Appalachian STATE UNIVERSITY
BOONE, NORTH CAROLINA

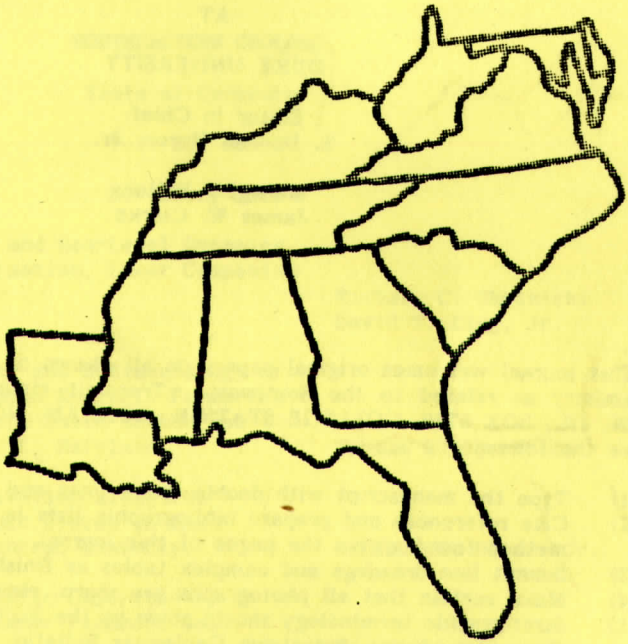
Southeastern Geology: Volume 27, No. 2
November 1986

Edited by: S. Duncan Heron, Jr.

Abstract

Academic journal published quarterly by the Department of Geology, Duke University.

SOUTHEASTERN GEOLOGY



PUBLISHED AT DUKE UNIVERSITY DURHAM, NORTH CAROLINA

VOL. 27, NO. 2 NOVEMBER, 1986

SOUTHEASTERN GEOLOGY

PUBLISHED QUARTERLY

AT

DUKE UNIVERSITY

This journal welcomes original papers on all phases of geology, geophysics, and geochemistry as related to the Southeast. Transmit manuscripts to S. DUNCAN HERON, JR., BOX 6729, COLLEGE STATION, DURHAM, NORTH CAROLINA 27708. Observe the following:

1) Type the manuscript with double space lines and submit in duplicate.
2) Cite references and prepare bibliographic lists in accordance with the method found within the pages of this journal.
3) Submit line drawings and complex tables as finished copy.
4) Make certain that all photographs are sharp, clear, and of good contrast.
5) Stratigraphic terminology should abide by the North American Stratigraphic Code (Am. Assoc. Petroleum Geologists Bulletin, v. 67, p. 841-875).

Proofs will be sent authors.

Reprints must be ordered prior to publication; prices available upon request. Subscriptions to Southeastern Geology are $9.00 per volume (US and Canada), $11.00 per volume (foreign). Inquiries should be sent to: SOUTHEASTERN GEOLOGY, BOX 6729, COLLEGE STATION, DURHAM, NORTH CAROLINA 27708. Make checks payable to: Southeastern Geology.

SOUTHEASTERN GEOLOGY

Table of Contents

Vol. 27, No. 2

November, 1986

1. Depositional Facies and Sea-Level Dynamics of the Blufftown Formation, Lower Campanian of east Alabama. Michael C. Skotnicki, David T. King, Jr. 53

2. Regional Stratigraphy and Sedimentology of the Lower Mississippian Rockwell Formation and Purslane Sandstone based on the new Sideling Hill road cut, Maryland. Thomas W. Bjerstedt 69

3. Mississippian Reactivation along the Irvin-PaintCreek Fault System in the Rome Trough, east central Kentucky. Garland R. Dever. Jr. 95

4. Mineralogy of Ultramafic Chlorite-Amphibole Schists, Inner Piedmont Belt, South Carolina. Richard D.Warner, Villard S. Griffin, Jr., Robert C. Hoover, J. Gregory Bryan, Stephen H. Poe, Jeffrey C. Steiner 107

DEPOSITIONAL FACIES AND SEA-LEVEL DYNAMICS OF THE BLUFFTOWN FORMATION, LOWER CAMPANIAN OF EAST ALABAMA

MICHAEL C. SKOTNICKI
DAVID T. KING, JR.

Department of Geology, Auburn University
Auburn, AL 36849

ABSTRACT

The Upper Cretaceous Blufftown Formation (Lower Campanian), approximately 150 m thick, crops out in the northern Coastal Plain of east Alabama and western Georgia. The Blufftown thickens to the west and in east-central Alabama intertongues with the age-equivalent Mooreville Chalk. The study area is the outcrop of the Blufftown Formation in Alabama and includes the Blufftown-Mooreville transition zone. The Blufftown Formation consists of the deposits of a linear clastic barrier-island shoreline and an associated open-marine inner shelf. This shoreline and shallow-marine system existed on the northeastern rim of the Gulf of Mexico in Early to Mid-Campanian time and deposited sedimentary facies over eastern Alabama according to its geographic position as dictated by relative changes in sea-level and apparent shifts in shoreline strike. The Blufftown is composed of nine sedimentary facies: 1) barrier-island lower shoreface, 2) barrier-island middle shoreface, 3) tidal-inlet, 4) back-barrier marsh, 5) washover fan, 6) palimpsest bed, 7) inner shelf, 8) storm sands, and 9) mid-shelf. Deposition of the Blufftown in the eastern part of the study area was dominated by barrier-island sands and associated back-barrier lagoonal sediments, while in the western part of the study area sedimentation was dominated by shallow marine sandy, glauconitic silts and slightly sandy, calcareous clays that accumulated in great thickness. At the same time, rare-event storm sands were deposited in this shallow-marine setting giving evidence of episodic punctuated mixing of shoreline clastics with open-marine sediments. Correlation has distinguished four genetic packages of facies bounded by facies discontinuities. Vertical changes in facies within and between genetic packages has allowed interpretation of sea-level history and a sea-level curve for Blufftown Formation time (Early to Mid-Campanian) has been created. This curve shows three unequal cycles of sea-level rise and fall.

INTRODUCTION

The purpose of this paper is to describe the depositional facies of the Blufftown Formation and to use the stratigraphic relations of the facies to interpret the history of sea-level change during Blufftown deposition. The Blufftown Formation is Early to Mid-Campanian in age (Sohl and Smith, 1980) and crops out in the northern Gulf Coastal Plain from western Georgia to east-central Alabama where it gradationally intertongues with the laterally equivalent Mooreville Chalk (Figure 1). The study area is the outcrop area of the Blufftown in Alabama including the transition zone with the Mooreville Chalk (Figure 2, Appendix 1). Appendix 1 lists the locations of typical exposures of the Blufftown depositional facies.

As Figure 1 shows, the Blufftown-Cusseta Sand contact is generally regarded as disconformable (Monroe, 1941) and an erosional surface with associated paleontological break (Sohl and Smith, 1980) in the western part of the study area confirms this. The nature of the contact between the underlying Tombigbee Sand Member of the Eutaw Formation and the lowermost eastern Blufftown is unclear (Sohl and Smith, 1980). Figure 1 shows this relationship as it is typically illustrated, as a disconformity (Monroe, 1941).

In Alabama the Blufftown is approximately 150 m thick. It crops out in a belt approximately 30 km wide that extends across portions of Russell,

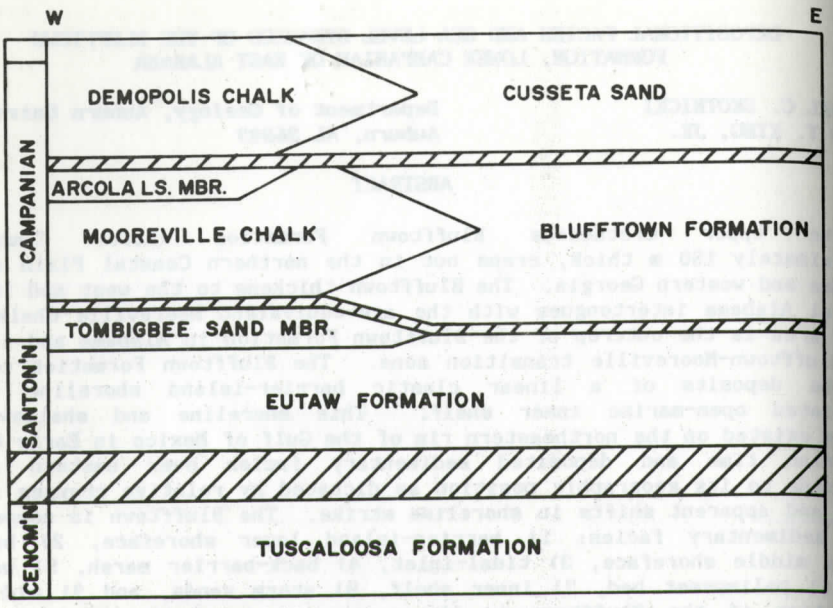


Figure 1. Generalized stratigraphic relations of formal stratigraphic units in the Upper Cretaceous of eastern and central Alabama. Stratigraphic relations after Monroe (1941) and age assignments after Sohl and Smith (1980).

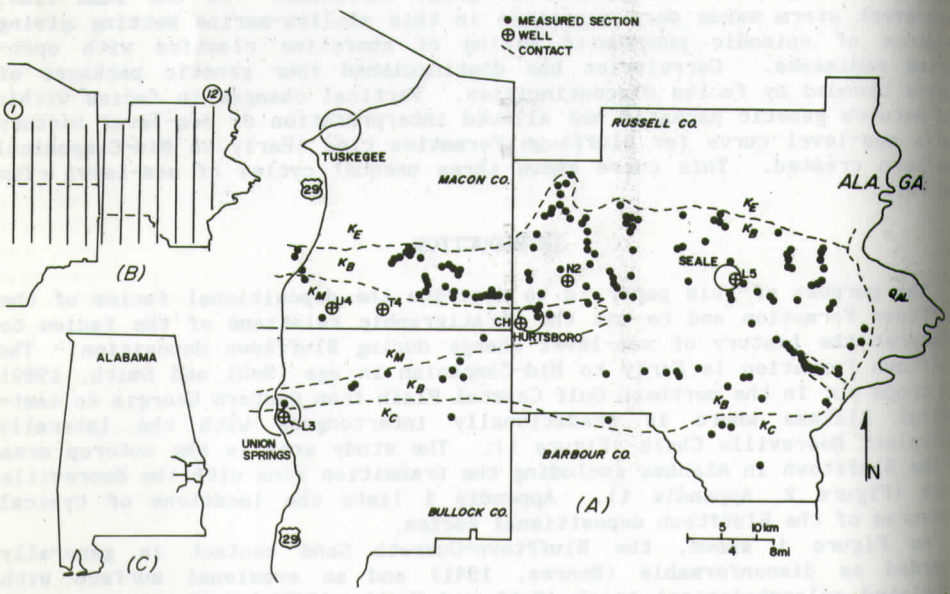


Figure 2. (A) Locations of measured sections and wells in the study area of Russell, Barbour, Macon, and Bullock Counties, Alabama. The contacts between stratigraphic units are modified from Eargle (1950). Lettering system for wells (L3, etc.) follows the Alabama Geological Survey reports (cited in text). Map units: Ke (Eutaw Fm.), Kb (Blufftown Fam.), Km (Mooreville Chalk), Kc (Cusseta Sand), Qal (alluvium). (B) Locations of the 12 lines used for composite measured sections in Figure 3. (C) Location of study area in Alabama (county outlines are shown).

Barbour, Bullock, and Macon Counties (Figure 2). The Blufftown extends in the subsurface to the south due to its dip of 7.5 m/km (Copeland, 1968).

Previous work on the Blufftown Formation in the study area has focused mainly on its physical stratigraphy (Monroe, 1941 and 1947; Eargle, 1950; and Copeland, 1968) or its paleontology (Nikravesh, 1967: Sohl and Smith, 1980; and Schwimmer, 1981). Only recently has any attempt been made to study the facies stratigraphy of the Blufftown (Reinhardt, 1980; Skotnicki, 1985; King and Skotnicki, 1986).

METHODS

Over 140 surface exposures of the Blufftown Formation in east Alabama were measured, described, and sampled for this study (Figure 2). The surface exposures consisted of road cuts, railroad cuts, creek banks, and sand pits. Surface exposures were used extensively because of their relative abundance and the fact that little subsurface data are available for the Blufftown. The measured sections of surface exposures have been used to graphically represent the nature and thickness of the Blufftown strata in a correlation diagram (Figure 3). This was done by projecting measured sections onto north-south trending lines (perpendicular to Blufftown strike) spaced at 5.5 km intervals (Figure 2B). For correlation, the relative positions of the measured sections were then adjusted using basal elevation and depositional dip. Driller's chip logs of six wells, taken from publications of the Alabama Geological Survey (Scott, 1960, 1962, and 1964) were added to the correlation diagram to give continuous columns of lithologic data.

Laboratory analyses of samples have included use of a binocular microscope for examination and identification of mineralogy, texture, fossils, and carbonate content. Unconsolidated samples were sieved or hydrometered and statistical analyses performed on individual data to determine mean grain size and standard deviation.

STRATIGRAPHIC CORRELATION

The correlation diagram of Blufftown depositional facies is shown in Figure 3. Because of the scale of this diagram, individual depositional facies have been grouped into facies ensembles (e.g., barrier sands or lagoonal sediments). The individual facies (e.g., barrier lower-shoreface and barrier middle-shoreface) that compose these ensembles are discussed separately in the following section. Correlation of the facies ensembles suggests the Blufftown Formation consists of four genetic packages of facies. The boundaries between packages are facies discontinuities that represent rapid, relative sea-level change.

The datum used for the correlation diagram (Figure 3) was the Blufftown Formation-Cusseta Sand contact. This contact was used because there are no continuous, distinctive beds within the Blufftown Formation which provide a natural datum.

DEPOSITIONAL FACIES

The Blufftown Formation is composed of nine major depositional facies distinguished on the basis of Selley's (1978 criteria: 1) lithology, 2) sedimentary structures, 3) fossil content, 4) paleocurrents, and 5) geometry, in addition to the facies' stratigraphic relations. The depositional facies of the Blufftown are: 1) barrier-island lower shoreface, 2) barrier-island middle shoreface, 3) tidal-inlet, 4) back- barrier marsh, 5) washover fan, 6) palimpsest bed, 7) inner shelf, 8) storm sands, and 9) mid-shelf. For correlation, facies of similar origin were grouped into facies ensembles. The ensembles and individual facies are described below in a general shore-to marine order.

Barrier-Island Ensemble Sands

Lower Shoreface: The lower shoreface deposits of the Blufftown are composed of clayey, micaceous, poorly sorted (ave. st. dev. = 1.08 phi), fine-grained (ave. = 2.69 phi), highly bioturbated sand similar to the modern lower-shoreface sands described by others workers such as McGowen and Lopez (1983) and Schwartz and others (1981). Pirmary sedimentary structures are generally lacking, having been destroyed by the intensive bioturbation of infaunal organisms, but where present consist of thin, planar laminations. Body fossils are rare, but where present consists of *Exogyra*. The *Exogyra* nearly always are disarticulated and commonly are not in life- position orientation. Secondary sedimentary structures generally consist of cylindrical, subvertical to subhorizontal, *Ophiomorpha* that are passively filled by the host sediment. Also present, but rare, are sub-horizontal burrows which show asymmetric, concentric infillings of clay. In surface exposures, where a vertical transition in facies occurs, the lower-shoreface sands are gradationally overlain by middle-shoreface sands.

Middle Shoreface: The middle-shoreface deposits of the Blufftown Formation are composed of micaceous, moderately sorted (ave. st. dev. = 0.79 phi), fine- to medium-grained (ave. = 2.33 phi) sand that contains wedge-shaped sets of thin, low-angle, planar cross-bedding. The sand is commonly burrowed by abundant well-preserved *Ophiomorpha nodosa* which are lined with pellets of fine sand and commonly branch into networks. This lithology is very similar to the modern middle-shoreface sands described by Davies and others (1971) and the ancient ones of Lerand (1983). Commonly, *Ophiomorpha* crowded surfaces are truncated by erosional surfaces with planar cross-bedded sand above. Alternating sequences of *Ophiomorpha*-burrowed and planar cross-bedded sands similar to the "lam-scram" beds of Howard (1972) occur. The burrowed part of the sequence represents a period of slow, continuous sedimentation and intricate burrowing of the stable substrate by suspension-feeding animals. This is followed by storm-wave truncation and sudden deposition of sand in planar cross-bed sets containing small rip-up clay clasts. In surface exposures these middle-shoreface deposits gradationally overlie lower-shoreface sands and are erosionally overlain by coarse-grained tidal-inlet fill sands.

