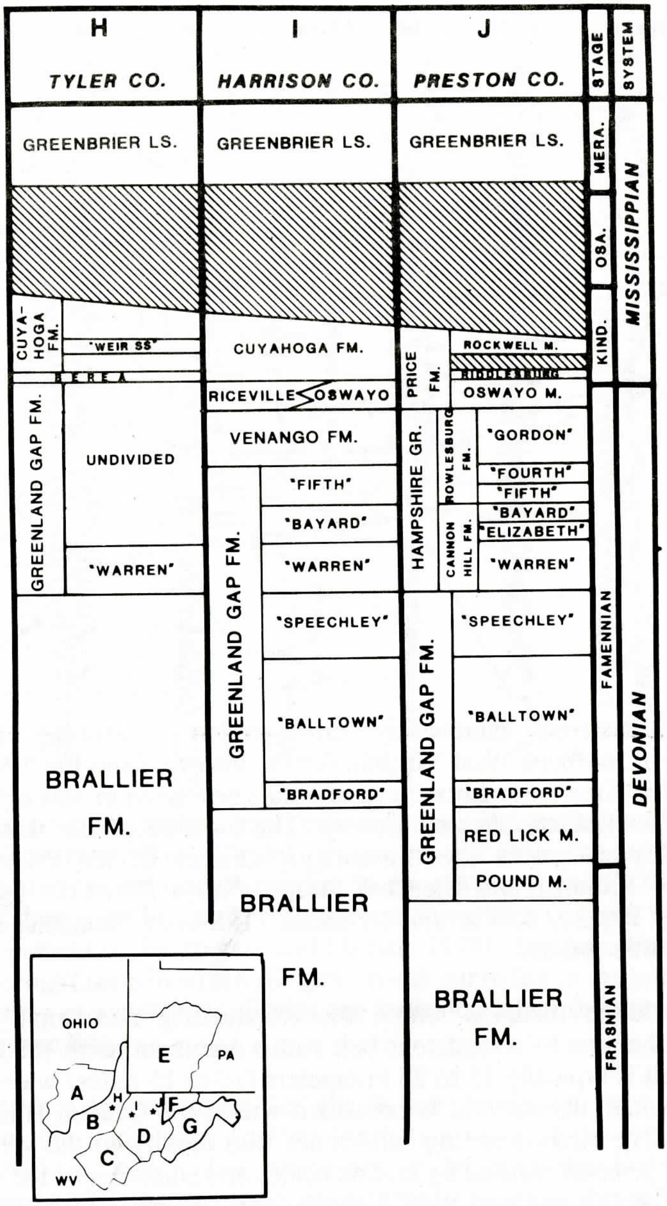

Figure 3. Correlation chart for northern West Virginia based upon this study.

Isopach maps of a Late Devonian sandstones from the study area typically show a tripartite division. To the east is a shale-rich zone containing minor east-west trending sandstones. Lithologic data for this part of the interval indicate that the shales are red. In outcrop, the fine-grained units within this facies are commonly red mudstones containing features indicative of flood plain deposition, including root mottling and concave-upwards, occasionally calcitic fractures resulting from shrinking and swelling known as pseudoanticlines. Here, an

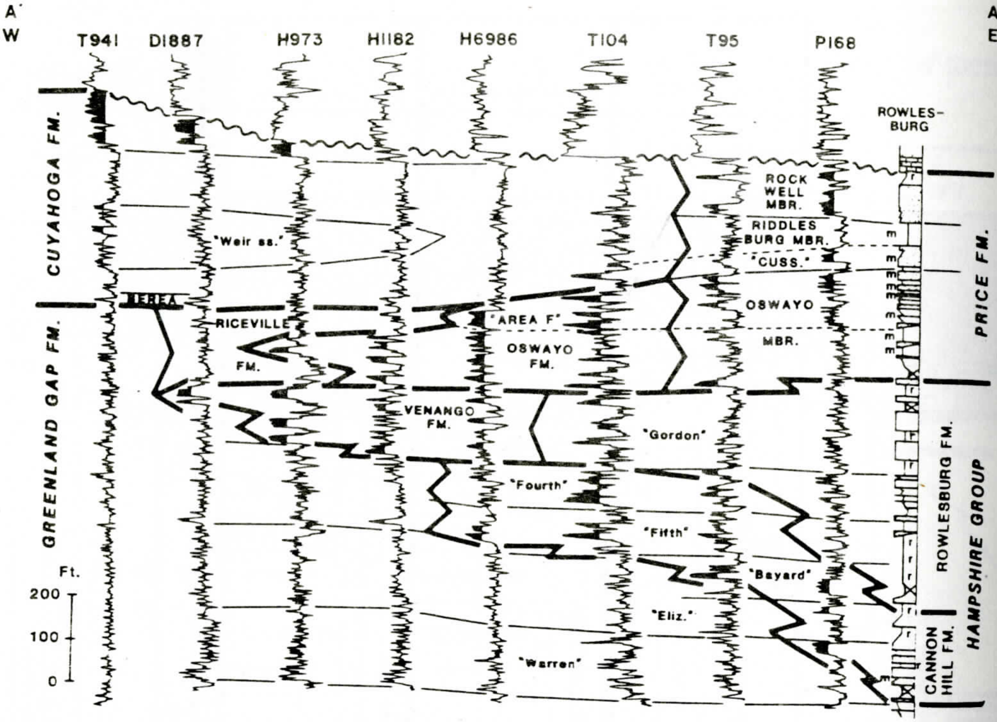

Figure 4. East-west gamma-ray cross-section illustrating the subsurface stratigraphy of northern West Virginia for the interval from the "Warren shale" to the base of the Greenbrier Limestone, and the correlation of subsurface units to the outcrop at Rowlesburg, Preston County. The location of the section is given in Figure 1. "Cuss." marks the Cussewago Member of the Price Formation as described by Kammer and Bjerstedt (1986). "Area F" marks the "area F fan" described by Pepper, deWitt and Demarest (1954). "r" indicates red color; "m" indicates marine fossils.

interval would be included within the Rowlesburg Formation. This area is bounded on the west by a sandstone belt with a dominant north-south (strike) trend. This sand belt is typically 15 to 25 kilometers (10 to 15 miles) wide and represents the area in which the interval is contained within the Cannon Hill or "Venango" Formation. The strike-trending sandstones thin rapidly to the west, where their stratigraphic position is filled by marine shales and siltstones of the Greenland Gap Formation. Within northern West Virginia, the "Warren", "Elizabeth", "Bayard", "Fifth" and "Fourth" sandstone intervals contain all three of these facies.

Greenland Gap Formation

The name Greenland Gap Formation (Greenland Gap Group of the Allegheny Front; comprised of the Scherr and Foreknobs Formations, Dennison 1970) is herein applied in the subsurface to the thick marine shale, siltstone and sandstone lithosome heretofore known commonly as the "Chemung". Dennison anticipated the lowering in rank of the Greenland Gap Group to formation status in the West

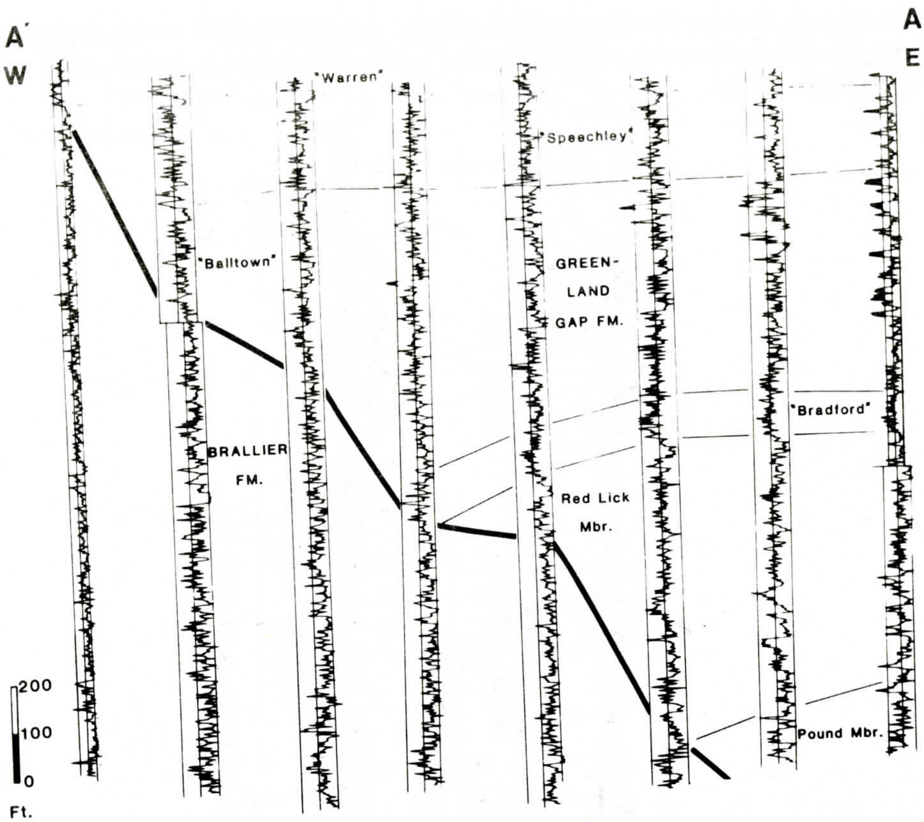

Figure 5. East-west gamma ray cross-section illustrating the subsurface stratigraphy of northern West Virginia in the interval from the top of the Pound Member of the Greenland Gap Formation to the "Warren shale". The location of the section is shown on Figure 1. The vertical lines bisecting the logs are "base lines" (in the manner of Piotrowski and Harper, 1979). The base lines are used to determine lithology; units in which the gamma-ray curve fall to the left of the center line are siltstones or sandstones. Shales are indicated when the curve falls to the right of the center line.

Virginia subsurface by indicating that the Scherr Formation may not be distinguishable from the underlying Brallier Formation west of the Allegheny Front (Dennison, 1970, p. 63). The ambiguity in locating the Foreknobs-Scherr boundary suggests that the safest course is simply to apply the name Greenland Gap Formation for those units that may be either Foreknobs or Scherr west of the Allegheny Front.

The transition from the Greenland Gap Formation to the overlying Venango and Cannon Hill Formations is marked in the subsurface at the base of the lowest, "clean", massive sandstone that is in association with a thick sequence of coarse-grained sandstones and red shales. This contact is distinct in northcentral West Virginia, where it occurs at the base of the Venango Formation ("Gordon" sandstone)(Figure 4 and 9a). Westward from the pinchout of the Venango, the contact is placed at the base of the Berea Sandstone. Eastward, the contact stratigraphically lowers; first to the base of the "Fifth" sandstone, then to the base of the "Bayard", and to the base of the "Warren" sandstone adjacent to the outcrop belt (Figure 4).

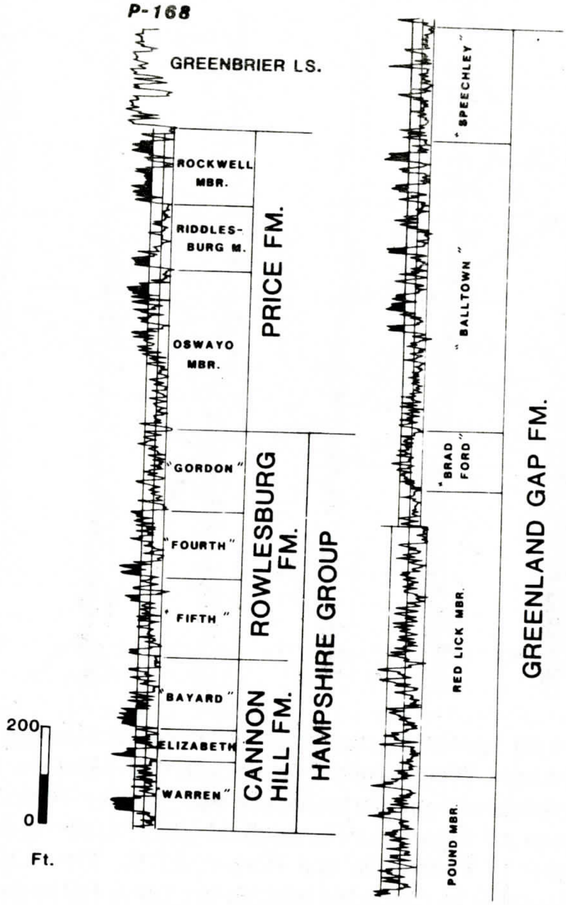

Figure 6. Gamma-ray well log Preston-168. This log is typical of logs from the eastern part of the study area.

The Greenland Gap Formation is distinguished from the underlying Brallier Formation by being slightly coarser-grained. In the subsurface, this transition is marked where the gamma-ray log signature passes from primarily 25 to 50 percent sandstone and siltstone (Greenland Gap) to less than 25 percent sandstone and siltstone (Brallier). This contact is highly time-transgressive, rising from below the Pound Member in the subsurface of eastern West Virginia, to the base of the "Warren" interval in the west, a stratigraphic thickness of approximately 420 meters (1300 feet) (Figures 5 and 9b). As a result, the Greenland Gap as recognized in eastern West Virginia (from the base of the Pound Member to the base of the "Warren"), is older than the Greenland Gap recognized by lithology in northwestern West Virginia (base of the "Warren" to the Berea Sandstone). The stratigraphic rising of the base of the Greenland Gap to the west necessitates the recognition of the Brallier Formation in units much younger than has previously been suggested. This Famennian part of the Brallier is equivalent in age to the Huron Shale in eastern Ohio, however, because only minor black shales occur within the Famennian section of northern West Virginia, these units are not assigned to the Huron.

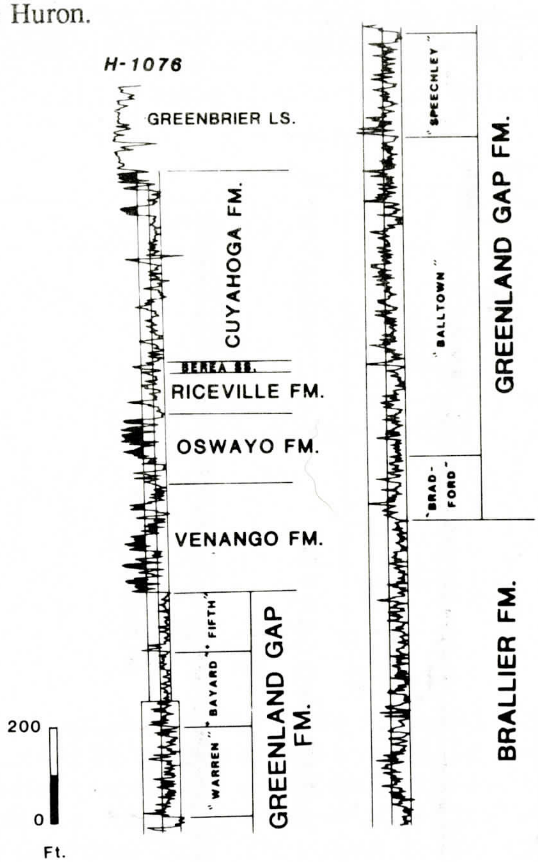

Figure 7. Gamma-ray well log Harrison-1076. This log is typical of logs from northcentral West Virginia and serves as a reference log for the Oswayo and Venango Formations in West Virginia.