Tidal Inlet: Tidal-inlet fill deposits of the Blufftown are composed of moderately sorted (ave. st. dev. = 0.86 phi), coarse-grained sand (ave. = 1.15 phi) which is trough cross-bedded and commonly contains herringbone trough cross-bedding indicative of bi-directional tidal-current flow. These sands are similar to the modern tidal-inlet sands of Kumar and Sanders (1974) and the ancients ones of Lerand (1983). The bedding is commonly accented by the presence of thick clay drapes and large clay clasts that originated as ripped-up clay linings of *Ophiomorpha*. *Ophiomorpha* are common in the tidal-inlet sand and have thick clay linings which were necessary to prevent burrow collapse within the coarse sand. Paleocurrent data from exposures of herringbone trough cross-bedding are biomodal, and readings are approximately 180 degrees apart. In one exposure, a large permineralized log lies in a paleocurrent orientation subparallel to the directions of the cross-bed sets suggesting it was deposited in a tidal-inlet and was oriented by linear tidal-current flow. In surface exposures, tidal-inlet fill sands erosionally overlie finer grained middle-shoreface deposits and in one place a tidal-inlet sand is erosionally overlain by a palimpsest bed (described below). In a large-scale detailed correlation of Blufftown facies (Skotnicki, 1985) tidal-inlet fill sands are interbedded with back-barrier lagoonal sediments.

Back-Barrier Ensemble Sediments

Marsh: The marsh deposits of the Blufftown are composed of very micaceous, highly carbonaceous, very poorly sorted (ave. st. dev. = 3.54 phi) siltstone and mudstone (ave. > 8.0 phi) similar to the modern back-

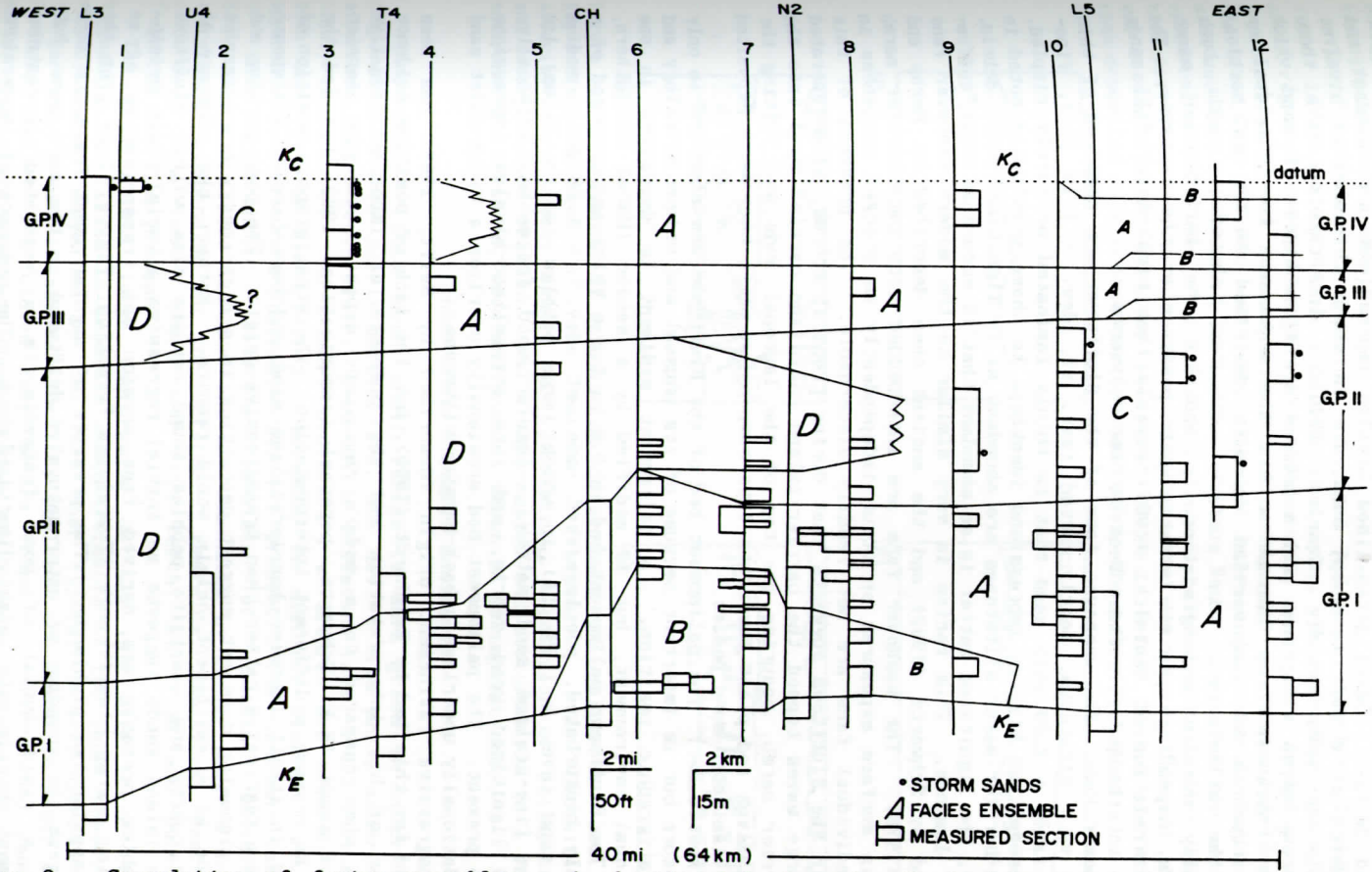


Figure 3. Correlation of facies ensembles: A, barrier-island sands; B, back-barrier sediments; C, inner-shelf sediments; and D, mid-shelf sediments. Boundaries between genetic packages I, II, III, and IV are also indicated. Datum is the Blufftown-Cusseta (Kc) contact. Base of section is the Eutaw (Ke) contact. Locations of storm deposits are indicated by small circles at side of measured sections.

barrier sediments described by Rampino and Sanders (1980). Lignitized plant matter is abundant and occurs as highly disseminated stem material and other macerated debris. Small streamlined bivalve impressions are abundant and are evidence of a soft to soupy substrate in which thin-shelled bivalves were able to move rapidly (Stanley, 1970). Intercalated with these carbonaceous marsh siltstones and mudstones are thin layers of sand which range in thickness from laminae within the mudstones to individual or stacked sequences of washover-fan deposits (described in the next section) within the siltstones. In surface exposures, shoreface sands both erosionally overlie and gradationally underlie the back-barrier marsh sediments, depending upon the stratigraphic position of the exposure. The facies correlation of Skotnicki (1985) reveals that tidal-inlet fill sands are also interbedded with these back-barrier sediments.

Washover Fan: The washover fans of the Blufftown are composed of thin beds of very micaceous, poorly-sorted (ave. st. dev. = 1.15 phi), fine-grained (ave. = 2.39 phi) sand that is thinly laminated or rarely rippled. Small streamlined bivalve impressions identical to those previously noted in the back-barrier marsh siltstones are abundant as is lignitized plant debris. Commonly, the lignitized matter is so abundant that it occurs as thin "coffee-grounds" laminae. This facies is very similar to the modern washover fans described by Schwartz (1982) and the ancient ones described by Horne and others (1978). The washover fans are interbedded with back-barrier marsh facies in surface exposures or occur independently as stacked sequences in which individual fans are not readily discernable. The presence of this facies in the Blufftown suggests that during tropical storms, wind-generated storm-surge waves topped the barrier-island chain and carried sand into the back-barrier marsh, scouring the top of the lagoonal muds and lifting the light bivalves and plant fragments into suspension until they were deposited into the laminated sand bodies.

Palimpsest Bed: The palimpsest bed of the Blufftown was observed in only one exposure but is important because of its unusual and unique petrology and its stratigraphic position. A palimpsest sediment is deposited in one depositional environment, but is modified by a second (Harms and others, 1982). The Blufftown palimpsest bed is 0.6 to 1.2 m thick and composed of a thoroughly bioturbated, poorly-sorted (ave. st. dev. = 1.26 phi), medium grained sand (ave. = 1.35 phi) in which large pebbles (up to 18 mm) lie within a fine-grained sand matrix. Angular wood fragments, vertically-oriented lignitized wood debris, and rare streamlined-bivalve impressions are also present. The palimpsest bed erosionally overlies a tidal-inlet sand and gradationally underlies a back-barrier siltstone.

Transgressive ravinement origin (described by Swift, 1968) has been suggested for this bed by Reinhardt (1980), but the lack of positive evidence in favor of his interpretation and the presence of important negative evidence when compared to a modern ravinement deposit overlying barrier shoreface sands (R.K. Schwartz, personal communication, 1985) has led the authors to propose a different interpretation. The stratigraphic position of this bed in itself (between barrier-island sand and back-barrier siltstone) suggests a regressive rather than transgressive origin. The poor sorting and lack of winnowing does not suggest deposition in a high-energy transgressive surf zone as a ravinement origin would require. Rather, the presence of abundant quartz and quartzite pebbles found nowhere else within Blufftown barrier- island sands suggests an initial regressive alluvial-plain origin. The pebbles probably are derived from erosion and transportation of an underlying coarse clastic unit (Tuscaloosa Formation, Figure 1) or from the nearby Appalachian highlands. The presence of angular wood fragments does not suggest the rounding of water-saturated driftwood in the surf zone, but instead a rapid burial of wood fragments in a vegetated terrestrial environment shortly after plant disarticulation. The presence of impressions of thin-shelled, streamlined bivalves and thorough bioturbation does not suggest a high-energy, surf zone origin for this unusual sand bed. As

discussed previously, such bivalves are commonly present in a back- barrier environment (Stanley, 1970). Further, thorough bioturbation requires low-energy rather than high-energy conditions as would be found in a transgressive surf zone.

Thus, ravinement origin for this sand bed as interpreted by Reinhardt (1980) is less likely than a palimpsest origin. We suggest (Figure 4) that after regression stranded barrier-island sands, they were eroded by surface processes and small streams deposited a poorly sorted alluvial plain sand on the erosional surface above the reworked barrier lithosome. Further, the sand was likely vegetated by coastal flora. Roots penetrated the underlying erosional surface and into the barrier sand. Transgression then submerged this poorly sorted, pebbly sand in a quiet, back-barrier environment where it was bioturbated by bivalves and other fauna. Normal lagoon sedimentation then deposited a carbonaceous siltstone above the palimpsest sand.

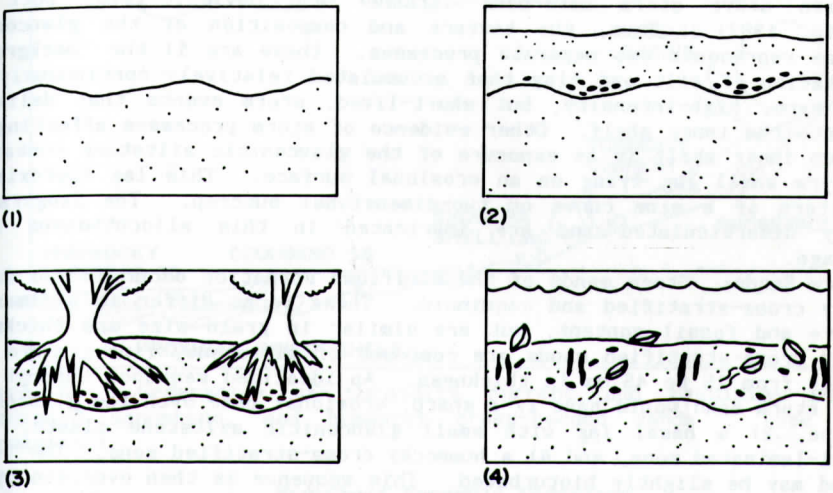


Figure 4. Sequence of events in development of palimpsest bed. (1) Regressive erosion of the barrier lithosome. (2) Deposition of alluvial-plain pebbly sand. (3) Terrestrial vegetation of palimpsest bed and incorporation of wood. (4) Transgression and bioturbation in a lagoonal environment (bivalves, wood and pebbles shown in palimpsest bed).

Inner Shelf Ensemble Sediments

Inner-Shelf: This inner shelf facies is dark green in color and is composed of a very poorly sorted (ave. st. dev. = 3.18 phi), sandy, highly micaceous, calcareous, clayey, glauconitic siltstone (ave. > 6.5 phi) that is highly bioturbated. No primary sedimentary structures are present, having been destroyed by bioturbation. Normally no distinct biogenic texture is apparent; however, rare *Thalassinoides* are present. The carbonate component of this glauconitic siltstone consists of abundant microfossils. Macrofossils are also abundant and are dominated by either *Exogyra* or large inoceramids, depending on the stratum. Ostreids, ammonites, gastropods, calcareous worm tubes, shark and fish teeth, and fish and other bone fragments are also present.

The *Exogyra* and large inoceramids do not typically occur together suggesting they lived in separate subenvironments of the inner shelf. This supposition is supported by the fact that typically, articulated *Exogyra* (larger than 8 cm) in approximate life position are found in sandy strata. Further, smaller articulated *Exogyra* are found in intercalated sands and marls. The sandy substrate suggests that the *Exogyra* are shallow, near-shore pelecypods. In contrast, *Exogyra* in sand-free marls are typically

disarticulated, oriented convex-up, and appear somewhat comminuted suggesting that they are a transported fauna. These disarticulated *Exogyra* occur with whole, large inoceramids. The inoceramids, all of which occur in living position, are interpreted as a deeper water fauna.

The glauconitic siltstone contains a textural inversion (term of Folk, 1980) of fine sand in a clayey silt matrix suggesting different source areas for the two sizes of framework silicates. Thorough bioturbation in the marls suggests that the textural inversion was created by punctuated mixing (term of Mount, 1984) of shoreline sand into the mud-dominated depositional environment of the inner shelf by rare storm events on the Blufftown shelf. The storm hypothesis is supported by the presence of hummocky cross-stratified and coquinoid sand beds (described in next section) within the glauconitic siltstone. Such sand beds are generally interpreted to represent deposition above storm wave-base (Brenner and Davies, 1973; Dott and Bourgeois, 1982). Thus, the texture and composition of the glauconitic siltstone represents two separate processes. These are 1) the "background" sedimentation of silt and clay that accumulated relatively continuously and 2) the rare, high-intensity, but short-lived, storm events that delivered sands onto the inner shelf. Other evidence of storm processes affecting the Blufftown inner shelf is an exposure of the glauconitic siltstone containing an *Exogyra* shell lag lying on an erosional surface. This lag approximates the pattern of a sine curve on two-dimensional outcrop. The *Exogyra* are commonly disarticulated and are imbricated in this allochthonous death assemblage.

Storm Sands: Storm sands of the Blufftown Formation occur in two forms: Hummocky cross-stratified and coquinoid. These forms differ in sedimentary structure and fossil content, but are similar in grain size and thickness. Hummocky cross-stratified sands are composed of micaceous, fine-grained sand and range from 15 to 45 cm in thickness. An idealized sequence through this type of storm sand would have 1) a sharp, erosional base over the glauconitic siltstone, 2) a basal lag with small glauconitic siltstone clasts, 3) a parallel-laminated zone, and 4) a hummocky cross-stratified zone. The top of the sand may be slightly bioturbated. This sequence is then overlain by the glauconitic siltstone. This idealized sequence is similar to those of Dott and Bourgeois (1982) and Walker and others (1983) and, like theirs, is not always complete. In one exposure hummocky cross-stratified sand occurs as a series of widely spaced starved hummocks.

The coquinoid sandstone is similar to the shelf sand facies described by Brenner and Davies (1973). The coquinoid sandstone is 20 to 45 cm thick and is composed of fine- to medium-grained, carbonate-cemented sand which is highly fossiliferous and massively bedded. The sandstone lies on an erosional base over the glauconitic siltstone and fills scours with a fossiliferous lag that encompases nearly the complete thickness of the sandstone. The fossil assemblage is mixed, consisting of both nearshore and deeper water forms. Nearshore forms include *Exogyra* (see previous discussion) and *Turitella* (Scott, 1974). Also commonly present are ostreids, inoceramids, calcareous worm tubes, shark and other teeth, and unidentified bone fragments. The fossils show a crude imbrication, but are not commonly graded according to size or weight. The tops of the beds are commonly rippled and show a lumpy, convex-up parting. When broken, the faint odor of natural gas can be detected escaping from a fresh surface, indicating the coquinoid sandstone is slightly petroliferous. Although the coquinoid sandstone is tightly cemented, moldic porosity created by originally aragonitic forms is common.