In the subsurface, the Greenland Gap Formation is subdivided into a series of informal units. These intervals were utilized in subsurface studies of the Upper Devonian within central West Virginia by Cardwell (1981, 1982). Although lithologically similar to one another, they are bounded by sub-regional shale marker beds and contain distinguishing characteristics that support their recognition as separate stratigraphic entities. The uppermost members of the Foreknobs Formation, the Pound (Dennison, 1970) and the overlying Red Lick (McGhee and Dennison, 1976), are recognized in the subsurface and are retained as formal members of the Greenland Gap Formation.

Pound Member: The Pound Member is predominantly sandstone in outcrops along the Allegheny Front where it is recognized as a member of the Foreknobs Formation of the Greenland Gap Group (Dennison, 1970). The Pound has been identified at the Elkins outcrop by Lewis (1983), and Barrell and Dennison (1986).

The upper part of the Pound Member has been correlated with the well-known "Benson" sandstone of the subsurface (Lewis, 1983). The occurrence of the "Benson" sandstone allows the recognition of the top of the Pound Member throughout much of the study area, where it contrasts with the shalier overlying Red Lick Member. The top of the Pound Member occurs approximately 500 to

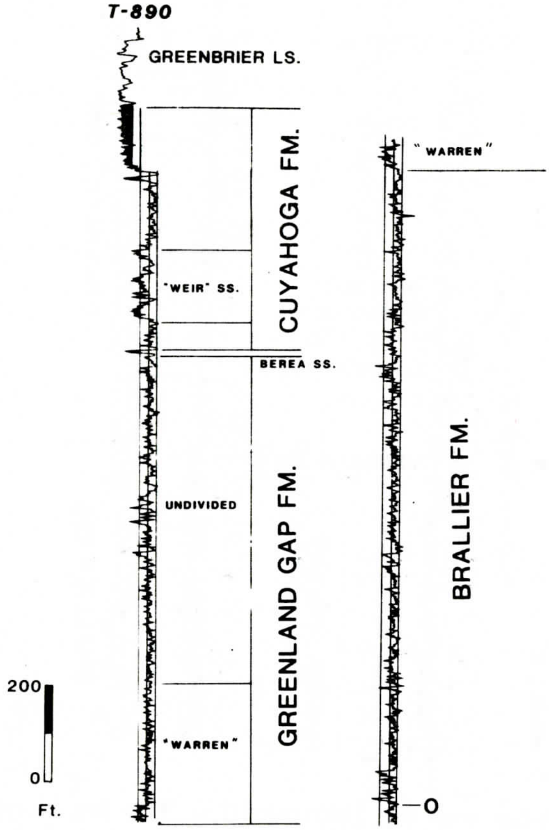

Figure 8. Gamma-ray well log Tyler-890. This log is typical of logs from the western part of the study area. "O" marks the horizon equivalent to the top of the Olentangy as recognized in Ohio (Majchszak, 1980).

700 feet below the base of the "Bradford" siltstone and encloses the Frasnian-Famennian boundary (McGhee and Dennison, 1980). This contact marks the base of the specific stratigraphic interval described in this report.

Red Lick Member: The Red Lick was described as a member of the Foreknobs Formation by McGhee and Dennison (1976). The unit occupies the interval between two easily identifiable marker beds, the subjacent Pound Member, and the overlying "Bradford" siltstone (Figure 6). The Red Lick Member is known in the subsurface as the "Riley", and is a generally finer-grained unit than either the Pound or "Bradford". A westward thinning siltstone wedge occurs in the middle of the member and extends roughly 25 kilometers (15 miles) into the subsurface. Where this wedge thins and siltstone percentage within the "Riley" drops below 40 percent, the unit is included within the Brallier Formation (Figures

5 and 9b). Well logs of the Red Lick Member are characterized across the extent of the unit by the occurrence of several well-developed vertically coarsening cycles.

The "Bradford" Interval: The "Bradford" (Figures 6 and 7) forms a distinctive marker bed throughout the eastern half of the study area. The unit is tabular, approximately 30 meters (100 feet) thick, and siltstone content is usually

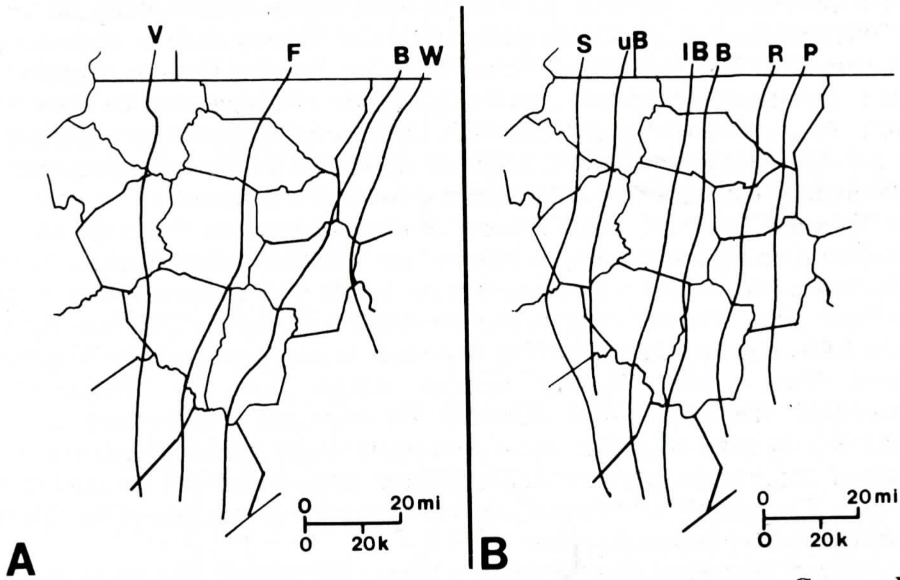

Figure 9. A) Stratigraphic position of the Greenland Gap-Venango-Cannon Hill contact. As traced to the east, the contact lowers to the base of the following units: West of V = Berea Sandstone; V = Venango Formation ("Gordon sandstone"); F = "Fifth" interval; B = "Bayard" interval; W = "Warren" interval. B) Stratigraphic position of the Greenland Gap-Brallier contact. As traced east, the contact occurs at the base of the following units; west of S = "Warren" interval; S = "Speechley" interval; uB = upper "Balltown" interval; 1B = lower "Balltown" interval; B = "Bradford" interval; R = Red Lick Member; P = Pound Member.

70 percent or greater. Where well-developed, The "Bradford" has a characteristic log signature of coarsening upwards in the lower half, and fining-upwards in the upper half, producing a nearly symmetrical log signature. Several thin sandstone "spikes" are common in the middle of the unit. As the unit fines farther to the west, it loses its characteristic signature, ultimately grading into Brallier lithologies approximately 30 kilometers (25 miles) west of the outcrop belt.

"Balltown" Interval: The "Balltown" interval (Figures 6 and 7) extends from the top of the "Bradford" siltstone bundle to a shale marker bed at the base of the "Speechley" interval. This unit is approximately 155 meters (500 feet) thick in Preston and Barbour Counties, West Virginia, and expands to a maximum of 230 meters (750 feet) in the northcentral part of the state. The "Balltown" is characterized by the occurrence of lenticular, north-south trending belts of siltstone in Marion and eastern Harrison counties, high sandstone content in the eastern part of the study area, and the presence of minor amounts of red shale beds in outcrop.

The red shales probably represent the edge of a tongue of Hampshire lithology, but are thin enough (approximately 3 meters (10 feet)) at the Elkins outcrop, that the interval is included within the Greenland Gap Formation (Lewis, 1983).