These shelf sands of the Blufftown represent rare, short-lived, but high-intensity storms which occurred within the Gulf of Mexico during Blufftown time (Marsaglia and Klein, 1983). During such storms cyclic loading of the barrier-island shoreface by large waves caused liquefaction of the sand and its associated faunal community into a shelf turbidity current (a process described by Walker, 1984) which then accelerated basinward down the shallow

gradient shelf into an area above storm wave base (Figure 5). The shell material and coarser sand were deposited first forming the coquinoid sandstone beds. The sand was so laden with shell material that no thin bedding could easily be created. Only the very tops of the beds became rippled as the wave energy affecting the bottom diminished.

The finer grained sand continued to move further basinward until it too was deposited, becoming reworked into hummocks and swales by the oscillatory flow of waves acting on the bottom. This resulted in the creation of the hummocky cross-stratified sand beds. Therefore, sand grain size and the presence or lack of shell material within the sand determined whether sedimentary structures were or were not formed in the Blufftown-shelf storm sands.

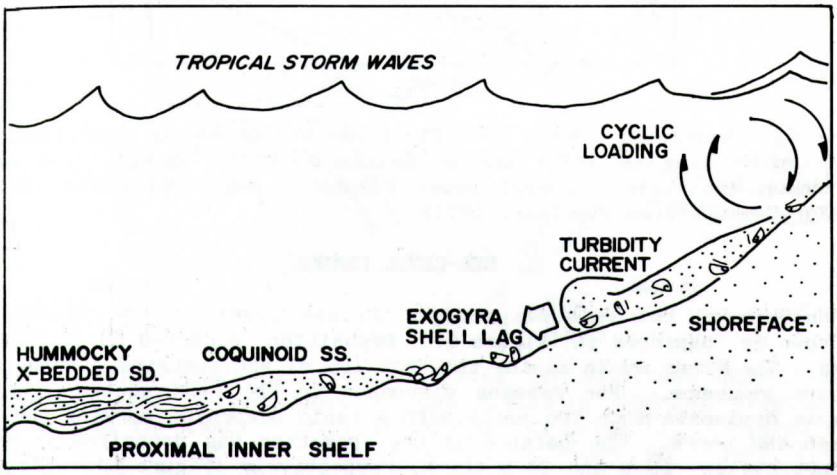


Figure 5. Generation of turbidity current and resulting shell lag, conquinoid sandstone, and hummocky cross-bedded sands. (See text for discussion.)

Mid-Shelf Ensemble Sediments

This ensemble is composed of a number of fine-grained facies described by King (this volume). Because of their textural similarity and the lack of significant surface exposure of the facies within the study area, the midshelf ensemble will be treated in a single description.

Mid-Shelf: These shelf deposits are olive-brown to tan and are composed of a calcareous clay that is slightly sandy. The carbonate content consists mainly of calcareous nannoplankton and foraminifera. Primary sedimentary structures are not abundant and bioturbation is generally complete indicating high rates of infaunal activity and a relatively low sedimentation rate. Thin planar laminations are the only primary sedimentary structures present, indicating pelagic deposition of clay and carbonate particles. Macrofossils are rare, but were present consist of small, disarticulated, broken fragments of *Exogyra*. The heavy-shelled *Exogyra* are not a soft-substrate form (Stanley, 1970) that would normally populate such a fine-grained facies. This fact and their fragmented nature suggests they were transported from the nearshore environment by storm processes. Any thin sand beds representing the associated finer grained storm deposits were biologically mixed into the clayey sediment. Thus, punctuated mixing and bioturbation created the texture of this facies. The physical relationship between the calcareous clay and other facies is not shown in surface exposures, but on the composite correlation (Figure 3) the clay overlies lower-shoreface sands and interfingers shoreward with the proximal inner-shelf glauconitic siltstone facies.

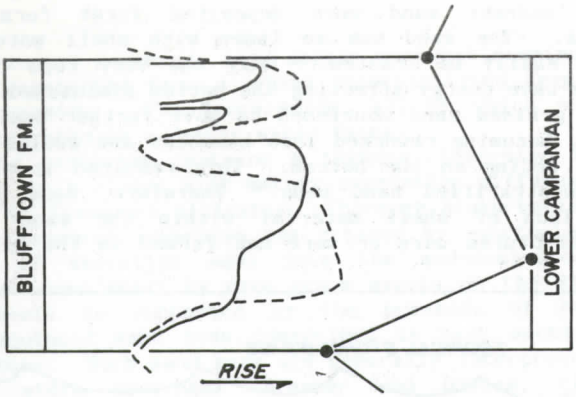


Figure 6. Sea-level curves for Blufftown Formation as described in this paper (solid line on left) and by Reinhardt (1980, dashed line on left). Also shown (on right), coeval Lower Campanian sea-level curve for Western Interior Seaway (from Kauffman, 1977).

SEA-LEVEL CHANGE

Analysis of the relative changes in sea level in the history of the Blufftown is depicted in Figure 6. Deposition occurred during three main cycles. The first cycle is the thickest (95 m) and contains a transgressive-dominant sequence. For reasons discussed in the next section, this first cycle is divisible into two parts with a rapid deepening marking the boundary between the parts. The balance of the formation was deposited in two other distinct cycles, 15 m and 20 m thick, respectively (Figure 6). These cycles are separated by distinct facies discontinuities described in the next section.

Reinhardt (1980) presented a sea-level curve for the Blufftown (also shown in Figure 6). His curve depicts two cycles of relative sea-level rise and fall. Because our work is based on more outcrops and a detailed correlation (Skotnicki, 1985), it is believed that Reinhardt's upper cycle is divisible into two cycles.

Genetic Packages

Correlation of Blufftown Formation facies ensembles has delineated four laterally continuous genetic packages of sediments (boundaries are indicated in Figure 3). Boundaries between these four genetic packages are facies discontinuities which represent rapid, relative changes in sea-level. For example, at the boundary between genetic package I and genetic package II, inner and mid-shelf facies overlie barrier-island sands without lateral interfingering or transition beds.

The vertical and lateral distribution of facies within each of the genetic packages and paleocurrent data suggest that average shoreline position and strike are different for each genetic package. Figure 7 depicts the paleogeographic reconstruction for each genetic package.

The first cycle deposited the facies of both genetic packages I and II. Deposition of genetic package I began with rapid stepwise transgression of the barrier system over an associated back-barrier facies ensemble (approximately 25 m thick). Rampino and Sanders (1980) and Franks (1980) have suggested that a thick sequence of back-barrier sediments, as is present in the Blufftown, can only be preserved in this way in a transgressive sequence. After the step-wise shoreward movement of the barrier system the rate of relative sea-level rise slowed and transgressive sequence of barrier sands (approximately 45 m thick) was deposited over much of the study area

(Figure 7).

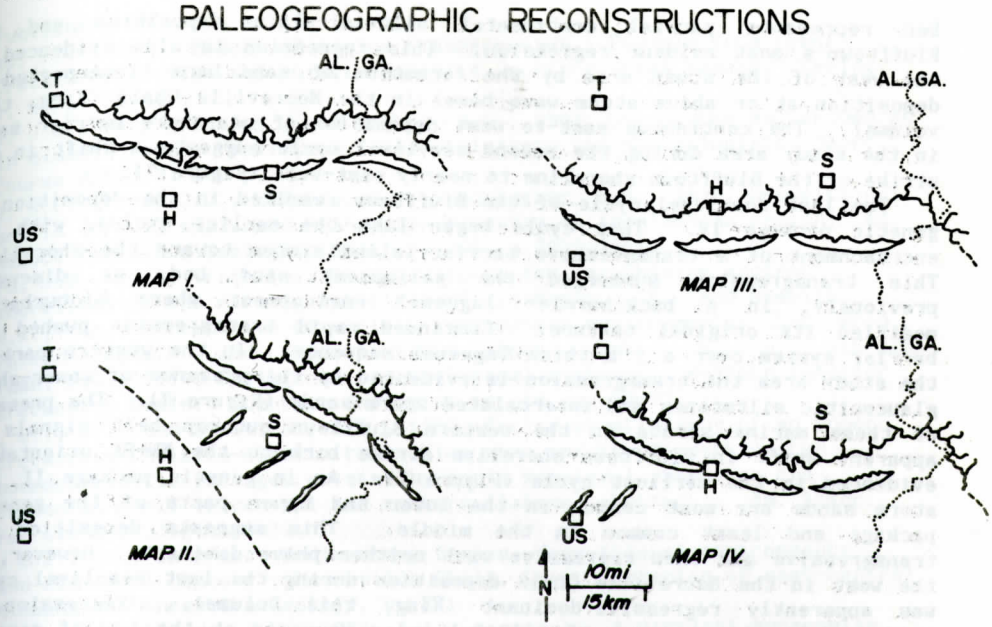


Figure 7. Paleogeographic reconstructions for Blufftown deposition. Maps I, II, III, and IV correspond to the similarly numbered genetic packages (described in the text). Locations of the key towns (T, Tuskegee; US, Union Springs; H, Hurtsboro; S, Seale) are also shown on Figure 2(A). For reference, the Alabama-Georgia border (dotted) is shown on each map above.

Deposition of genetic package II began with a rapid increase in the rate of relative sea-level rise and the resulting development of a facies discontinuity between underlying barrier-island sands and the overlying inner-shelf and mid-shelf facies (Figure 3). This occurred without the lateral interfingering of barrier sands and inner-shelf facies that likely would have developed given slow rates of transgression.

Hummocky cross-bedded storm sands (where present, an indicator of deposition above storm wave-base) are found within the inner-shelf facies near the base of genetic package II. Their absence above the occurrence at the base of genetic package II suggests water depth reached a maximum at this point in the relative sea-level rise. This deepening is represented to the west of the study area by the presence of nearly pure chalks at the same stratigraphic level within the Mooreville Chalk (King, this volume).

The presence of storm sands near the top of genetic package II suggests the beginning of a regressive phase. This regression is also evident to the west of the study area in the Mooreville Chalk where shelf sand bars containing hummocky cross-stratification (an indicator of conditions at or above storm wave base) are present (King, this volume).

The second sea-level cycle of the Blufftown developed genetic package III, which composes a lesser thickness of strata than the other genetic packages (Figure 3). Deposition began like genetic package I, with the stepwise transgression of a barrier system over an approximately 5 m thick back-barrier facies. Transgression, however, was not nearly so great as in the upper part of the first cycle, and inner-shelf marine sediments are only present in the extreme western part of the study area. Regression at the end of the deposition of genetic package III is marked by the diastemic erosional surface between an underlying barrier facies and the overlying palimpsest bed present in the eastern study area. As discussed previously, the palimpsest

bed represents partial continental alluvial-plain deposition and the Blufftown's most evident regression. This regression is also evidenced to the west of the study area by the formation of sand bars (interpreted as deposition at or above storm wave base) in the Mooreville Chalk (King, this volume). The continuous east-to west deposition of nearshore barrier sands in the study area during the second sea-level cycle suggests a shift in the strike of the Blufftown shoreline to nearly east-west (Figure 7).

The last sea-level cycle of the Blufftown resulted in the deposition of genetic package IV. This cycle began like the earlier cycles, with the encroachment of a transgressive barrier-island system toward the shoreline. This transgression submerged the palimpsest sand bed, as discussed previously, in a back-barrier lagoonal environment where bioturbation modified its original texture. Continued rapid transgression pushed the barrier system over a 5 m thick lagoonal sequence. In the western part of the study area the transgression is evidenced by the presence of inner-shelf glauconitic siltstones and intercalated storm sands (Figure 7). The presence of these marine strata in the western Blufftown outcrop belt signals an apparent shift in Blufftown shoreline strike back to the NW-SE orientation evidenced in the earliest cycle (Figure 7). As in genetic package II, the storm sands are most common in the lower and upper parts of the genetic package and least common in the middle. This suggests deposition was transgressive and then regressive with neither phase dominant. However, to the west in the Mooreville Chalk deposition during the last sea-level cycle was apparently regressive-dominant (King, this volume). Regression is evidenced by the presence of a back-barrier sequence at the top of genetic package IV in the eastern part of the study area (Figure 3). This final regression marks the end of Blufftown Formation deposition because a significant stratigraphic break occurs at the top of the formation (Sohl and Smith, 1980).

Because little information on Campanian sea-level cycle at the scale of this study exists in the literature, it is difficult to determine whether control of sea-level during Blufftown deposition was local or eustatic. However, Kauffman (1977) described only one cycle of sea-level rise and fall during the same time span as that of Blufftown deposition. The more complex sea-level history of the northern Gulf rim (Figure 6) may suggest, along with apparent shifts in shoreline strike, that local basin tectonics played a greater role than eustatic sea-level change on the depositional history of the Blufftown Formation.

ACKNOWLEDGEMENTS

This research was supported in part by funding from Sigma Xi.

REFERENCES

Brenner, R.L., and Davies, D.K., 1973, Storm-generated coquinoid sandstone: Genesis of high energy marine sediments form the Upper Jurassic of Wyoming and Montana: Geol. Soc. America Bull., v. 84, p. 1685-1698.

Copeland, C.W., 1968, Facies changes in the Selma Group in central and eastern Alabama, *in* Scott, J.C., chm., Facies changes in the Selma Group in central and eastern Alabama: Alabama Geol. Soc. Guidebook, 4th Annual Field Trip, p. 2-26.

Davies, D.K., Ethridge, F.G., and Berg, R.R., 1971, Recognition of barrier environments: Am. Assoc. Petroleum Geologists Bull., v. 55, p. 550-565.

Dott, R.H., and Bourgeois, J., 1982, Hummocky stratification: Significance of its variable bedding sequences: Geol. Soc. America Bull., v. 93, p. 663-680.

Eargle, D.H., 1950, Geologic Map of the Selma Group in eastern Alabama: U.S. Geol. Survey Oil and Gas Inv. Prelim. Map 105.

Folk, R.L., 1980, Petrology of sedimentary rocks: Austin, Texas, Hemphill

Publishing Co., 184 p.
Franks, P.C., 1980, Models of marine transgression - Example from Lower Cretaceous fluvial and paralic deposits, north-central Kansas: Geology, v. 8, p. 56-61.
Harms, J.C., Southard, J.B., and Walker, R.G., 1982, Structures and sequences in clastic rocks: Soc. Econ. Paleontologists and Mineralogists, Short Course Notes 9, 249 p.
Horne, J.C., Ferm, J.C., Carucio, F.T., and Baganz, B.P., 1978, Depositional models in coal exploration and mine planning in Appalachian region: Am. Assoc. Petroleum Geologist Bull., v. 62, p 2379-2411.
Howard, J.D., 1972, Trace Fossils as criteria for recognizing ancient shorelines in the stratigraphic record, *in* Rigby, J.K., and Hamblin, W.K., eds., Recognition of ancient sedimentary environments: Soc. Econ. Paleontologists and Mineralogists Spec. Pub. 16, p. 215-225.
Kauffman, E.G., 1977, Geological and biological overview: Western Interior Cretaceous basin: Mountain Geologist, v. 14, p. 75-99.
King, D.T, Jr., and Skotnicki, M.C., 1986, Facies stratigraphy of an Upper Cretaceous chalk-to-clastic transition zone and the discontinuity-bounded genetic packages created by rapid sea-level change on the Campanian shelf of central and eastern Alabama (abstract): Am. Assoc. Petroleum Geologist Bull.
King, D.T., Jr., in press, Sedimentary facies, depositional environments, and sea-level history - Mooreville Chalk, Lower Campanian of east-central Alabama: Southeastern Geology, v. 27.
Kumar, N., and Sanders, J.E., 1974, Inlet sequence: A vertical succession of sedimentary structures and textures created by the lateral migration of tidal inlets: Sedimentology, v. 21, p. 491-532.
Lerand, M., 1983, The Blood Reserve Sandstone at the Jenson Reservoir, *in* Lerand, M., ed., The sedimentology of the Blood Reserve Sandstone in southern Alberta: Canadian Soc. Petroleum Geologists, Fieldtrip Guidebook, p. 13-43.
McGowen, M. K., and Lopez, C.M., 1983, Depositional systems in the Nocatoch Formation (Upper Cretaceous), northeast Texas and southwest Arkansas: Texas Bur. of Econ. Geology Rept. of Inv. 137, 55p.
Marsaglia, K.M., and Klein, G.D., 1983, The paleogeography of Paleozoic and Mesozoic storm depositional systems: Jour. Geology, v. 91, p. 117-142.
Monroe, W.H., 1941, Notes on deposits of Selma and Ripley age in Alabama: Alabama Geol. Survey Bull. 48, 150 p.
Monroe, W.H., 1947, Stratigraphy of outcropping Cretaceous beds of southeastern states: Am. Assoc. Petroleum Geologists Bull., v. 31, p. 1817-1824.
Mount, J.F., 1984, Mixing of siliciclastic and carbonate sediments in shallow shelf environments: Geology, v. 12, p. 432-435.
Nikravesh, R., 1967, The foraminifera and paleoecology of the Blufftown Formation (Upper Cretaceous) of Georgia and eastern Alabama: unpublished Ph.D. dissertation, Louisiana State Univ., 159 p.
Rampino, M.R., and Sanders, J.E., 1980, Holocene transgression in south-central Long Island, New York: Jour. Sed. Petrology, v. 50, p. 1063-1080.
Reinhardt, J., 1980, Upper Cretaceous stratigraphy and depositional environments, *in* Frey, R.W., ed., Excursions in southeastern geology, v. II: Geol. Soc. America and American Geol. Inst., Guidebook, p. 386-392.
Schwartz, R.K., 1982, Bedforms and stratification characteristics of some modern small-scale washover sand bodies: Sedimentology, v. 29, p. 835-849.
Schwartz, R.K., Hobson, R., and Musialowski, F.R., 1981, Subsurface facies of a modern barrier island shoreface and relationship to the active nearshore profile: Northeastern Geology, v. 3, p. 283-296.
Schwimmer, D.R., 1981, A distinctive Upper Cretaceous fauna, 3-4 meters

below the Blufftown-Cusseta contact in the Chattahoochee River valley, *in* Reinhardt, J., and Gibson, T.G., Upper Cretaceous and Lower Tertiary geology of the Chattahoochee River valley, western Georgia and eastern Alabama: Georgia Geol. Soc., Fieldtrip Guidebook, p. 81-88.

Scott, J.C., 1960, Ground-water resources of Macon County, Alabama, a reconnaissance: Alabama Geol. Survey Inf. Series 16, 97 p.

Scott, J.C., 1962, Ground-water resources of Bullock County, Alabama, a reconnaissance: Alabama Geol. Survey Inf. Series 29, 120 p.

Scott, J.C., 1964, Ground-water resources of Russessl County, Alabama, a reconnaissance: Alabama Geol. Survey, Bull., v. 75, 77 p.

Scott, R.W., 1974, Bay and shoreface benthic communities in the lower Crestaceous: Lethaia, v. 7, 315-330.

Selley, R.C., 1978, Ancient sedimentary environments (2nd ed.): Ithaca, N.Y., Cornell University Press, 287 p.

Skotnicki, M.C., 1985, Stratigraphy and depositional environments of the Upper Cretaceous Blufftown Formation, eastern Gulf Coastal Plain of Alabama: unpublished M.S. thesis, Auburn Univ., 162 p.

Sohl, N.F., and Smith, C.C., 1980, Notes on Cretaceous biostratigraphy, *in* Frey, R.W., ed., Excursions in southeastern geology, v. II: Geol. Soc. America and American Geol. Inst., Guidebook, p. 382-402.

Stanley, S.M., 1970, Relation of shell form to life habits of the Bivalvia (Mollusca): Geol. Soc. America Mem. 125, 296 p.

Swift, D.J.P., 1968, Coastal erosion and transgressive stratigraphy: Jour. Geology, v. 76, p. 444-456.

Walker, R.G., 1984, Shelf and shallow marine sands, *in* Walker, R.G., ed., Facies models: Geoscience Canada, Reprint Series 1, p. 141-170.

Walker, R.G., Duke, W.L., and Leckie, D.A., 1983, Hummocky stratification: Significance of its variable bedding sequences: Discussion: Geol. Soc. America Bull., v. 94. p. 1245-1249.

APPENDIX 1

TYPE LOCALITIES OF BLUFFTOWN FORMATION DEPOSITIONAL FACIES

Barrier-Island Ensemble Sands

Lower Shoreface:

In Macon County on the north side of Macon County Hwy. 45, 0.1 mile east of intersection with U.S. Hwy. 29. (T 15N, R 24E; SW1/4, SW1/4, Sec. 7)

In Bullock County on the south side of Ala Hwy. 26, 1.0 mile south of the intersection with Bullock County Hwy. 103. (T 14N, R 25E; NW1/4, NE1/4, Sec. 23)

Middle Shoreface:

In Russell County on the west side of Russel County Hwy. 43, 3.9 miles north of intersection with Alabama Hwy. 165. (T 14N, R 29E; SW1/4, SW1/4, Sec. 1)

In Russell County on the north side of Russell County Hwy. 14, 1.0 mile west of intersection with Russell County Hwy. 9. (T 15N, R 26E; NE1/4, SE1/4, Sec. 23)

Tidal Inlet:

In Russell County on the east side of Russell County Hwy. 65, 0.6 mile south of intersection with Russell County Hwy. 22. (T 16N, R 27E; NE 1/4, SW1/4, Sec. 25) (The tidal inlet facies truncates a middle shoreface sand in this exposure.)

In Russell County on the east side of Russell County Hwy. 39, 1.3 miles north of intersection with Russell County Hwy. 24. (T 15N, R 30E; SE1/4,

SE1/4, Sec. 31) (The exposure is highly weathered and must be thoroughly scraped in order to see characteristic herringbone trough cross-bedding.)

Back-Barrier Ensemble Sediments

Marsh:

In Russell County on both sides of Russell County Hwy. 65, 0.75 mile north of intersection with Russell County Hwy. 22. (T 16N, R 27E; NE 1/4, NW1/4, Sec. 23) (Intercalated within the exposure are thin washover sands.)

In Russell County at Alabama Kraft Paper Company railroad cut on the north side of entrance road # 2, 0.4 miles east of intersection with Alabama Hwy. 165. (T 14N, R 30E; E1/2, SW1/4, Sec. 29)

Washover Fan:

In Russell County on the west side of Russell County Hwy. 7, 0.1 mile north of intersection with Russell County Hwy. 2. (T 15N, R 27E; SE1/4, NW1/4, Sec. 4) (The washover fan facies overlies marsh facies.)

In Russell County on the south side of Russell County Hwy. 7, 2.4 miles west of intersection with Russell County Hwy. 22. (T 16N, R 27E; NE1/4, NW1/4, Sec. 19) (The washover fan facies overlies marsh facies.)

Palimpsest Bed:

In Russell County at Alabama Kraft Paper Company railroad cut on the north side of entrance road # 2, 0.4 miles east of intersection with Alabama Hwy. 165. (T 14N, R 30E; E1/2, SW1/4, Sec. 29)

Inner-shelf Ensemble Sediments

Inner Shelf:

In Russell County in the banks of Hatchechubbee Creek approximately 0.5 miles east of bridge on dirt road approximately 4.0 miles south of Colbert on Alabama Hwy. 26. (T 14N, R 28E; NE1/4, NW1/4, Sec. 10) (The exposure shows an erosional surface filled by a shell and teeth lag.)

In Russell County in the banks of Hatchechubbee Creek, 0.5 mile south of Pittsview under bridge on Hwy. 431. (T 14N, R 29E; NE1/4, SW1/4, Sec. 19)

Storm Sands:

In Bullock County on both sides of Central of Georgia Railroad cut approximately 0.5 mile north of intersection with Bullock County Hwy. 40 at Peachburg. (T 14N, R 24E; NE1/4, NE1/4, Sec. 27)

In Russell County in the banks of Briar Creek, 1.7 miles east of intersection with Russell County Hwy. 39 under bridge over Russell County Hwy. 4. (T 14N, R 29E; NE1/4, SE1/4, Sec. 28)

Mid-Shelf Ensemble Sediments

Mid-Shelf:

In Macon County on both sides of Macon County Hwy. 99, 0.4 mile north of intersection with Macon County Hwy. 2. (T 15N, R 24E; SE1/4, SE1/4, Sec. 25) (The exposure is a ditch cut going up the hill.)

In Bullock County on west side of Bullock County Hwy. 103, 2.1 mi. north of intersection with Alabama Hwy. 26. (T 14N, R 25E; SW1/4, NE1/4, Sec. 2)

REGIONAL STRATIGRAPHY AND SEDIMENTOLOGY OF THE LOWER MISSISSIPPIAN ROCKWELL FORMATION AND PURSLANE SANDSTONE BASED ON THE NEW SIDELING HILL ROAD CUT, MARYLAND

THOMAS W. BJERSTEDT

Department of Geology and Geography
West Virginia University
Morgantown WV 26506

ABSTRACT

A new road cut exposes 247 m (810 ft.) of the Lower Mississippian Rockwell Formation and Purslane Sandstone in a spectacular exposure through the synclinal mountain of Sideling Hill, Maryland. This new exposure provides a valuable link between Early Mississippian marine and terrestrial facies in the foreland basin of West Virginia, Pennsylvania and Maryland.

The lower Rockwell Formation records sedimentation in barred-bay, back-bar lagoon and estuarine depositional systems that include a unique interbedded sandstone-diamictite lithofacies. Lagoon subenvironments contain thin coaly shales and restricted faunas in black to dark grey lithofacies that record bar-washover, marsh and probable flood tidal-delta systems. This dark-colored facies suite is the poorly understood Riddlesburg Shale that has been recognized in older Sideling Hill exposures in Pennsylvania and in the Broad Top Basin area.

The upper Rockwell Formation was deposited in alluvial plain environments that included suspension-load fluvial systems, crevasse splays, overbank siltstones, thin coaly shales, incipient soil horizons, and red mudstones. A red mudstone lithofacies at the top of the Rockwell Formation is recognized as the Patton Shale of central Pennsylvania. The overlying Purslane Sandstone records braidplain sedimentation in bed-load fluvial systems.

Polymictic diamictites at the base of the Rockwell Formation at Sideling Hill and the Spechty Kopf Formation in the Anthracite region are essentially chronostratigraphic equivalents of the Riddlesburg Shale in Pennsylvania and were deposited during the same major Early Mississippian sea level rise. In central Pennsylvania and Maryland, the Riddlesburg Shale is part of a diachronous lithosome that records a range of stenohaline and euryhaline environments in a nearshore complex of estuarine, bay and restricted lagoon facies. Riddlesburg Shale deposystems are nearshore equivalents of the basinal Sunbury Shale and were deposited along an environmental gradient from the basin-plain in Kentucky and Ohio (Sunbury) through shallow-shelf and nearshore environments in West Virginia (Riddlesburg), and restricted back-bar lagoon and marsh environments in south-central Pennsylvania and western Maryland (Riddlesburg).

INTRODUCTION

A spectacular new road cut has been excavated through the synclinal mountain of Sideling Hill along U.S. Route 40 in Washington County, Maryland. This new exposure is 10.2 km (6.4 mi.) west of the Rt. 40 and I-70 interchange at Hancock and has been accessible to highway traffic since August, 1985. The cut exposes almost all of the Lower Mississippian Rockwell Formation and Purslane Sandstone in 247 m (810 ft.) of stratigraphic section. This new Sideling Hill exposure provides a valuable link between Devonian-Mississippian marine facies in the foreland basin and terrestrial equivalents deposited near the ancestral Acadian Mountains.

Stose and Swartz (1912) named the Rockwell Formation for exposures along Rockwell Run, a small Potomac River tributary flowing northward between Sideling Hill and Purslane Mountain in Morgan County, West Virginia (Figure 1). The Rockwell was described as a course, arkosic sandstone to conglomerate with buff shales and darker shales containing thin coaly beds. At that time, the best exposures occurred in the Potomac River gap through

Sideling Hill and in Rockwell Run gap. Stose and Swartz (1912) named the Purslane Sandstone for white, conglomeratic sandstone overlying the Rockwell Formation. The Purslane Sandstone is exposed in the Meadow Branch syncline in Morgan and Berkeley counties and forms caps on regional synclinal mountains at Spring Gap Mountain, Town Hill and Sideling Hill (Figure 1). Stose and Swartz (1912) named the Hedges Shale for coal-bearing shale overlying the Purslane Sandstone near Hedges Mountain, in the northern part of the Meadow Branch syncline (Figure 1). The Hedges Shale is exposed only in the Meadow Branch syncline, however, and is not in the sequence exposed at Sideling Hill. Grimsley (1916) largely reviewed Stose and Swartz's (1912) earlier work in the West Virginia county report for Jefferson, Berkeley and Morgan counties. Since these early studies, little detailed work has been carried out on these formations because exposures are generally poor. In light of the accessibility, completeness and quality of the new Sideling Hill section, it is designated as the principal reference section for the Rockwell Formation and Purslane Sandstone in the panhandle regions of western Maryland and eastern West Virginia.

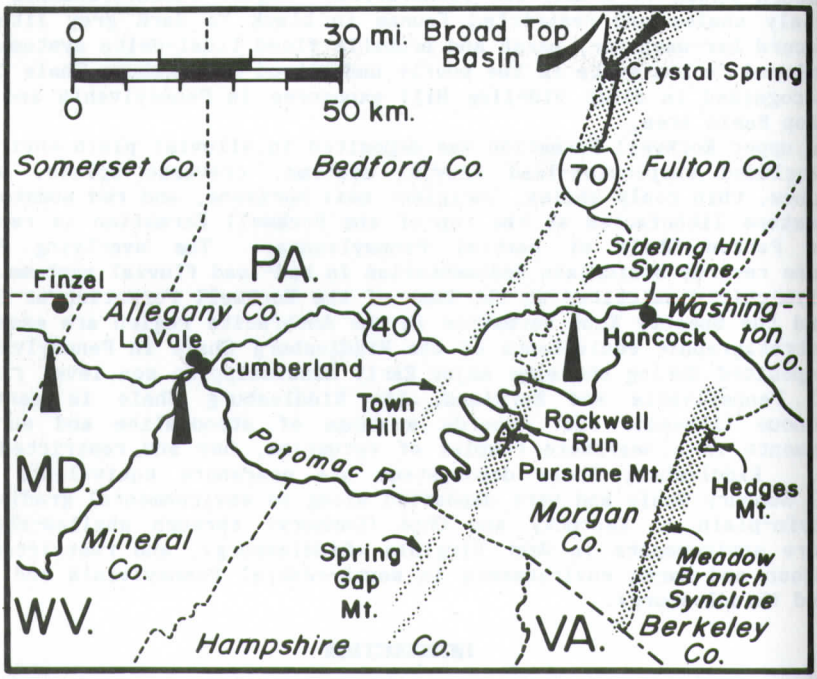


Figure 1. Index map showing the panhandle regions of western Maryland and eastern West Virginia. Synclinal mountains capped by Purslane Sandstone are stippled and original type sections for the stratigraphic units described by Stose and Swartz (1912) are shown. Large arrows show the locations of stratigraphic sections and place names discussed in text.

This paper presents a measured and described section of this new exposure. Following discussions concern regional Upper Devonian-Lower Mississippian stratigraphic relationships in order to place the Sideling Hill exposure into its proper stratigraphic and sedimentologic context.

SIDLING HILL MEASURED SECTION

The Sideling Hill exposure was measured under sunny conditions, on July 24, 1985 (Figure 2). A steel tape was held normal to dip for measuring the

thickness of field units. The thickness of unit 24 was determined by pace/compass technique. Total section thickness will vary according to the syncline limb chosen as the base of the section. The stratigraphically lowest lithologies are exposed on the western limb of the syncline. Field units were measured mostly along the north cut face that receives the most sunlight, or where best exposed. The broken section symbol in Figure 3 indicates that a change took place in the location where the section was measured.

Field units 1-17 were measured along the highway level. Units 1-3 were measured along the western limb of the south cut face, and 4-17 on the eastern limb of the north cut face. Units 18 and 19 were measured along the first bench of the east limb north cut, 20 and 21 along the second bench of the east limb north cut and units 22-24 along the third bench and summit of the west limb north cut. The section begins just below the polymictic diamictite, which is exposed only on the west limb of the syncline. Rock colors were assigned according to the Geological Society of America rock color chart (Goddard and others, 1975). Fresh and weathered colors are abbreviated with "F" and "W" respectively. Qualitative bedding thickness scale is from Ingram (1954). In unit descriptions, small-scale refers to structures that can be seen in hand samples or slabs; medium-scale refers to structures that can be recognized from the highway shoulder; and large-scale refers to structures that can be recognized at the scale of the entire exposure. Field units are shown graphically in a columnar section (Figure 3) and are described in Table 1.