"Speechley" Interval: The "Speechley" interval (Figures 6 and 7) is, in general, more shale-rich than the intervals that bound it. Throughout the northcentral West Virginia subsurface, the "Speechley" is delineated by two marker shale units. The top of the interval occurs at the shale marking the base of the "Warren" interval, often informally called the "Warren shale". Approximately 60 meters (200 feet) below the "Warren shale" is a second shale, recognized by a similar gamma-ray signature, commonly known to drillers as the "Speechley shale". As in the underlying "Balltown", the "Speechley" contains a siltstone belt that parallels depositional strike; however, unlike the "Balltown", the eastern half of the interval contains very little sandstone and no red shales.

"Warren" Interval: The "Warren" interval (Figures 6, 7 and 8) marks the initiation of the final major progradation of the "Catskill" deltaic wedge. The base of the unit is marked by a massive siltstone bundle that is easily traced westward into Ohio. The "Warren" interval is approximately 60 meters (200 feet) thick and marks the base of the Greenland Gap Formation in northwestern West Virginia and eastern Ohio. The "Warren" interval differs from the "Balltown" and "Speechley" intervals in that siltstones, although still concentrated in a belt paralleling depositional strike, occur over much wider areas with more numerous siltstones bridging the gap between the offshore siltstone belt and the sandstones to the east. The top of the "Warren" interval is marked at the base of the "Bayard" sandstone and equivalent horizons.

"Upper Greenland Gap" Interval: Above the Bayard, and up to the Berea Sandstone, the Greenland Gap Formation is a heterogeneous sequence of interbedded shales, siltstones and minor sandstones (Figure 8). This interval is comprised of near-shore marine sediments intermediate between the finer-grained deposits of the Chagrin Shale to the west, and the coarser-grained deltaic deposits in the interval between the "Bayard" and "Cussewago" sandstones to the east. The boundary between equivalents of the Cannon Hill and Venango Formations is roughly marked by a coarse-grained layer equivalent to the upper "Fourth" sandstone. This bed can be traced widely across western West Virginia and eastern Ohio ("Gordon" of Filer, 1985). A second conspicuous bed, the "Gordon Stray" sandstone, separates Venango equivalents from the overlying Riceville-aged deposits. The top of the "Upper Greenland Gap" interval is marked by Berea-equivalent sheet sandstones.

Hampshire Group

The Hampshire Formation is recognized as the red portion of the "Catskill" wedge (Darton, 1892). However, definition of the boundaries of the unit, particularly the lower boundary, is not clear. Many workers place the base of the Hampshire at the youngest bed that contains marine fossils. Others include more section within the Hampshire by marking the boundary at the stratigraphically lowest occurrence of red color. Lewis (1983) recommended subdivision of the Hampshire Formation into two units, which he informally named the "Upper Hampshire" and the "Lower Hampshire". It is proposed in this paper to elevate these units to formational status; the upper non-marine, red mudstone-rich unit to be named the Rowlesburg Formation, and the lower unit, commonly bearing marine or brackish fauna in sandstone-rich strata with subordinant beds of red shale to be the Cannon Hill Formation. These two formations together comprise the Hampshire Group. Typical gamma-ray well log expression of the Hampshire Group is shown in Figure 6.

Cannon Hill Formation: The Cannon Hill Formation is named and described from the Rowlesburg outcrop (Figure 10 and appendix). An extremely well-exposed section along Route 33, east of Elkins, West Virginia (appendix) has been included as a reference section. The Elkins outcrop has been described previously by Lewis, (1983) and McColloch and Schweietering, (1985). Cannon Hill lithologies are typically tabular, flat-bottomed, quartz-rich sandstones with interbedded red, green and gray shales. This unit becomes younger to the west, as a result of the overall westward shift of the shoreline during Late Devonian time. The stratigraphic position of the base of the Cannon Hill Formation across the study area is given in Figure 9a.

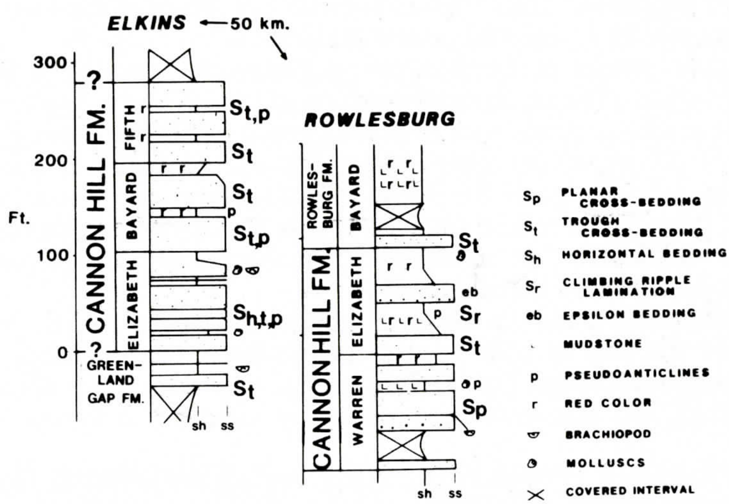

Figure 10. Measured sections of the Cannon Hill Formation along Rt. 33 east of Elkins, Randolph County (reference section), and at Rowlesburg, Preston County (type section). The locations of the outcrops are given in Figure 14.

The Cannon Hill Formation is subdivided into a number of intervals closely corresponding to subsurface drillers terminology. Because of the time-transgressive nature of the Cannon Hill Formation, these intervals persist into the Rowlesburg Formation as traced to the east and the Greenland Gap Formation to the west. However, these intervals are defined, and most clearly recognized, by the position of their sandstone facies recorded in the Cannon Hill Formation.

Rowlesburg Formation: The Rowlesburg Formation is described and named from an exposure near the town of Rowlesburg, Preston County (appendix). Due to the relative inaccessibility of the Rowlesburg outcrop, a well-exposed, albeit incomplete section located on Route 72, midway between Hannahsville and St. George, is used as a reference section (appendix). In the subsurface, the

Rowlesburg Formation consists of a sandstone-poor, red shale and mudstone-rich interval. The Rowlesburg grades upwards into gray, marine-fossiliferous siltstones of the Oswayo Member of the Price Formation. The lower contact of the Rowlesburg Formation is placed at the base of the red shale sequence. This contact is arbitrarily marked where sandstone abundance decreases below 50 percent. At Rowlesburg, the lower contact is located at the base of the "Fifth" sandstone interval. Forty miles to the west, the contact rises 60 meters (180 feet) stratigraphically through the "Fifth" and "Fourth" intervals and is located at the base of the "Gordon" sandstone (Figure 4). The Rowlesburg Formation may locally contain intervals of appreciable sandstone, however, these sandstones, which often have a "fining-upwards" gamma-ray signature, are highly lenticular.

The red shale lithosome thins westward as a result of the stratigraphic rising of the underlying sandstones of the Cannon Hill Formation. The unit thins from 530 feet at Rowlesburg, to approximately 150 feet in eastern Harrison, eastern Lewis and central Marion Counties. Farther to the west, the Rowlesburg rapidly grades into a thick sequence of massive sandstones. These sandstones represent the Venango Formation. This "vertical" contact is drawn at the location where sandstone occurrence exceeds 50 percent (Figure 5).

The upper contact of the Rowlesburg Formation is placed at the highest extensive red shale. The stratigraphic position of the top of the Rowlesburg shifts only slightly across the study area, in marked contrast with the highly diachronous lower contact. However, the upper contact does shift upsection as traced both to the south, in the vicinity of the West Virginia Dome, and to the east. The stratigraphic rising of the contact from the base of the Oswayo Member of the Price Formation to the base of the Riddlesburg Member in the western panhandle of Maryland is described by Bjerstedt (1986).

Environmentally, the Rowlesburg Formation represents the sub-aerially exposed delta and alluvial plain deposits that followed the prograding "Catskill" shoreline westward through Late Devonian time. The lithologies described from the Rowlesburg and Hannahsville outcrops are indicated in Figure 11 and the appendix.

The Rowlesburg Formation is divided into several informal intervals that correspond to the "Fifth", "Fourth", and "Gordon" sandstones of well drillers. The unit also includes equivalents of the lower part of the driller's "Thirty-foot" sandstone. However, because the "Thirty-foot" straddles the formational boundary, it is not particularly useful for subsurface study and is not recognized in this report.

Venango Formation

The Venango Formation, described from northwestern Pennsylvania (Lesley, 1892), has never been formally defined in West Virginia. In Pennsylvania, the Venango is currently described as a "Group" although no formations have been assigned to it (Berg and others, 1985). In West Virginia, this interval is known as the "Gordon" sandstone, a thick sequence of amalgamated sandstones that occurs in a narrow belt with north-south trend in the central part of the state. Venango Formation sandstones mark the westernmost progradation of Catskill sandstones. The Venango Formation is subjacent to the marine Riceville Formation, and overlies the marine Greenland Gap Formation. Because the unit does not crop out in West Virginia, and no adequately-defined names for the unit exist in neighboring areas, the authors have elected to propose the term Venango Formation be extended to West Virginia and be accepted as the formalized terminology. The well-log of the Harrison 1076 well (Figure 7) is presented as a reference log for the unit.

The Venango Formation represents the shoreline sandstones deposited at the

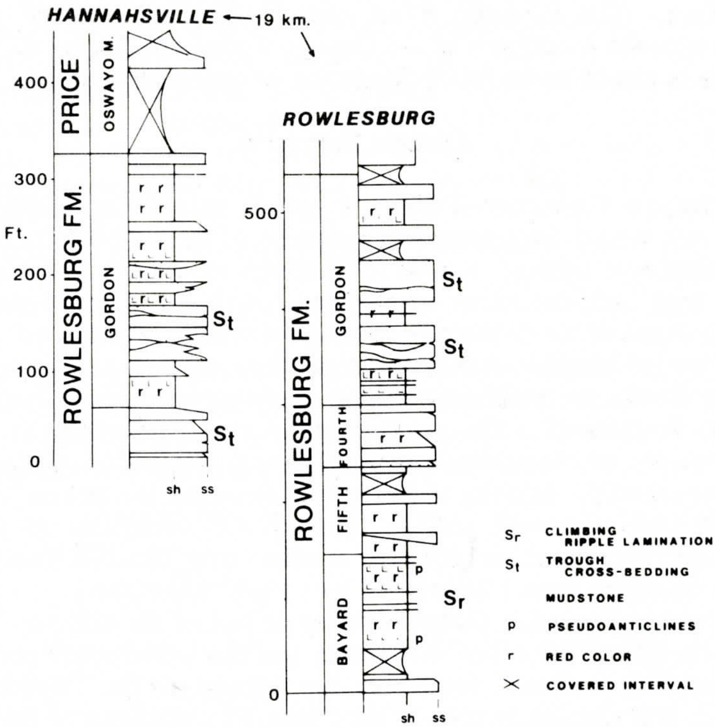

Figure 11. Measured sections of the Rowlesburg Formation from the outcrops located near Hannahsville, Tucker County (reference section) and Rowlesburg, Preston County (type section). The location of the outcrops is shown in Figure 14.

westernmost extent of "Catskill" progradation (Boswell, 1985). Slowed progadation of the clastic wedge resulted in the vertical stacking of these sandstones, creating a thick, narrow belt of "clean" sandstones known to drillers as the "Gordon" sandstone. Isopach maps of individual Venango shoreline sandstones indicate that marine processes were highly effective in sweeping sand into straight, continuous belts trending along depositional strike. The greater marine dominance, as well as the vertical stacking, results in fewer shale interbeds in the Venango Formation than in the genetically-similar shoreline deposits of the Cannon Hill Formation. The top of the Venango is marked by a thin tongue of sandstone that prograded up to 30 kilometers (18 miles) westward of the Venango trend in Wetzel and Marshall counties known as the "Gordon Stray sandstone tongue". The Venango grades into shale and siltstone units of the "Upper Greenland Gap" interval to the west, and the red mudstone-rich Rowlesburg

Formation to the east.

Riceville Formation

The Riceville Formation (Figures 4 and 7) represents the Greenland Gap-like lithologies that transgressed eastward across Venango shorelines in latest Devonian time. The Riceville forms eastward-thinning siltstone wedges that intertongue with the sandstones of the Oswayo Formation (Figure 4). The top of the Riceville is placed at the Berea Sandstone or equivalent horizon.

Oswayo Formation

The Oswayo Formation (Figures 4 and 7) consists of onlapping marine sandstones that record the transgressive demise of the "Catskill" clastic wedge. Oswayo sandstones, although similar to Venango sandstones as interpreted from gamma-ray logs, lack the minor interbedded red shales of the Venango. Also, eastern equivalents of the Oswayo Formation are tabular, bioturbated, fossiliferous siltstones (Oswayo Member of the Price Formation), whereas eastern equivalents of the Venango Formation are channel-filling sandstones and intercalated red shales (Rowlesburg Formation). The presence of marine equivalents to the east of Oswayo Formation sandstones indicates that Oswayo Formation sandstone belts do not represent shoreline deposits as has been interpreted for similar Venango and Cannon Hill units (Boswell, 1985). Instead, the sandstones of the Oswayo Formation are interpreted as offshore marine sand bars (Boswell, 1985) or periodically emergent barrier islands fed by longshore currents.

The Oswayo Formation includes the upper part of the drillers' "Thirty-foot" sandstone, the entire "Fifty-foot" sandstone, and the westernmost portions of the "Gantz" sandstone. These sandstones are known as the "Hundred-foot" in Pennsylvania, and also likely include the "Area F" sandstones of Pepper, deWitt and Demarest (1954).

The Oswayo Formation grades eastward into gray, marine-fossiliferous siltstones of the Oswayo Mamber of the Price Formation. Because the Oswayo Formation sandstones extend farther westward into the subsurface than the Rockwell or Riddlesburg Members of the Price, the Oswayo Member is upgraded to formational status where sandstone concentration within the interval exceeds 50 percent (Figure 4).

Berea Sandstone

The Berea Sandstone (Figures 4 and 8) is poorly developed in the northern West Virginia subsurface. It occurs as either a thin (1.5 meters or 5 feet) sheet sandstone represented by a single, or occasionally double, gamma-ray "spike"; or as a sandstone filling the Gay-Fink channel (Pepper, deWitt and Demarest, 1954). The unit is a convenient marker bed in the areas to the west of the pinchout of the sandstone-rich part of the "Catskill" wedge, where it serves to divide the Devonian Greenland Gap and Riceville Formations from the Mississippian Cuyahoga Formation. In central West Virginia, south of the Gay-Fink channel, and above the feature known as the West Virginia Dome, no sandstone occurs at the Berea horizon. The sheet sandstone lithology also disappears as the Berea is traced eastward. The lack of any sandstone at the Berea horizon greatly complicates correlation of the Berea of western West Virginia and Ohio, to the latest Devonian sandstones of eastern West Virginia. Because of its importance along the western margin of the basin, many workers have attempted to identify the Berea on outcrop to the east, resulting in suspect correlations. Units ranging from the Cloyd Conglomerate (Potter and others, 1983) through the Rockwell Member of the Price Formation (in Bjerstedt, 1986) and the Rockwell Formation of the Broad Top basin area (Reger, 1927) have been labelled "Berea". Because it is lithostratigraphically unsound to attempt to apply the name Berea to units that may be time equivalent, but are lithologically distinct, and for which correlation cannot be demonstrated, the name Berea should be limited to those units that can be physically correlated with the Berea trends mapped and described in Ohio and western West Virginia by Pepper, deWitt and Demarest (1954). In northern West Virginia, this includes only the channel and sheet sandstone units.