VERTICAL SEQUENCE

Rockwell Formation

The uppermost Hampshire Formation is poorly exposed at Sideling Hill and it is exposed only on the southern cut face of the western syncline limb about 40 m (130 ft.) below a basal Rockwell diamictite lithofacies. The Rockwell Formation is 191 m (627 ft.) thick and almost completely exposed at Sideling Hill. It consists of a sequence of fine- to coarse-grained sandstones, siltstones and mudstones. The lower Rockwell is characterized by flaky and crumbly centimeter-to decimeter-thick coaly shales that are not found higher than 120 m (394 ft.) above the base of the measured section (unit 14). The upper 71 m (233 ft.) of the Rockwell Formation contains red mudstones. As will be subsequently discussed, black to dark grey lithofacies in the lower Rockwell contain evidence of deposition in marginal-marine environments.

Stose and Swartz (1912, p. 13) described Rockwell sandstones as arkosic, a description perpetuated by later authors (Dally, 1956, p. 48). This was apparently a megascopic description because it has no basis petrographically. Dennison and Wheeler (1975, p. 79) pointed out that Pelletier's (1958) petrographic determinations of 14 "Pocono" thin section samples from Maryland and Pennsylvania (mostly lithic arenites) did not agree with Stose and Swartz's description. Eight sandstone samples from the Sideling Hill section were examined petrographically. All samples fall within the sublitharenite field in Folk's (1968) classification as modified by Folk and others (1970) (Figure 4). In fact, virtually no feldspar was identified in these sandstones. The Huntley Mountain Formation of north central Pennsylvania, a lateral equivalent of the Rockwell, similiarly contains little or no feldspar (Berg and Edmunds, 1979, Figure 14).

A 23 m (75 ft.) polymictic diamictite and interbedded fine- to medium-grained sandstone (unit 2) occurs near the base of the Rockwell Formation on an erosional surface at the top of unit 1. The diamictite is completely unbedded, very poorly sorted and contains large-scale dewatering and diapiric structures and large-scale flow rolls. Quartz pebbles and pebbly-to-cobbly rock fragments of siltstone, sandstone, coaly siltstone, chert, rhyolite and

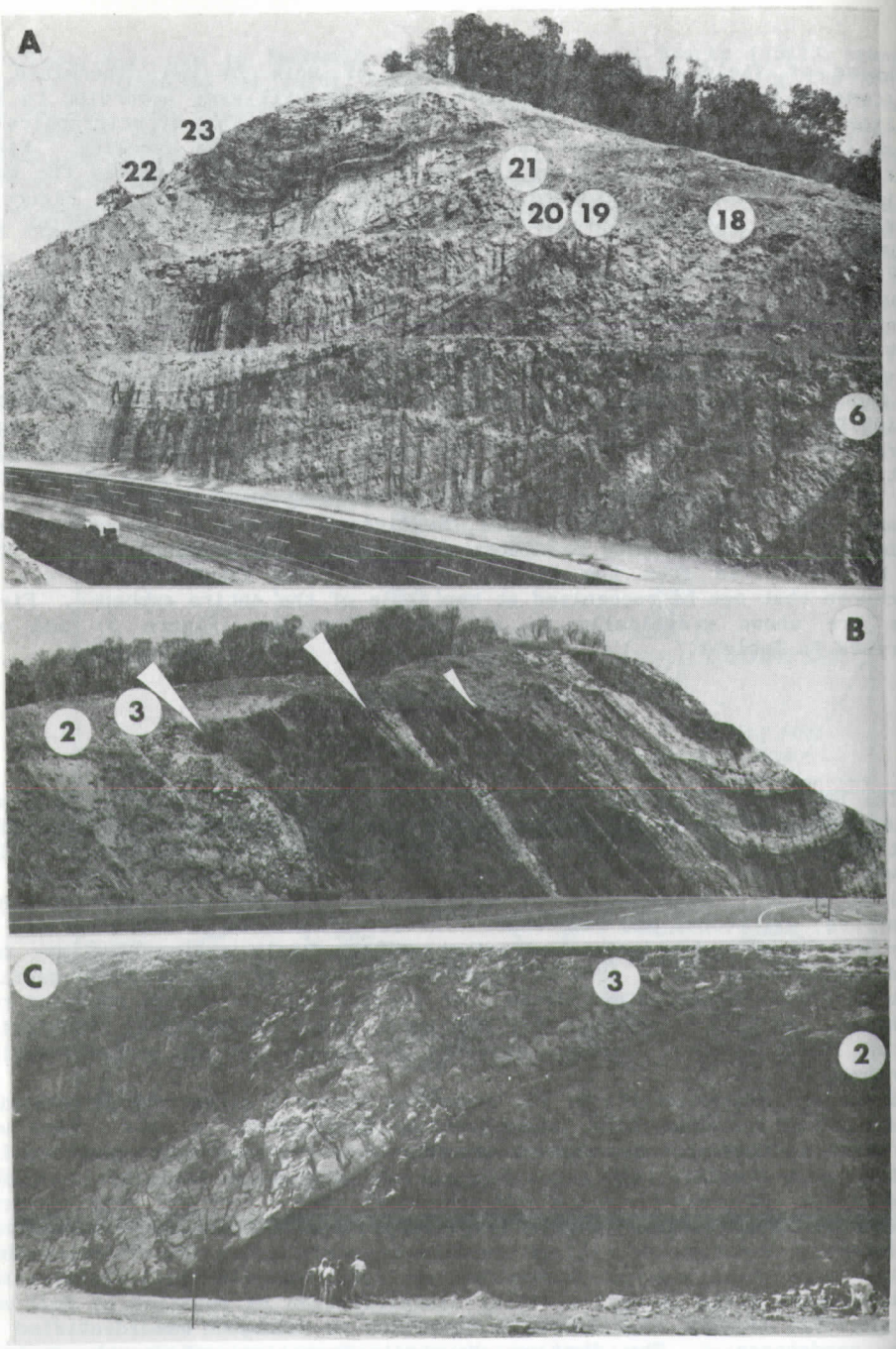


Figure 2. A. North cut face of the eastern limb of the Sideling Hill exposure showing field units from the measured section in Figure 3. B. North cut face of the western syncline limb showing the stratigraphically lowest units including massive polymictic diamictite (unit 2) that overlies a thick basal sandstone. An overlying tidal inlet-fill sandstone lithofacies is recognized (unit 3), and the Riddlesburg Shale Member (units 4 - 5 of the

assorted metamorphic lithologic suites occur in a mud-supported, very sandy matrix. Granitic and gneissic(?) rock fragments are less abundant. The diamictite records subaqueous mud and debris flows in estuarine and barred-bay environments.

A fining-upward, medium- to coarse-grained sandstone 13.1 m (43 ft.) thick overlies the diamictite (unit 3). Sedimentary structures in this sandstone lithofacies include large- to medium-scale trough and planar crossbed sets near the base that grade upward to small-scale wedge crossbed sets, bidirectional planar crossbed sets (herringbone) and apparent current-ripple lamination. Vertical *Skolithos* burrows occur in the upper part of the unit. This sandstone lithofacies is interpreted to represent a tidal inlet-fill sequence from a marine-bar complex.

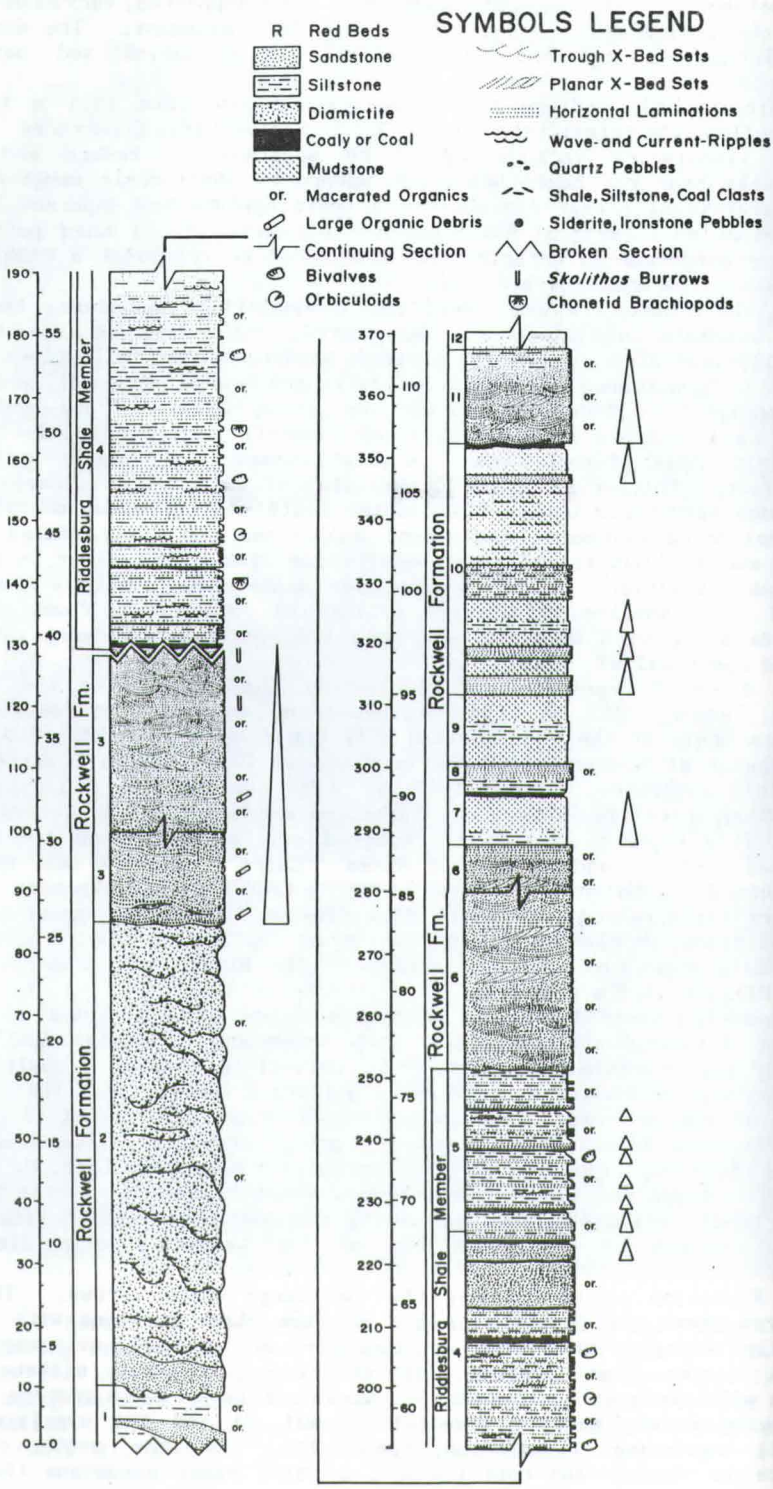
Above the diamictite and overlying channel-fill sandstone, the lower Rockwell consists of black to dark grey, current- and wave-rippled, horizontally-laminated, very fine-grained sandstones and siltstones. This lithology is punctuated with medium- to thick-bedded, tabular and scour-based sandstones. These sandstones are storm washover fans and splays deposited in a back-bar lagoon. They are commonly capped with 5 to 20 cm of coaly shale that formed upon re-establishment of lagoon and marsh sedimentation. Thicker sandstone bodies (top of unit 4), or closely-spaced, thick-bedded sandstones within this facies suite may represent deposition in flood tidal-delta systems. Individual sandstones and coaly shales usually cannot be successfully traced from one syncline limb to the other in the same stratigraphic position. Black to dark grey siltstones contain a restricted, fauna of thin-shelled bivalves, orbiculoid brachiopods and chonetid brachiopods along with abundant macerated and larger plant debris including *Adiantites spectabilis*? Read.

Units 4 and 5 represent the Riddlesburg Shale Member of the Rockwell Formation. Reger (1927, p. 404) discussed the occurrence of fossil-bearing Riddlesburg Shale at the old Sideling Hill tunnel (now covered), 7.5 km (4.5 mi.) northeast of Breezewood, Pennsylvania, and 45 km (28 mi.) north of the new Maryland exposure. Laird (1942, p. 149) reported other localities on Sideling Hill where the Riddlesburg Shale was exposed, 1) near Cassville (now covered), 2) a highway exposure he reported east of Kimmel was the Sideling Hill tunnel section, and 3) west of Hiram. Laird's exposure near Hiram was not relocated. Laird (1942, p. 149) noted that thicknesses of the Riddlesburg Shale ranged from 22 to 33 m (72-108 ft.) in exposures along the Structural Front at Altoona and in the Broad Top Basin, which included the Sideling Hill exposures. The thickness of the Riddlesburg Shale Member at Sideling Hill is 37.3 m (122.5 ft.).

An upsection trend from dark grey, fossil-bearing siltstones in units 4 and 5 to fining-upward sequences with brownish, yellowish and reddish siltstones and mudstones in units 7-17 reflect a gradual transition from subaqueous, brackish-water deposition to subaerial deposition. The change to subaerial deposition seems to be characterized by the development of incipient paleosol horizons comprised of glaebule-bearing, crumbly, yellowish mudstones up to 1 m thick with apparent gilgai structures (Collinson, 1978, p. 41). An incursion of brackish, lagoonal environments onto subaerial facies is shown by a 1 m thick, black mudstone containing abundant thin-shelled bivalves (in unit 10, exposed in the north cut of the western syncline limb). The

---

Rockwell Formation occurs between the two large white arrows. The small white arrow shows the position of a 1 m thick black mudstone with abundant thin-shelled bivalves that marks an incursion of lagoonal environments onto subaerial facies that include lighter-colored, crumbly siltstones and mudstones with incipient paleosols. C. South cut face of the western syncline limb showing thick, massive diamictite (unit 2) and the overlying tidal inlet-fill sandstone lithofacies (unit 3). On the south cut, the diamictite is thicker and contains a 2 m thick basal sandstone that fills the relief on an erosional surface.


SYMBOLS LEGEND
R Red Beds
Sandstone
Siltstone
Diamictite
Coaly or Coal
Mudstone
or. Macerated Organics
Large Organic Debris
Continuing Section
Bivalves
Orbiculoids
Trough X-Bed Sets
Planar X-Bed Sets
Horizontal Laminations
Wave-and Current-Ripples
Quartz Pebbles
Shale, Siltstone, Coal Clasts
Siderite, Ironstone Pebbles
Broken Section
Skolithos Burrows
Chonetid Brachiopods
Riddlesburg Shale Member
Rockwell Fm.
Rockwell Formation

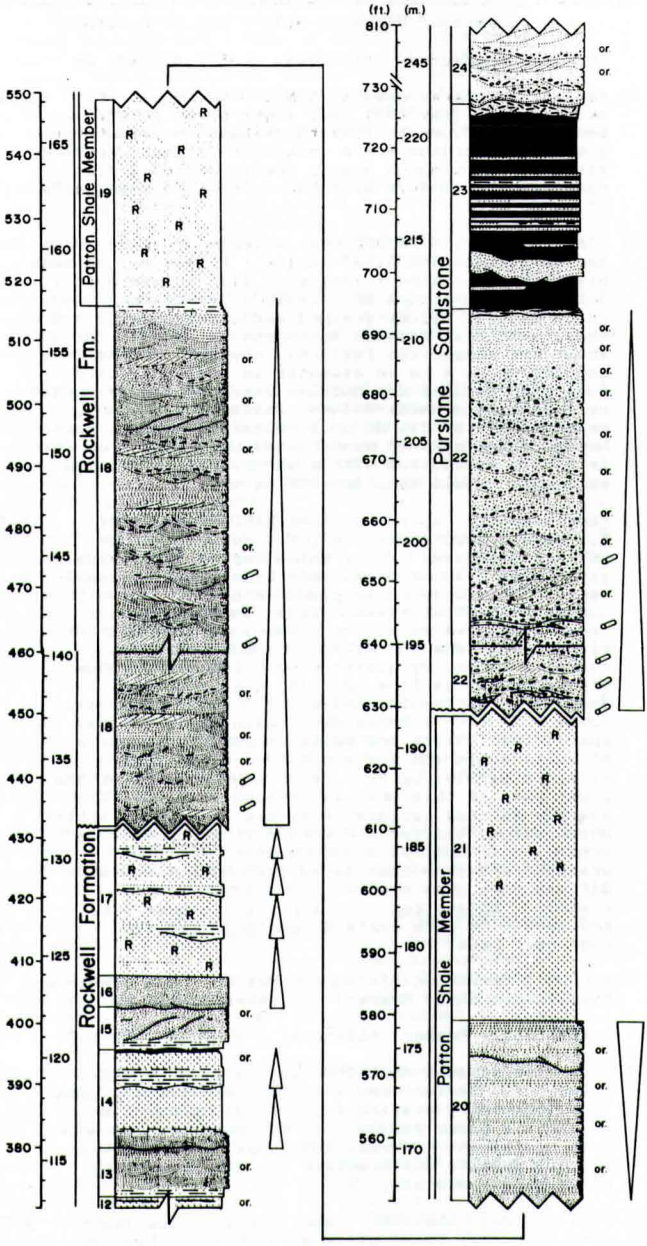


Figure 3. Detailed measured and described section at the new Sideling Hill exposure. Numbers at left refer to field stratigraphic units defined during measurement. Lithologic symbols are defined in the legend. Descriptions of the field units are contained in Table 1.

Table 1. Measured and described section of the Sideling Hill road cut along U.S. Rt. 40 in Washington County, Maryland.

UNIT	DESCRIPTION	THICKNESS (m)	(ft)
	Summit of Sideling Hill - End of Section		
	Purslane Sandstone thickness = 55.7 m: 183 ft		
24	medium- to coarse-grained SANDSTONE: color - F., medium light grey (N6), W., light brown (5YR5/6); bedding - medium- to very thick-bedded; structures - scour base, medium-scale trough and planar crossbed sets, semicontinuous quartz pebble and coal clast conglomerates mostly near base, isolated coal balls in sandstone.	26.2	86
23	COAL and coaly MUDSTONE with interbedded fine- to medium-grained SANDSTONE: color - F. and W., greyish black (N2); bedding - coal-rich lithologies mostly unbedded, sandstones thin-bedded; structures - one, thick-bedded, medium-grained sandstone channel 1.5 m above unit base showing thickness variations due to scour and compaction features, abundant marcasite nodules up to 4 cm in diameter in coal and coaly mudstone, nodules and nodular layers commonly marked by marcasite's white-colored weathering product, melanterite; middle of unit contains tabular, thin-bedded, fine-grained sandstones; upper 0.6 m of unit is highly brecciated with a sandy, coal clast and marcasite nodule conglomerate at top.	10	33
22	conglomeratic, coarse-grained SANDSTONE: color - F., medium light grey (N6), W., moderate brown (5YR3/4); bedding - very thick-bedded to massive; structures - scour base, unit fines upward, small-, medium- and large-scale planar crossbed sets with common tangential bases, large- to medium-scale trough crossbed sets less common; foreset toes of planar crossbeds and axes of trough crossbeds commonly marked by quartz pebble lags, individual planar crossbeds fine upwards, quartz pebbles angular to subangular and up to 3 cm in diameter, lower 5 m of unit contains vitrain streaks (coal spar), coal clasts and balls, pods and stringers of silty shale and coaly mudstone and oblong siderite pebbles up to 7 cm in length; most of the lower part of this unit is striped with very thin sharply-defined laminae of macerated organic matter which become thicker and more diffuse upsection; upper 3 m of unit is a medium grey (N5) fine-grained sandstone that lacks pebbles but contains diffuse dark grey organic bands and streaks; fossils - Adiantites in a silty shale near base and coalified pith casts of Lepidodendropsis logs near base.	19.5	64
	Rockwell Formation (including Patton Shale Member and Riddlesburg Shale Member) thickness = 191 m: 627 ft		
	Patton Shale Member thickness = 33.9 m: 111.5 ft		
21	very fine-grained SANDSTONE and silty MUDSTONE: color - F., medium grey (N5), W., variegated, upper 10 m dominantly greyish red (5Y4/2), lower 5 m dominantly dusky yellow (5Y6/4); bedding - unbedded, unit breaks into curved, conchoidal surfaces with overall texture and fracture pattern resembling block milk chocolate.	15.1	49.5
20	fine-grained SANDSTONE: color - F., medium light grey (N5), W., light olive grey (5Y6/1); bedding - thin- to thick-bedded; structures - scour base; coarsens and thickens upward to thick-bedded, medium-grained sandstone, horizontal laminations, large-scale reactivation surfaces on south cut face color banding in unit from grey to reddish grey.	8.8	29

19	silty and sandy MUDSTONE: color - F., greyish red (5R4/2), W., light brown (5YR6/4); bedding - unbedded, weathers crumbly, silty and sandy zones have a more resistant, blocky weathering profile; structures - none observed.	10	33
18	fine-grained SANDSTONE: color - F., medium grey (N5), W., light brown (5YR5/6); bedding - thin- to very thick-bedded; structures - scour base, unit fines upward, medium-scale lateral accretion bedding; large-scale planar crossbed sets some with tangential bases, medium- to large-scale trough crossbed sets, horizontal laminations, abundant large- to medium-scale internal reactivation surfaces, shale pebbles and minor siderite pebbles on bedding planes, coaly siltstone and siltstone interbeds up to 1 m thick.	25.6	84
17	MUDSTONE with interbedded SILTSTONE and very fine-grained SANDSTONE: mud/silt-sand = 3/1; color - F. and W., greyish red (5R4/2); bedding - unbedded; structures - scour bases on sandstone and siltstone interbeds.	7.3	24
16	very fine-grained SANDSTONE: color - F., medium dark grey (N4), W., moderate yellowish brown (10YR5/4); bedding - massive; structures - scour base on unit.	1.5	5
15	fine-grained SANDSTONE with interbedded SILTSTONE: sand/silt = 3/1; color - F., medium grey (N5); W., light brown (5YR5/6); bedding - medium- to thick-bedded; structures - medium- to large-scale lateral accretion bedding, lower 0.6 m of unit is a siltstone with scour base grading upward to medium-grained sandstone.	2.1	7
14	MUDSTONE with interbedded SILTSTONE and very fine-grained SANDSTONE: mud/silt/sand = 5/1/1; color - F., olive grey (5Y3/2), W., medium grey (N5); bedding - mudstone is unbedded, weathers crumbly; structures - scour base on unit and on a thick siltstone interbed near unit top, lower 0.6 m of unit is very fine-grained sandstone, 15 cm thick coaly shale at top of unit.	4.9	16
13	fine- and very fine-grained SANDSTONE: color - F., medium dark grey (N4), W., light brown (5YR5/6); bedding - medium-bedded; structures - medium-scale trough crossbed sets at the base and horizontal laminations near the top, lower 0.4 m of unit is a siltstone.	2.4	8
12	MUDSTONE: color - F., olive grey (5Y3/2), W., dark yellowish brown (10YR4/2); bedding - unbedded, weathers crumbly; structures - upper 0.5 m contains thin coaly shale lenses.	1.5	5
11	medium-grained SANDSTONE: color - F., medium light grey (N6), W., light brown (5YR5/6); bedding - thick- to very thick-bedded at base, medium-bedded at top; structures - scour base, unit thins and fines upward, medium- and large-scale trough crossbed sets, small-scale planar crossbed sets and horizontal laminations.	4.6	15
10	MUDSTONE with interbedded SILTSTONE and very fine-grained SANDSTONE: mud/silt/sand = 4/2/1; color - F., olive grey (5Y3/2), W., dusky brown (5YR2/2); bedding - unbedded, siltstone and sandstone medium- to thick-bedded; structures entire unit is a series of fining-upward sequences, interbedded sandstones and siltstones have horizontal laminations, a 0.3 m thick coaly siltstone at top; body fossils - a 1 m thick black mudstone with abundant thin-shelled bivalves occurs in this unit on the north cut face of the western syncline limb and thins to near zero in the south cut.	12.4	40.6
9	MUDSTONE with interbedded SILTSTONE: color - F. medium dark grey (N4), W., brownish grey (5YR4/1); bedding - unbedded, weathers crumbly, siltstone	3.4	11

shows a more resistant, blocky weathering profile; structures - 1.2 m thick siltstone in middle of unit.

8	fine-grained SANDSTONE: color - F., medium grey (N5), W., moderate brown (5YR4/4); bedding - one thick bed; structures - sharp upper and gradational lower contact, horizontal lamination.	0.9	3
7	MUDSTONE: color - F., medium dark grey (N4), W., brownish grey (5YR4/1); bedding - unbedded, weathers crumbly; structures - lower 0.8 and upper 0.6 m of unit is more silty with a more resistant, blocky weathering profile, 15 cm coaly shale near unit top.	3.1	10.3
6	medium-grained SANDSTONE: color - F., light grey (N7), W., light brown (5YR5/6); bedding - medium- to thick-bedded; structures - scour base, large-scale trough crossbed sets and planar crossbed sets with tangential bases, horizontal laminations, some interbeds of fine-grained sandstone. South cut face on the eastern syncline limb shows a major siltstone split which results in a reduction in unit thickness to half that shown on north face.	11	36

Riddlesburg Shale Member thickness = 37.3 m: 122.5 ft

5	SILTSTONE with interbedded fine-grained SANDSTONE: silt/sand = 3/1; color - F., dark grey (N3), W., light brown (5YR5/6); bedding - unweathered siltstone is massive, weathering to very thin-bedded slivers and pencils, sandstones are single beds from medium- to thick-bedded; structures - unit is a series of fining-upward sequences from sandstone to siltstone, some sandstones with scoured bases and some sequences capped by thin coaly shale, observations on the north cut of the western syncline limb show a gradation to lighter-colored lithologies toward the unit top with apparent incipient soil horizons (gilgai) up to 1 m thick composed of glaebule-bearing, crumbly yellowish mudstone; body fossils - similiar to unit 4 and mostly in lower part of the unit.	8.1	26.5
4	SILTSTONE with interbedded very fine- and fine-grained SANDSTONE: silt/very fine sand/fine sand = 5/3/1; color - F., dark grey (N3), W., dark yellowish orange (10YR6/6); bedding - unweathered siltstone is massive, weathering to crumbly slivers and pencils, sandstones thin- to thick-bedded; structures - siltstone and sandstone show horizontal laminations, wavy bedding and current- and wave-ripple lamination, pyrite nodules 1 to 3 cm in diameter, less common thin sideritic layers up to 3 cm thick, micaceous bedding planes, 30 cm thick coaly shale at base with numerous flaky, coaly shales from 10 to 20 cm thick throughout unit; trace fossils - Skolithos, indistinct horizontal burrows; body fossils - thin-shelled bivalves, chonetids and orbiculoids, macerated organic matter and larger plant debris are abundant, including Adiantites spectabilis Read.	29.3	96
3	medium- to coarse-grained SANDSTONE: color - F., medium grey (N6) W., light brown (5YR5/6); bedding - thick-bedded to massive; structures - unit fines upward to medium-grained sandstone, large- to medium-scale trough and planar crossbed sets especially near base, horizontal laminations, small-scale wedge crossbed sets and bidirectional planar crossbed sets (herringbone), current-ripple lamination, scale of sedimentary structures generally decreases upsection; thin partings of fine-grained sandstone and siltstone; wavy, internal reactivation surfaces, grey color bands near unit top beneath coaly bed at base of unit 4; trace fossils - Skolithos, uppermost bedding plane is bioturbated.	13.1	43
2	POLYMICTIC DIAMICTITE: color - F., medium dark grey (N4), W., medium grey (N3); bedding - massive, completely unbedded; structures - this unit contains a scoured, erosional base with moderate relief, 2 m of fine-grained sandstone fills the relief on the scour-base and is itself scoured by the diamictite lithofacies; the diamictite is very poorly sorted, pebbles and cobbles up to 30 cm in diameter are	22.9	75

supported in a sandy mudstone matrix, large-scale dewatering and diapiric structures and flow rolls, pebbles and cobbles composed of quartz and rock fragments of sandstone, siltstone, coaly siltstone, chert, rhyolite and assorted metamorphic lithologic suites with minor granitic and gneissic(?) rock fragments; comments - The diamictite lithofacies shows complex facies changes between the north and south cuts. These include marked thickness changes and interbedding with a basal fine- to medium-grained sandstone. The diamictite is thinner on the north cut face (approx. 16 m (52 ft)) and overlies at least 17 m (55 ft) of sandstone, the base of which is not exposed. This sandstone unit contains horizontal laminations, is badly faulted, shows slickensides and contains interbedded diamictite near the upper contact where a 1 m thick lens(?) of reddish mudstone also occurs.

1	silty SHALE and silty MUDSTONE: color - F. and W.; blackish grey (N2); bedding - thin-bedded to laminated, weathers crumbly; structures - moderate relief on a sharp erosional upper contact with the basal sandstone of the diamictite in unit 2. A few apparently channeled, medium-grained sandstones with scour bases are poorly exposed in the road bank of the south cut, entire thickness of unit not measured. Red mudstones of the Hampshire Formation are poorly exposed in the road bank 40 m (130 ft) below the upper contact of unit 1, on the west syncline limb.	3+	10+

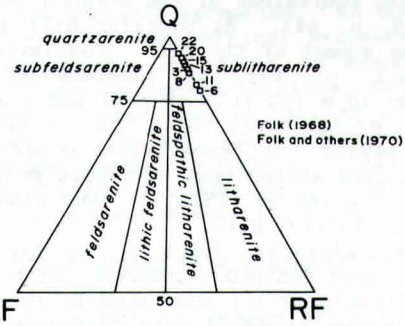


Figure 4. Triangular plot of sandstone composition from eight sandstone samples from the Sideling Hill exposure after Folk (1968) as modified by Folk and others (1970). Numbers refer to field units in the columnar section.

stratigraphic transition from lower Rockwell coaly beds to upper Rockwell red beds reflects an environmental transition from brackish-water and ever-wet conditions in lagoon and marsh facies to well-drained interfluvial areas on the alluvial plane.

Channel or crevasse splay sandstones in the vertical sequence from units 6-18 show a series of fining-upward sequences. Most channel sandstones are fine- to medium-grained but may be coarse-grained as well. Channel sandstones show abundant internal reactivation surfaces and lateral accretion bedding from channel and bedform migration. Sedimentary structures in channel sandstones include medium- to large-scale planar crossbed sets, with or without tangential bases, medium- to large-scale trough crossbed sets and horizontal laminations. Fining-upward sequences, with or without thinning-upward of bedding, are a recurrent pattern in the Rockwell Formation.

Red beds occur only in the upper part of the Rockwell. The first red mudstone lithofacies occurs in field unit 17, 125 m (410 ft.) above the base of the Rockwell. This is the stratigraphically highest field unit that can be examined entirely along the highway. Field unit 18 is a channel sandstone, 25.6 m (84 ft.) thick, that is the thickest individual sandstone unit in the Rockwell Formation. A significant unconformity occurs at the base of this channeled lithofacies.

The 34 m (111.5 ft.) of dominantly red mudstone overlying the unit 18 channel sandstone is the Patton Shale of Campbell (1904, p. 6). He first recognized this sub-Burgoon red shale in Jefferson County, west- central Pennsylvania and reported that its maximum thickness was 24 m (80 ft.) in that area. This red shale was subsequently recognized in other geologic folios prepared by the U.S. Geological Survey in Pennsylvania, (Butts, 1905). Correlation of these red beds with the Patton Shale is consistent with the work of Berg and Edmunds (1979, p. 34) and Craig and Connor (1979, Plate 15A). An upper red mudstone lithofacies (unit 21) has a texture and fracture pattern resembling block milk chocolate and is separated from a lower red mudstone lithofacies by an 8.8 m (29 ft.) thick coarsening- upward, horizontally-laminated sandstone of possible crevasse splay origin. The Patton Shale is not a continuous lithologic unit through the outcrop belt of central Pennsylvania. Berg and Edmunds (1979, p. 43) reported that the Patton is locally missing and that the Burgoon Sandstone overlies the Rockwell and equivalent Huntley Mountain Formation in north-central Pennsylvania. Because of poor exposures or erosional nonpreservation, the Patton cannot be consistently recognized on outcrop and should be relegated to member status in the Rockwell Formation.

Purslane Sandstone

The Purslane Sandstone of the Maryland and West Virginia panhandle region is the lithostratigraphic equivalent of the Burgoon Sandstone of Pennsylvania (Berg and Edmunds, 1979, p. 44). At Sideling Hill it is 55.7 m (183 ft.) thick and extends to the summit of the hill. Consistent with descriptions by Stose and Swartz (1912, p. 13), it consists of two major sandstone lithofacies separated by 10 m (33 ft.) of coal and coaly mudstone (unit 23). Stose and Swartz (1912, p. 14) noted that, "The sandstone caps Sideling Hill from the river north beyond the Pennsylvania state line and exhibits in places the coal-bearing dark shales that have been prospected on the mountain slope above the Potomac." Grimsley (1916, p. 146) also noted this coaly unit in Purslane exposures on Sideling Hill.