Price Formation

Price Formation is the favored terminology for the stratigraphic units in West Virginia previously referred to as "Pocono Formation" or "Pocono Group" (Kammer and Bjerstedt, 1986). The Price Formation at Rowlesburg (Figures 6 and 12) is lithologically diverse, containing both marine and non-marine shales,

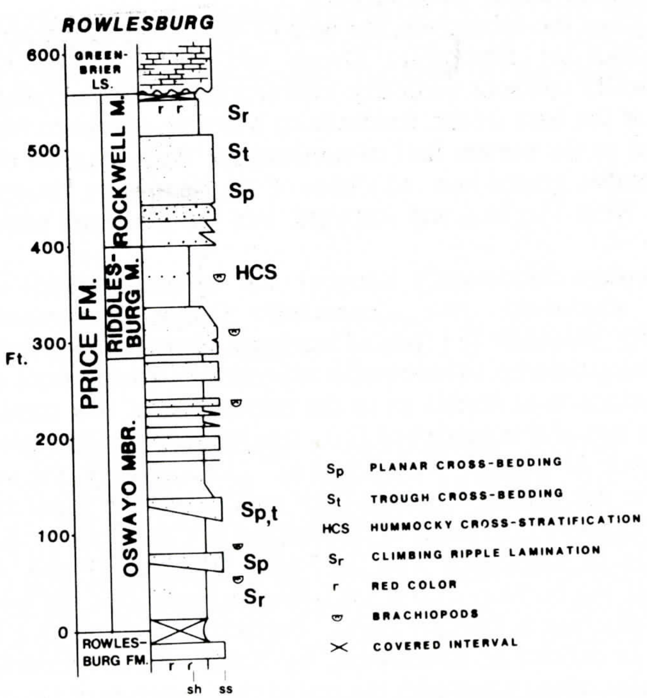

Figure 12. Measured section of the Price Formation at Rowlesburg, Preston County. The location of the outcrop is given in Figure 14.

siltstones and sandstones. Consequently, the Price is defined by its position in sequence, occurring between the uppermost Hampshire red bed and the lowermost Maccrady or Greenbrier unit. The basal contact with the Hampshire Group becomes slightly younger both to the south, over the West Virginia dome, and to the east. However, in most of northern West Virginia, the contact remains consistently within the "Thirty-foot" sandstone interval. Throughout northern West Virginia, the upper boundary of the Price Formation is marked by an unconformable contact with the Greenbrier Limestone. In outcrop, Kammer and Bjerstedt (1986) recognized four members of the Price Formation in northern West Virginia: the chiefly Devonian Oswayo Member, and the Lower Mississippian Cussewago Sandstone, Riddlesburg Shale, and Rockwell members. This paper recognizes the Oswayo, Riddlesburg and Rockwell as Members of the Price Formation.

Oswayo Member: The Oswayo Member of the Price Formation is described from the Rowlesburg outcrop by Kammer and Bjerstedt (1986). Their identification extended the Oswayo southward from Pennsylvania where it is generally related to the type Oswayo of southwestern New York State (Caster, 1934). In West Virginia, the Oswayo consists of thin, tabular, tan-to-gray siltstones and shales. Many units are thoroughly bioturbated and contain a brackish-marine fauna. The occurrence of marine-influenced Oswayo overlying Hampshire lithologies indicates a major and rapid transgression in latest Devonian time. This same transgression is responsible for the tongue of marine Riceville Shale in the northcentral West Virginia.

Throughout the subsurface, the base of the Oswayo Member is marked by the last appearance of Hampshire Group red shales. The Oswayo Member characteristically coarsens vertically, culminating in the overlying massive, marine sandstones at the base of the Riddlesburg Member. Oswayo Member lithologies are restricted to the eastern half of northcentral West Virginia (Figure 13a). The Oswayo Member grades into red shales of the Hampshire Group both southward, into central West Virginia, and eastward, into the Maryland panhandle (Bjerstedt, 1986).

"Cussewago Sandstone": Kammer and Bjerstedt (1986) extend the name Cussewago southward from Conemaugh Gorge, Pennsylvania to describe approximately 19 meters (60 feet) of marine sandstones that overlie the Oswayo at the Rowlesburg outcrop. However, considerations arising from study of this unit in the subsurface raise doubts as to the correctness of this correlation. The best test as to the age of this sandstone is its stratigraphic position relative to the Berea sheet sandstone of western West Virginia. Unfortunately, the subsurface data is ambiguous on this matter. The unit thins rapidly into the subsurface, pinching out within 24 kilometers (15 miles) of the outcrop belt and does not extend far enough to the west to be juxtaposed with the Berea Sandstone (Figure 4). Nonetheless, correlation of the horizon of the sandstone suggests that the unit is most likely stratigraphically higher than the Berea. Furthermore, correlation indicates that the Cussewago as defined at Rowlesburg by Kammer and Bjerstedt (1986) is in a slightly angular relationship with the underlying upper-most Oswayo units (Figure 14). If this angularity represents a slight unconformity, most likely formed during Berea time, then Kammer and Bjerstedt's Cussewago might represent a post- Berea shoreline system, in which case the name "Cussewago" would not apply. As presently defined, the Cussewago is older than Berea (Pepper, deWitt and Demarest, 1954).

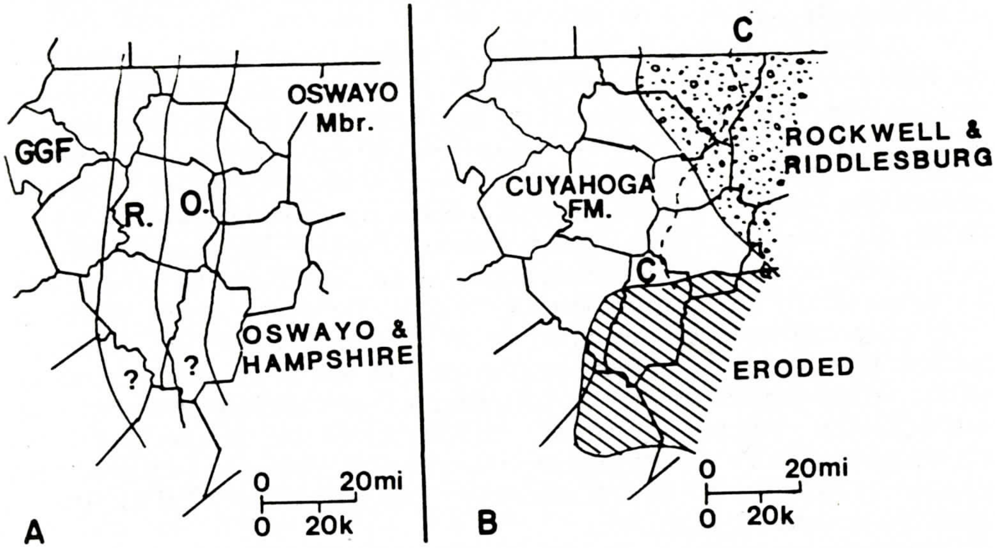

Figure 13. A) Map showing the areal extent of units equivalent to the Oswayo Formation; GGF = Greenland Gap Formation; R = Riceville Formation; O = Oswayo Formation; "Oswayo and Hampshire" = in this area, only a very thin sandstone occurs which may be equivalent to the Oswayo Formation. It is not clear if the Oswayo is missing in this area because of erosion, or because it has changed facies into red shales of the Rowlesburg Formation of the Hampshire Group.

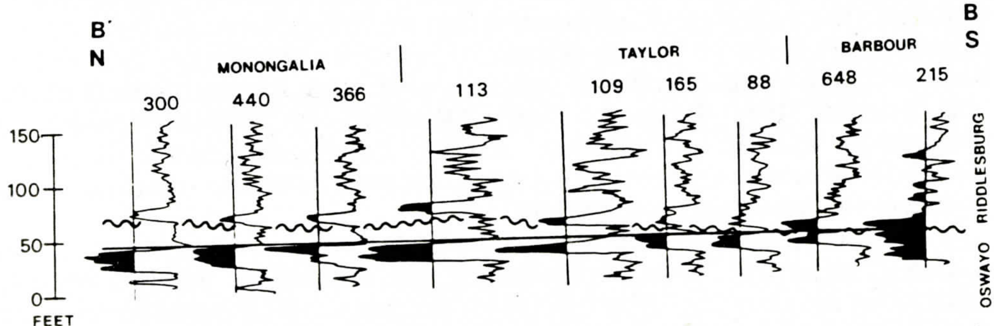

Figure 14. North-south gamma-ray cross-section showing the stratigraphic relationship between the basal Riddlesburg sandstone and underlying upper Oswayo units. The interpreted sub-Riddlesburg unconformity cuts down to the south, and may have been formed during Berea time. Kammer and Bjerstedt (1986) place this unconformity at the top of the basal Riddlesburg sandstone. Note also the increased percentage of sandstone in the Riddlesburg interval in the southernmost well. The position of this cross-section is given on Figure 1.

Although the authors prefer to correlate the Berea Sandstone with units below this 60 foot-thick sandstone, the possiblility of correlating to some position above the "Cussewago" can not be precluded. Bjerstedt (1986) reports finding a 3 centimeter thick clay layer at the top of the Cussewago Sandstone Member that he believes is evidence of subaerial exposure. However, in the subsurface, the top of this unit commonly grades into the overlying Riddlesburg Member with no unconformity suggested.

If Berea equivalents are located below this unit, as the authors suspect, then the "Cussewago" of Kammer and Bjerstedt (1986) is in fact a basal Riddlesburg sandstone, and "true" equivalents of the Cussewago would occur in beds attributed to the uppermost Oswayo. These upper Oswayo units grade to the west into siltstones which are locally removed by erosion along the Berea "Gay-Fink" channel (Rittenhouse, 1946). Although suspected to be Cussewago by Pepper, DeWitt and Demarest (1954), these sandstones ("area F fan") correlate to less sandy marine beds at the Rowlesburg outcrop and as such should be included with the lithologically and environmentally similar Oswayo. As a result, no Cussewago sandstone can be conclusively identified within the subsurface. Because of the ambiguity of the subsurface data, as well as the poor development of this particular sandstone, we do not recognize the Cussewago Member in the northern West Virginia subsurface. Having given the proper caveat, our correlations will show that the sandstone identified as Cussewago by Kammer and Bjerstedt (1986) to be the basal Riddlesburg. The extent of the basal Riddlesburg sandstone is shown in Figure 13b.

Riddlesburg Member: The Riddlesburg Member is a tongue of marine lithologies of the Cuyahoga Formation that penetrates eastward into the Hampshire-Price sandstones (Figure 4). The Riddlesburg Member is bounded by sandstones of the Oswayo Member below and the Rockwell Member above. There is a massive sandstone at the base of the unit in the Rowlesburg outcrop (Cussewago Sandstone Member of Kammer and Bjerstedt, 1986), overlain by 16 meters (52 feet) of brown-to-gray marine siltstones and shales exhibiting storm-generated features. The percentage of sandstone within the interval increases greatly in the southern part of the study area (Figure 14), where the Riddlesburg-aged shelf is believed to have shoaled under the influence of the West Virginia Dome (Kammer and Bjerstedt, 1986; Bjerstedt, 1986). Westward, the boundary between the Riddlesburg Member of the Price, and the Cuyahoga Formation is located arbitrarily where the thickness of overlying Rockwell sandstones decreases below 3 meters (10 feet) (Figures 4 and 13b).

Rockwell Member: The Rockwell Member consists of approximately 37 meters (120 feet) of "clean", cross-bedded, and occasionally pebbly sandstones with minor red siltstones. Bjerstedt (1986) interprets a major unconformity at the base of the Rockwell Member that indicates a substantial part of the underlying Riddlesburg Member in northeastern West Virginia has been eroded. The Rockwell Member can be traced approximately 40 kilometers (25 miles) westward into the subsurface where it grades into shale and siltstone lithologies of the Cuyahoga Formation (Figure 13b). The Rockwell Member of the Price is commonly referred to as "Weir" sandstone in the subsurface of northeastern West Virginia.

Cuyahoga Formation

The Cuyahoga Formation (Figures 7 and 8) consists of Lower Mississippian

(Kinderhookian) marine shales, siltstones and sandstones that are the offshore shelf, fan, and basinal equivalents of the Riddlesburg and Rockwell members of the Price Formation. Equivalents of the Cuyahoga are subdivided into a plethora of formations, members, and even sub-members, both formal and informal, in Ohio and Pennsylvania. The relationship of these units with the various lithologies of the West Virginia Cuyahoga are not yet clear. The Cuyahoga of central West Virginia, where preserved north of the West Virginia Dome, is a homogenous sequence of silty shales approximately 110 meters (350 feet) thick. In the western part of the state, thin-bedded sandstones appear in the lower third of the interval. This sandstone lithosome becomes stratigraphically lower to the west, occurring within 30 meters (90 feet) of the Berea Sandstone. When mapped, this unit is a narrow sandstone belt that trends parallel to depositional strike and has been interpreted as a submarine fan deposit (Boswell, 1985). The unit is commonly called the "Weir" sandstone in the West Virginia subsurface. A marked shale reflection in gamma-ray logs at the top of the "Weir" sandstone can be traced eastward across the finer-grained part of the Cuyahoga to within the Rockwell Member of the Price Formation. The "Weir" is overlain by 40 meters (130 feet) of shales and siltstones, and eventually by a massive clean sandstone up to 60 meters (200 feet) thick, known to well drillers as the "Big Injun". The "Big Injun" is truncated by the sub-Greenbrier unconformity to the east. This sandstone occupies the same stratigraphic position as the Black Hand Member of the Cuyahoga Formation of central Ohio (Majchszak, 1984). The extent of the Cuyahoga Formation in West Virginia is shown in Figure 13b.

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APPENDIX

The Rowlesburg outcrop, designated as the type section of the Cannon Hill and Rowlesburg Formations, consists of a series of small exposures along the railroad to the west of Rowlesburg (Figure 15a). The outcrop can be reached by an unimproved road that crosses the railroad 1.2 kilometers (.75 miles) west of the center of Rowlesburg. This unimproved road passes by the east end of the railroad bridge over Fill Hollow. The Greenland Gap-Cannon Hill contact is placed at the base of the uppermost sandstone that occurs in the outcrop on the east end of the bridge. The Cannon Hill-Rowlesburg contact is placed 0.8 kilometers (2700 feet) farther west at the base of a red sandstone that crops out near two large train signal standards. The upper contact of the Rowlesburg Formation occurs at the top of the red shale that occurs on the west bank of Tray Run, approximately 0.95 kilometers (3200 feet) northwestward along the track. The locations of the Hannahsville and Elkins reference sections are shown in Figures 15b and c.

The gamma-ray log from the Harrison 1076, presented as a reference log for the Venango and Oswayo Formations in West Virginia is located 2.24 miles south, and 3.38 miles west of the intersection of latitude 39° 20' and longitude 80° 20'.