The lower sandstone lithofacies (unit 22) is 19.5 m (64 ft.) thick and consists of pebbly, coarse-grained sandstone with local clast-supported conglomerate lenses. Structures are dominated by small- to very large-scale (up to 1 m in height) planar crossbed sets, commonly with tangential bases. Planar crossbed sets (Sp) that dominate this lithofacies can be codified as "Sp" bedsets in Miall's (1978) classification scheme of fluvial deposystems.

A photograph typical of small-scale Sp bedsets from the Purslane Sandstone on Sideling Hill is shown in Pettijohn and Potter (1964, plate 32B). Individual planar beds fine upwards and commonly have quartz pebbles at foreset toes. Trough crossbed sets are less abundant. Quartz pebbles in this lower lithofacies are angular to subangular. The lower 5 m of this unit contains abundant coal balls, coal clasts and vitrain streaks (coal spar), siderite pebbles, black coaly shale and silty shale lenses with *Adiantites* fronds and coalified pith casts of *Lepidodendropsis* log fragments. The lower sandstone lithofacies is striped with very thin, sharp, organic-rich bands that become more diffuse and increase in thickness and number upsection as the unit grades into a fine-grained sandstone beneath the coasly mudstone lithofacies.

The coal and coaly mudstone lithofacies (unit 23) is 10 m (33 ft.) thick and contains interbedded fine- to medium-grained sandstones with locally channeled bases. The fresh color of the coaly mudstone is dull black rather than vitreous and contains abundant marcasite concretions up to 4 cm in diameter. Individual concretions and concretionary layers are marked by the white-colored weathering product of marcasite, melanterite. The top of this lithofacies is marked by a 0.6 m (2 ft.) thick sandy coal breccia and marcasite concretion conglomerate. This brecciated layer may indicate a significant diastem. The upper sandstone lithofacies scours deeply into the

coaly lithofacies that thins from the north cut face to the southern face. On the southern face of the western syncline limb, the coaly facies has been completely scoured-out and the upper sandstone lithofacies locally lies on or scours into the upper meter of the lower sandstone lithofacies.

The upper sandstone lithofacies (unit 24) is 26.2 m (86 ft.) thick and consists dominantly of medium-grain sandstone with abundant quartz pebble conglomerate in semicontinuous lenses and stringers, especially near the base. Incorporated into the quartz pepple lags are pebble- to cobble- sized rip-up clasts of the underlying coaly facies. These rounded clasts and coal balls are also found isolated in the sandstone. These occurrences suggest erosion from cut banks and entrainment of this collapsed material into the stream bed-load.

Cotter (1978) interpreted the Pocono Formation of central Pennsylvania (=Burgoon and Purslane) as a braided-channel deposystem. The morphology of braided rivers is complex, but generally shows individual bedforms or bars, bar systems or complexes and vegetation stabilized islands (Walker and Cant, 1984, p. 80). Large-scale planar crossbed sets in the Purslane result from migrating transverse and longitudinal bars on the braidplain. Because of its apparent areal extent, the coaly lithofacies may have been a vegetated, and thus stabilized, interfluvial area between major braided channels. Vertical accretion and lateral migration eventually caused the upper sandstone channel system to erode, significantly scour and overstep the swampy environments that characterized the coaly lithofacies.

Kammer and Bjerstedt (1986, Figure 2) placed an unconformity of unknown magnitude at the top of the Rockwell Formation. An unconformity is suggested by the stratigraphic context of the vertical sequence. The upper Rockwell records sedimentation on an alluvial plain in meandering, suspension-load systems. The Purslane Sandstone records braided, bed-load deposystems that necessitates major changes in gradient, as well as type, size and amount of sediment load (Walker and Cant, 1984). Conformable juxtaposition of deposits from these two deposystems appears unlikely. Further evidence for this relationship is based on the Patton Shale Member which is locally missing due to erosion that accompanied Burgoon deposition in Pennsylvania.

REGIONAL STRATIGRAPHIC RELATIONSHIPS

In this section I will attempt to outline supporting evidence for the two major contentions of this work, namely that, 1) the basal Rockwell and Spechty Kopf diamictites are essentially chronostratigraphic equivalents of the Riddlesburg Shale in Pennsylvania and Maryland and, 2) that the erosional surface upon which the diamictites were deposited represents a major hiatus at the Devonian-Mississippian boundary that is partly concurrent with Early Mississippian Berea Sandstone deposition in the foreland basin of western West Virginia.

Kammer and Bjerstedt (1986) revised the stratigraphic nomenclature of Lower Mississippian rocks in West Virginia and replaced the name "Pocono" with Price in reference to these deposits throughout the state. Lithostratigraphically, the Price Formation encompasses all uppermost Devonian and Lower Mississippian clastics in both Virginias above either the "Chemung" Formation or red beds of the Hampshire Formation and below either the Greenbrier Limestone or the red beds of the Maccrady Formation. Kammer and Bjerstedt (1986) used the Sunbury Shale Member, a basinal dark shale, and its eastern nearshore equivalent, the Riddlesburg Shale Member of the Price Formation, as a stratigraphic datum to trace a major eustatic sea level rise of Early Mississsippian time (Ross and Ross, 1985). Beginning from a marine point of reference in the foreland basin, Kammer and Bjerstedt found that Riddlesburg nearshore sandstones, siltstones and silty shales could be traced from outcrops in central and northern West Virginia northward and eastward into thick, nonmarine deltaic and fluvial clastics.

Sevon (1969, 1979, 1985) reported polymictic diamictites from the base of

the Devonian-Mississippian Spechty Kopf Formation in the Anthracite region of eastern Pennsylvania and mapped their thickness and distribution. Sevon (1979) further identified this distictive lithofacies in the Rockwell Formation of south-central Pennsylvania at Crystal Spring and in the Maryland panhandle at LaVale (Figure 1). The diamictite, however, does not occur in the vertical sequence exposed along I-48, south of Finzel in Garrett County (Figure 1). Based on thickness variations of the diamictite in the Spechty Kopf Formation, Sevon (1969, 1979, 1985) interpreted an unconformable basal contact (Edmunds and others, 1979, Figure 10) with relief on this erosional surface of up to 100 m (Sevon, 1979) in the Anthracite region. Sevon (1969, 1979) noted that the thickest diamictites occur in the hypothesized axes of major sediment-dispersal systems that were subjected to erosion during the Devonian-Mississippian transition.

Depositional Mechanism for the Diamictites

Sevon (1969, 1979) favored a subaqueous debris or mudflow origin for the diamictites, with maximum thicknesses corresponding to eroded channels on the relict Catskill alluvial plain. Sevon (1979) reported that the basal Spechty Kopf and Rockwell diamictite lithofacies had widths of about 1 to 3 km parallel to depositional strike (northeast-southwest) and about 64 km normal to depositional strike indicating that the diamictites were deposited in dip-oriented systems.

A major argument against a subaerial origin for the diamictites rests in the stratigraphic context of this lithofacies. If the diamictites had resulted from subaerial debris flows on alluvial fans, then other alluvial fan facies should be expected in the vertical sequence (Rust and Koster, 1984). Bull (1972, p. 72) noted that very heterogenous bedding sequences are important clues for recognizing alluvial fan deposystems. Debris flows deposits are common in alluvial fan systems, however, water-laid, bedded gravels and seive deposits may likewise be present (Zaitlin and Rust, 1983). Subaqueous deposition is further suggested by large-scale dewatering and diapiric structures in the diamictite caused by rapid syn- and post-depositional expulsion of water from the saturated matrix. Deposition of the diamictite in alluvial fan environments is difficult to imagine because the tectonic event(s) needed to make such a subaerial system possible, ie. a proximal, high-relief source area, is not reflected in suprajacent lithofacies at Sideling Hill which contain a restricted, marginal-marine fauna.

Sevon (1981, p. 43, Figure 2D) presented an interpretation of the tectonic events that led to diamictite sedimentation. Citing the work of Dewey and Kidd (1974), he suggested a single depositional event or multiple, closely-spaced events that released pebble- to boulder-sized alluvium from a tectonic basin that formed by extensional tectonism and rifting during separation of the North American and South American plates at the end of the Acadian Orogeny. Continued filling of the tectonic basin eventually breached its rifted western margin and the impounded debris was washed into the foreland basin during a catastrophic flood(s).

Terminal Acadian tectonic events that preceeded deposition of the Mississippian clastic wedge in the Appalachian basin has been attributed to oblique-slip collision of the Avalon terrane with the New York and Virginia promontories (Ettensohn, 1985). This mechanism involves convergence rather than extensional tectonism. Could the diamictites have resulted from coarse-grained alluvium deposited in an ephemeral "piggyback" basin that formed on advancing thrust sheet(s)? A breach through the topographic front of a thrust sheet that carried such a basin would allow a rapid change in gradient and furnish the erosional power to account for the emplacement of the exotic, very coarse-grained debris in the diamictite.

Seven and Berg (1986) regarded a laterally restricted but regionally recurrent lithologic sequence at the base of the Spechty Kopf and Rockwell

formations to represent lacustrine deposition on the relict Catskill Formation alluvial plain following a period of erosion at the Devonian-Mississippian transition. This lithologic sequence consists of; 1) a basal polymictic diamictite deposited on an unconformable surface, 2) pebbly mudstone, 3) laminite, and 4) planar bedded, well-sorted quartz arenite. They favored a lacustrine origin because of the lack of fossils in any of the above lithofacies and because of the thin-bedded, varve-like lithofacies (laminite) that overlies diamictite and pebbly mudstone in their idealized sequence (Sevon and Berg, 1986; oral communication, 6/86).

Based on regional stratigraphic study, I would argue for a marine depositional environment, namely, the barred-bay and estuarine facies that record a nearshore facies complex related to a major Early Mississippian transgression. The fining-upward, medium-grained sandstone lithofacies overlying the diamictite (unit 3; Figure 2C) contains sedimentary structures usually considered important indicators of marine influence. Coarse-grained sandstone near the base fines upward to medium-grained sandstone and large- to medium-scale trough crossbed sets near the base generally decrease in scale upsection. In the upper few meters, current-ripples, small-scale herringbone and wedge crossbed sets indicate tidal influences and waning current flow (Reinson, 1984, p. 127). Although *Skolithos* burrows are not diagnostic of marine influence (Fitzgerald and Barrett, 1986), they are further supporting evidence for it. A tidal inlet-fill sequence is suggested based on the above criteria and comparison with published vertical sequences (Clifton, 1982, p. 186).

Overlying black to dark grey, siltstones and very fine-grained sandstones (units 4-5) contain a restricted fauna of bivalves, orbiculoids and chonetid brachiopods. Abundant pyrite nodules from 1 to 3 cm in diameter and thin sideritic layers up to 3 cm thick further suggest deposition in organic- rich lagoon and marsh environments. Wavy bedding that results from mud- draping of current- and wave-ripples is a common feature of estuarine deposition (Clifton, 1982, p. 187). The very thin coaly shales on top of washover deposits of tabular and scour-based sandstones suggest deposition in back-bar lagoon environments (Reinson, 1984, p. 129).

The marine influences that are preserved in the lower Rockwell Formation in western Maryland may be reconciled with Sevon and Berg's (1986) interpretation of lacustrine deposition if the paleoclimatic setting is considered. The Appalachian basin was latitudinally disposed along the equator during the Devonian-Mississippian transition. According to the paleogeographic and paleoclimatic interpretation of Ettensohn and Barron (1981, p. 349) the northern Appalachian basin lay near the zone of convective precipitation that lies within 5 degrees latitude of the equator. High fresh-water influx from rainfall may have caused further dilution of already brackish-water environments in the shallow and restricted eastern Penn-York Embayment, an area far removed from basinal, normal-marine circulation.

Unconformity at the top of the Hampshire and Catskill Formations

A disconformable contact occurs between the Cussewago Sandstone and Riddlesburg Shale members of the Price Formation at the principal reference section for northern West Virginia at Rowlesburg (Kammer and Bjerstedt, 1986) (Figure 5). This unconformity is related to widespread regional emergence and sediment bypassing of the exposed coastal plain during the extensive Early Mississippian regression marked by the Berea Sandstone of West Virginia and Ohio and the Cloyd Conglomerate Member of the Price Formation in southern West Virginia and Virginia. Berea-Cloyd fluvial deposystems built westward across relict Cussewago and age-equivalent facies and carried sediment to the Berea shelf edge in western West Virginia (Larese, 1974) (Figure 6A).

This regression corresponds to the end of the third tectophase of the Acadian Orogeny (Ettensohn, 1985). Oblique-slip collision of the Avalon terrane with the New York and Virginia promontories is a likely cause of this

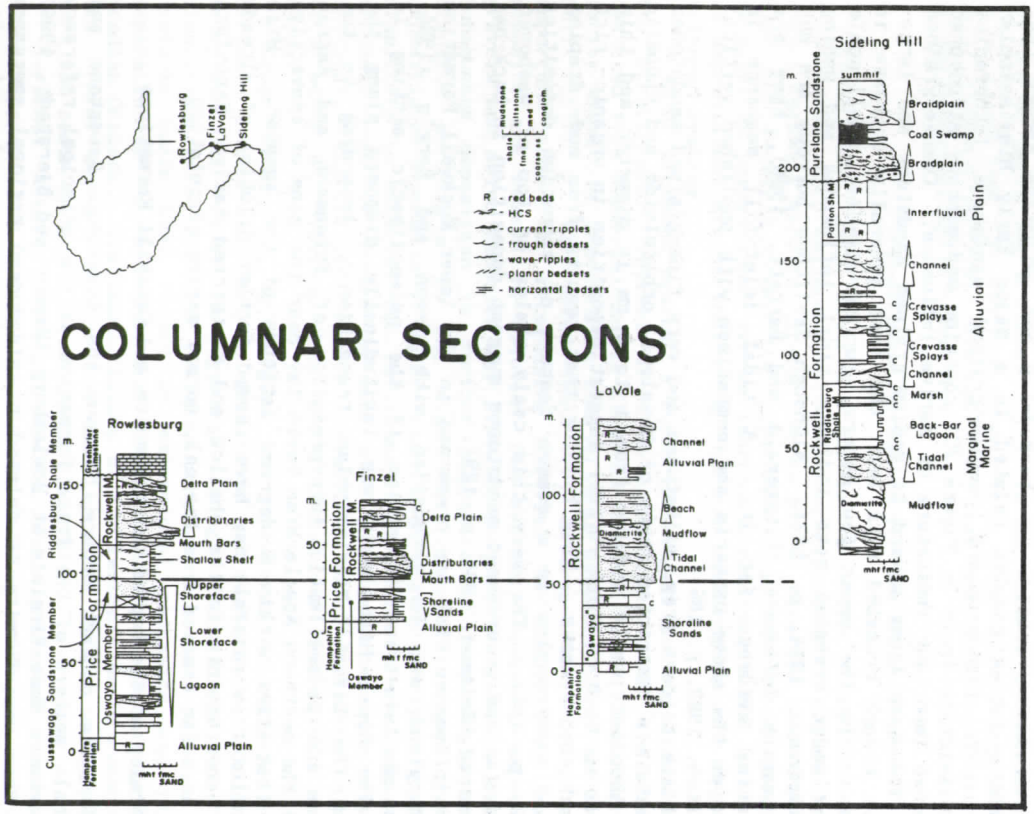


Figure 5. East-west stratigraphic sections between Rowlesburg, West Virginia and Sideling Hill, Maryland showing interpreted depositional systems and Upper Devonian-Lower Mississippian rock-stratigraphic nomenclature recognized for the area. Datum is the Early Mississippian Riddlesburg Shale Member in West Virginia and interpreted correlatives to the east. The base of the Early Mississippian transgression is also regarded as the sub-Berea unconformity surface. Riddlesburg marine facies transgressed Cussewago and age-equivalent facies that were subaerially exposed as a result of the extensive Berea Sandstone regression in West Virginia, Pennsylvania and Maryland. *C=coaly horizons; triangles refer to large-scale grain size trends. Lithologies, structures and grain sizes are defined in the legend.*