Columnar sections of the Rowlesburg and Cannon Hill Formations exposed west of Rowlesburg, south of Hannahsville, and east of Elkins are presented in Figures 10, 11, and 12. Below is the lithologic description of the Rowlesburg type section of the Cannon Hill and Rowlesburg Formations.

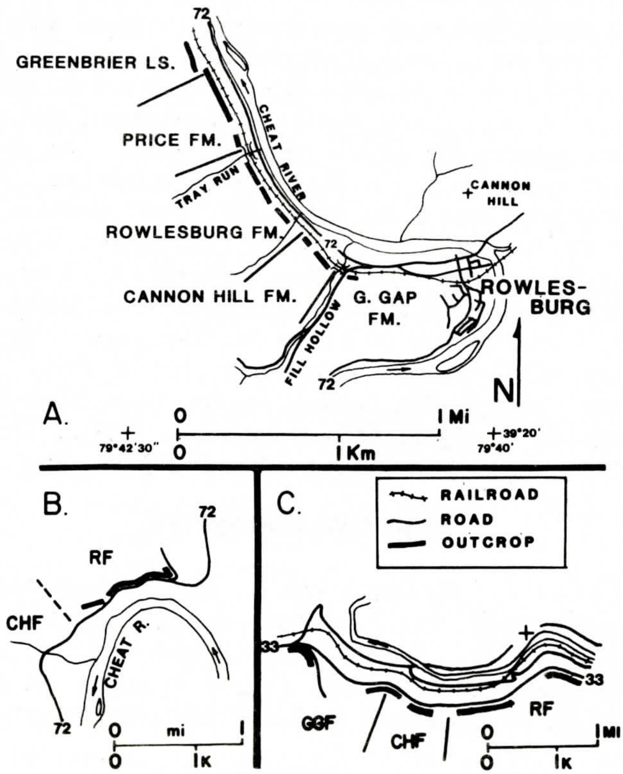

Figure 15. Location maps for outcrops described in this report. A) The outcrop west of Rowlesburg, Preston County; B) The outcrop south of Hannahsville, Tucker County; C) The outcrop east of Elkins, Randolph County. The following symbols were used: GGF = Greenland Gap Formation; CHF = Cannon Hill Formation; RF = Rowlesburg Formation.

Type section of the Cannon Hill and Rowlesburg Formations
Rowlesburg, West Virginia

HAMPSHIRE GROUP (763 ft.)

Cannon Hill Formation (233 ft.)

"Warren" interval (123 ft.)

SANDSTONE, medium-grained, argillaceous, yellowish-gray, large trough cross-sets, minor conglomeratic beds, internal scour surfaces, basal lag of quartz, shale clasts, and large plant debris............14 ft.

COVERED INTERVAL..30 ft.
SANDSTONE, basal medium-grained, fining-up, friable, conglomeratic, clean, very light gray sandstone. Planar cross-beds at base, graded horizontal beds at top, upper contact sharp, minor lenses of dark shale..11 ft.
SILTSTONE and SHALE, argillaceous, medium grey siltstone and grey to greyish-red shales, thinly laminated, horizontally burrowed, with small rhynconellid brachipods...3 ft.
SANDSTONE, medium-grained, clean, very pale orange, heavy mineral laminations, slightly scoured basal contact, bedding alternates between low-angle planar and horizontal sets, increasing size and inclination of foresets vertically..27 ft.
SHALE and MUDSTONE, coarsening upwards into SILTSTONE, light brown, medium light gray and greenish gray at base, become pale reddish brown mudstone with pseudoanticline structures...........................8 ft.
SANDSTONE, basal fine-grained, coarsening-up, argillaceous, light brown, very large high-angle, bidirectional trough-cross sets, casts of molluscs at base. Thin, unfossiliferous dark grey shale at middle...20 ft.
SHALE and SILTSTONE, poorly exposed, thin-bedded, argillaceous, horizontally laminated...11 ft.

"Elizabeth" interval (110 ft.)

SANDSTONE, fine-grained, fining-upwards, very pale orange, basal lag of shale clasts, quartz pebbles and rock fragments, trough cross-bedded, minor planar cross-bedding; internal friable, medium-grained sand containing large macerated plant fragments. Top of unit is massive, ripple-drift laminated very fine-grained sandstone...32 ft.
MUDSTONE and SHALE with interbedded SILTSTONE, pale reddish brown with minor amounts of light gray and light brown, thinly bedded, internal scour surfaces, pseudoanticlines are common in the mudstones..22 ft.
SANDSTONE, fine-grained, fining-up, very pale orange, quartz-rich, trough cross-bedded, indistinct lateral accetion surfaces (epsilon bedding), large plant fossils at base, grades into overlying siltstone..15 ft.
SILTSTONE, fining-upward, pale red, climbing-ripple laminations, top of unit in channel morphology...19 ft.
SHALE, pale red, changing vertically to pale olive, minor tabular siltstones, contains an assemblage of molluscs, and a single specimen of rhinocarid arthropod at top of unit.....................................22 ft.

Rowlesburg Formation (Hampshire Group) (530 ft.)

"Bayard" interval (148 ft.)

SANDSTONE, fine-to-medium grained, argillaceous, moderate red, with large planar cross-beds..13 ft.
COVERED INTERVAL..30 ft.
MUDSTONE, SHALE and SILTSTONE with minor SANDSTONE, tabular moderate red siltstones and pale olive, fine-grained sandstones encased in thick light red mudstones and shales. Mudstones commonly contain root traces and pseudoanticlines..105 ft.

"Fifth" interval (92 ft.)

SHALE, moderate red, with minor ripple-bedded sandstone..28 ft.
SANDSTONE, medium-to-fine grained, basal lag of quartz and plant fragments, indistinct epsilon bedding. Removed in places by light olive gray channel sandstone..11 ft.
SHALE, moderate red..10 ft.
SANDSTONE, fine-grained, light brown, trough cross-bedded, with internal reactivation surfaces marked by lags of shale clasts..9 ft.
COVERED INTERVAL..34 ft.

"Fourth" interval (68 ft.)

SANDSTONE and SHALE, fine-grained, cross-bedded, very pale orange sandstones, planar cross-bedded, intertounging laterally with light red and greenish-gray shales. Shales commonly show lateral accumulation and slump features..38 ft.
SANDSTONE with SILTSTONE and SHALE, fine-grained, very pale orange, friable, large macerated plant fragments, intertonguing laterally with thin-bedded siltstones and shales, grey, tan, and red..20 ft.
SANDSTONE, fine-grained, grayish-orange, wavy bedding at base, replaced vertically by trough cross-sets, large plant fragments..10 ft.

"Gordon" interval (including "Gordon Stray" and equivalents of lower "Thirty-foot" sandstones) (222 ft.)

MUDSTONE and SHALE, moderate red, mottled green by rooting. Thin tabular siltstones and fine sandstones abundant near base..36 ft.
SANDSTONE, fine-grained, argillaceous, pale red or very pale orange, basal 20 feet is amalgamtion of three distinct sands separated by lenses of dark-grey shale exhibiting slump features. Bedding grades vertically from small scale festooned trough sets into climbing-ripple cross-lamination..38 ft.
SHALE and SILTSTONE, moderate to light red, siltstones thin-bedded

and tabular..21 ft.
SANDSTONE, pale red, argillaceous, trough and ripple-cross bedded, poorly exposed..42 ft.
COVERED INTERVAL..15 ft.
SANDSTONE, medium-to-fine grained, light brown, argillaceous, trough cross-bedded..12 ft.
MUDSTONE and SHALE, basal moderatered mudstone with rooting, coarsening-up..24 ft.
SANDSTONE, medium-to-fine grained, very pale orange, large tangential planar cross-bed sets at the base..16 ft.
COVERED INTERVAL..18 ft.

PRICE FORMATION (486 ft.)

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