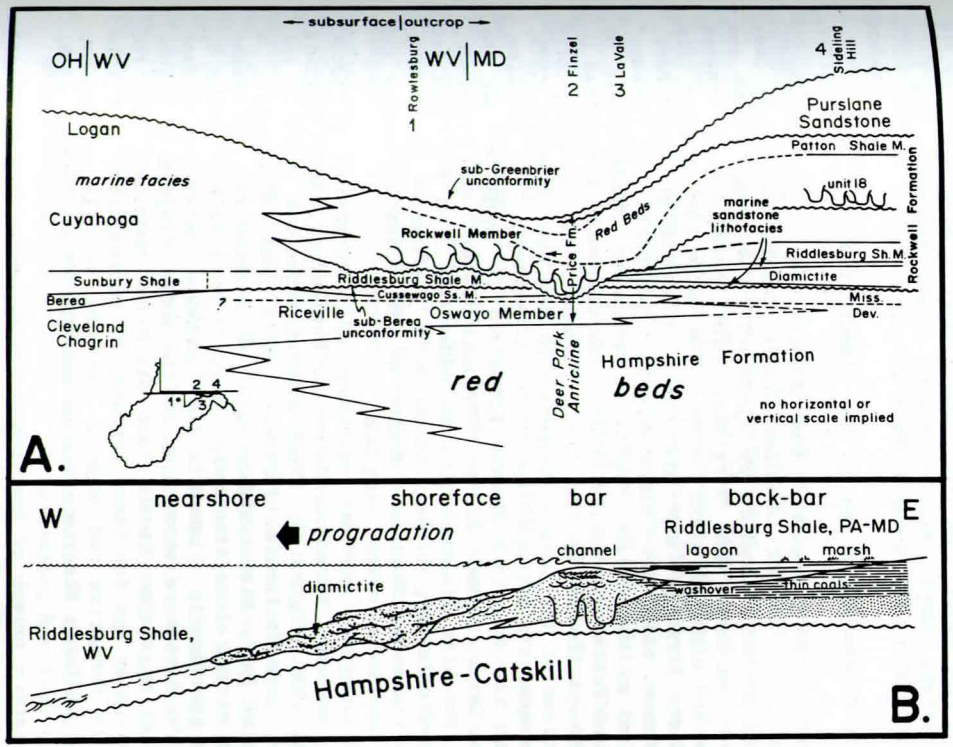


Figure 6. A. Cartoon wedge diagram showing interpreted stratigraphic relationships from the northern West Virginia subsurface to the Sideling Hill section. Kammer and Bjerstedt (1986) preferred to recognize Price Formation rock-stratigraphic nomenclature where the Oswayo Member could be recognized on the outcrop except at LaVale where red bed lithofacies occur above and below an Oswayo "tongue". Generalized subsurface data after Boswell (1985). A thinner Oswayo stratigraphic section at Finzel and absence of Riddlesburg-age deposits may have resulted from incipient activation of the Deer Park Anticline at the close of the Acadian Orogeny. No scale is implied. B. Cartoon reconstruction of depositional environments and the relationship of the Riddlesburg Shale of Pennsylvania and Maryland with marine-bar and diamictite lithofacies and also the Riddlesburg Shale Member of the Price Formation in West Virginia. No scale is implied.

extensive regression (Ettensohn, 1985; Thomas, 1977). The unconformity below the basal Rockwell diamictite at Sideling Hill and Crystal Spring is interpreted to be the same unconformity that occurs below basal diamictites of the Spechty Kopf Formation in the Anthracite region recognized by Berg and others (1983). This interpretation is based on the unique nature of the diamictite lithofacies in this spatial and temporal setting which makes this lithofacies analogous to a time-stratigraphic marker. The relatively minor, sub-Berea Sandstone unconformity in the foreland basin of West Virginia coalesces eastward with the major unconformity at the top of the Hampshire and equivalent Catskill Formation in Maryland and Pennsylvania (Figure 6A). The major unconformity was probably not caused solely by tectonic events leading to the Berea regression, but the depositional hiatus or erosional interval at the top of the Hampshire and Catskill formations spans the time during which Berea deposition took place in West Virginia.

Early Mississippian Age for the Diamictites

Sevon (1969, 1979, 1985) reported the diamictites to be Devonian-Mississippian in age. The latest Devonian Oswayo transgression has been traced on the outcrop in West Virginia by Bjerstedt (1986) and Kammer and Bjerstedt (1986). In northern West Virginia, this transgression drowned Hampshire delta plain and alluvial plain facies. The rocks that express this transgression in West Virginia are green-grey bioturbated sandstones, siltstone and mudstones that are lithologically identical to "Oswayo" lithofacies identified in the inlier region of southwest Pennsylvania by Laird (1941, p. 12) and Fettke and Bayles (1945). As currently recognized in West Virginia the Oswayo Member of the Price Formation is lithologically identical to descriptions of the Oswayo Formation in New York (Tesmer, 1975, p. 76) and is in the same stratigraphic position in sequence. In addition, this correlation with the New York Oswayo Formation is supported by the fossil content of the Oswayo Member in West Virginia (Kammer and Bjerstedt, 1986). This latest Devonian transgression is separated from the earliest Mississippian Sunbury-Riddlesburg transgression by a regressive phase that deposited the marine Cussewago Sandstone Member of the Price Formation and the subsurface Berea Sandstone in West Virginia.

The Devonian-Mississippian systemic boundary is interpreted to occur about 9 m below the Cussewago Sandstone Member at Rowlesburg, West Virginia (Kammer and Bjerstedt, 1986, Figure 4). This determination was based on the first occurrence of *Schellweinella inflata* (White and Whitfield), a brachiopod indicative of Early Mississippian age. An unnamed red bed member 11.5 m (38 ft.) thick occurs above the Oswayo "tongue" at LaVale (Figures 5, 6A) and appears to be genetically related to the Cussewago regression in West Virginia. Based on the facies relationships in the wedge diagram of Figure 6A, earliest Mississippian marine facies that pinch out eastward into red beds support a Mississippian age for these terrestrial equivalents. For this reason, upper Hampshire Formation red beds to the east of Allegany County, Maryland, are regarded as Early Mississippian in age (Craig and Connor, 1979, Plate 15A).

The approximate eastern extent of the Oswayo transgression can be traced along the Allegheny Front in Grant and Mineral Counties, West Virginia and Allegany County, Maryland (Beuthin and Dennison, 1985; Bjerstedt, 1986). Green-colored, cross-bedded sandstones that interbed with red beds of the Catskill Formation as far east as the Broad Top Basin in Pennsylvania (Willard, 1939, p. 277, 303) represent the extreme eastern feather-edge of the Oswayo transgression. Oswayo shoreline and tidal flat facies occur at LaVale below the diamictite (Figure 5) and it is likely that the Oswayo pinches-out west of the Sideling Hill exposure although this section is not deep enough to demonstrate this conclusively. The diamictites, therefore, were not deposited during the latest Devonian Oswayo transgression, but during the Early Mississippian Sunbury-Riddlesburg transgression.

The position of the diamictite in the vertical sequence is one of the strongest arguments for an Early Mississippian age. The stratigraphic sections in Figure 5 and the diagrammatic cross section in Figure 6A show that the Riddlesburg Shale cannot be directly traced eastward to the Maryland exposures at Finzel and LaVale, however, marine facies genetically related to the major base level rise of Early Mississippian time can be identified. At LaVale it is represented, from oldest, by a tidal inlet-fill sandstone lithofacies, diamictite, and an overlying horizontally-laminated sandstone lithofacies representing shoreface or foreshore beach(?) deposits. At Sideling Hill, it is represented, from oldest, by an interbedded sandstone and diamictite lithofacies that is incompletely exposed, a tidal inlet-fill sandstone and overlying dark grey siltstones and very fine- grained sandstones deposited in back-bar lagoon environments (Riddlesburg Shale Member).

A thinner Oswayo section occurs at Finzel (Figure 5) between thicker sections to the west at Rowlesburg and to the east at LaVale. Neither the Riddlesburg Shale or sandstone-diamictite are preserved at the Finzel exposure possibly due to activation of foreland detached structures during the Devonian- Mississippian transition, specifically, the Deer Park Anticline. Incipient movement of this feature at the close of the Acadian Orogeny may have resulted in erosional thinning of the Oswayo Member at Finzel and removal of Early Mississippian Riddlesburg-age facies (Figure 6A).

The upper Rockwell Formation at Sideling Hill is preserved in northern West Virginia as the Rockwell Member of the Price Formation (Kammer and Bjerstedt, 1986). A regional unconformity occurs at the base of the Rockwell Member that has contributed to thinning of Early Mississippian marine deposits in north-central West Virginia.

The lower Rockwell Formation at Sidling Hill exposes diamictite, marine-bar and back-bar lagoon facies that were never deposited in northern West Virginia. At Sideling Hill, these marginal-marine facies span 76 m (250 ft.) of stratigraphic section, whereas only 18 m (60 ft.) of Riddlesburg Shale is preserved at the Rowlesburg exposure. Based on subsurface evidence, Boswell (1985, p. 79) mapped and interpreted the "Weir sand" interval to represent elongate, fluvially-dominant deltaic systems that are approximately equivalent to the Rockwell Member on outcrop. As shown by study of outcrop correlatives to the east, these elongate deltaic systems in the subsurface of West Virginia were part of a complex of deltaic and interdeltaic bays and lagoons. These marginal-marine environments have been completely preserved to the east at Sideling Hill. As shown by outcrop study in northern West Virginia (Bjerstedt, 1986), there is erosional relief up to 12 m (40 ft.) on the top of the Riddlesburg Shale Member. Lateral planation and erosional incision by Rockwell Member deltaic systems has removed some of these marine environments. Large-scale removal of Riddlesburg-age facies has taken place in western Maryland (Figure 6A).

Rockwell and Spechty Kopf Diamictites and the Relationship with the Riddlesburg Shale of Pennsylvania, Maryland and West Virginia

I argue for a genetic relationship between the diamictites of the basal Rockwell and Spechty Kopf formations and the Riddlesburg Shale Member of the Rockwell Formation of Maryland and Pennsylvania. Both systems are directly related to the same major Early Mississippian sea level rise.

The exact age- and facies-relationship of the marine Riddlesburg Shale to the thick fluvial facies in which it occurs has never been fully understood. Kammer and Bjerstedt (1986) expressed no significant doubt as to the Mississippian age of the Riddlesburg in Pennsylvania. The type section of the Riddlesburg Shale (Reger, 1927) occurs in Bedford County, 2.4 km (1.5 mi.) north of Riddlesburg, Pennsylvania and 24 km (15 mi.) north of Crystal Spring (Figure 1). Descriptions of the Riddlesburg Shale (White, 1885; Lesley, 1895; Reger, 1927; Laird, 1942) noted its fossil content, black to

dark grey color and tendency to weather into elongate slivers and pencils. An interbedded lithology of thin-bedded sandstones, siltstones and shales with wave-rippled bedding planes is characteristic in exposures along the Structural Front at Altoona. The Riddlesburg fauna varies in type, abundance and diversity at different exposures in the Broad Top Basin and along the Structural Front but appears to contain a normal-marine fauna only at its type section (Reger, 1927, p. 400; Girty, 1928). More restricted environments generally occur to the east.

As currently recognized, the Riddlesburg Shale in Pennsylvania represents a variety of nearshore marine to marginal-marine environments deposited in near contemporaneity with diamictite sedimentation. To the west, the Riddlesburg Shale Member of the Price Formation in West Virginia records shallow-shelf facies based on the preservation of hummocky cross-stratified, fine-grained sandstones at Rowlesburg. In Maryland and Pennsylvania, Riddlesburg deposystems included restricted back-bar lagoon and marsh systems (Sideling Hill, Crystal Spring) to nearshore and estuarine systems bearing *Cruziana* trace fossil assemblages (Altoona). The Riddlesburg fauna also varies at different exposures from more euryhaline (Altoona) to stenohaline (type section) assemblages. The wedge diagram in Figure 6A suggests that the Riddlesburg Shale is time transgressive and younger to the east. The Riddlesburg Shale in West Virginia represents slightly older lithologies of the same diachronous lithosome.

In Pennsylvania and Maryland, Riddlesburg dark grey siltstones and shales were deposited in areally extensive marginal-marine environments. So far, these environments have been recognized in west-central and south-central Pennsylvania. In his initial description of the Riddlesburg Shale, Reger (1927, p. 405) suggested that this unit occurred far to the northeast of its type section located on the western margin of the Broad Top Basin. He stated, "Farther northeast the same peculiar lithology, apparently without marine fossils, but in several cases with rather abundant plant fossils, is visible in the same relative stratigraphic position in the gaps of the Susquehanna River between Harrisburg and Millersburg, in the gap of the Schuylkill below Pottsville, and in the gap of the Lehigh opposite Mauch Chunk (now Jim Thorpe)." Further detailed work in the Anthracite region may confirm this relationship.

A section at Crystal Spring (Figure 1) along the I-70 highway cut through Ray's Hill at the southern margin of the Broad Top Basin, provides the most complete exposure in south-central Pennsylvania that clearly shows the relationship between the diamictite lithofacies and the Riddlesburg Shale (Figure 7). Descriptions of the field units shown in Figure 7 are contained in Table 2. The Crystal Spring section clearly shows a regressive stratigraphic sequence. Up to 1 m of erosional relief occurs on the top of Catskill Formation red beds. The overlying diamictite lithofacies is interbedded with fine- to medium-grained shoreface sandstone and thin-bedded, wave-rippled siltstone and silty shale. These shoreface facies coarsen upward in large-scale and are capped by a tidal inlet-fill sequence. This composite shoreface-bar system is overlain by back-bar lagoon and marsh facies comprised of a thick sequence of barren, olive-grey to dark grey siltstone and silty shale that weathers to thin slivers and pencils (Riddlesburg Shale Member). Interbedding of sandstone and diamictite resulted from mud and debris flows that accumulated in the axes of estuary channels. This material continued to move seaward, down sedimentary dip, and eventually passed through tidal channels and larger-scale inlets between bar systems to interbed with sandy shoreface facies (Figure 6B).

In accord with Walther's Law of Correlation of Facies (Middleton, 1973), the position in sequence of the Riddlesburg Shale indicates not only a genetic relationship with diamictite deposystems but lateral contemporaneity as well (Figure 6B). At Sideling Hill, a thickness of about 40 m (130 ft.) was determined using pace-compass technique from the base of the massive diamictite (unit 2; south cut face) to the first red mudstone lithofacies

(definite Hampshire Formation). Based on the excellent supplementary section at Crystal Spring, this partly covered interval at Sideling Hill conceals a similiar interbedded sandstone-diamictite sequence that was deposited on an erosional surface at the top of the Hampshire Formation.

CONCLUSIONS

The Lower Mississippian Rockwell Formation preserves a regressive sedimentary sequence that built westward into the foreland basin following a major sea level rise in Early Mississippian time. The stratigraphic sequences at Sideling Hill, Crystal Spring and LaVale demonstrate that the Riddlesburg Shale Member of the Rockwell Formation in Pennsylvania and basal Rockwell polymictic diamictites were deposited in genetically related, laterally adjacent marine systems. The thickest basal Rockwell diamictites accumulated in the hypothesized axes of Hampshire-Catskill sediment-

Table 2. Brief description of lower Rockwell Formation field stratigraphic units at the Crystal Spring section, along the I-70 road cut through Ray's Hill. Most rock colors are light olive grey (5Y5/2) due to weathering, only different colors are noted.

UNIT	DESCRIPTION	THICKNESS (m)	(ft)
13	very fine-grained SANDSTONE: medium- to thick-bedded, scour base, large- to medium-scale trough crossbed sets and horizontal laminations, units fines and bedding thins upsection.	10.6	34.7
12	silty SHALE and MUDSTONE with minor SILTSTONE interbeds: weathers to slivers and pencils, sandstone and siltstone interbeds in lower part.	17.4	57.3
11	very fine-grained SANDSTONE and SILTSTONE: medium-bedded, common current-ripple laminations.	8.2	27
10	silty SHALE and MUDSTONE: olive grey (5Y4/1); weathers to thin slivers and pencils, barren.	20	65.5
9	pebbly, fine- to coarse-grained SANDSTONE: very thick-bedded to massive, conglomeratic sandstone in basal 2 m with large- to medium-scale trough crossbed sets, upper part contains horizontally-laminated sandstone with small-scale wedge, and bidirectional (herringbone) bed sets, units fines and bedding thins upward.	12.1	40
8	SILTSTONE with interbedded silty SHALE and very fine-grained SANDSTONE: thin- to medium-bedded, wave-ripples (short wavelengths) on most bedding planes, load casts and ball and pillow structures common, becomes coarser-grained and thicker-bedded upsection.	9.1	30
7	SILTSTONE: very thin-bedded, horizontally-laminated. "laminite" of Sevon and Berg (1986).	1.5	5
6	POLYMICTIC DIAMICTITE: greyish red (5R4/2); unbedded, scour base, cobbles up to 40 cm.	7	23.1
5	very fine-grained SANDSTONE: medium- to thick-bedded, wave-ripples (long wavelengths), flow rolls at upper contact with diamictite.	11.7	38.5
4	SILTSTONE: very thin-bedded, horizontally-laminated "laminite" of Sevon and Berg (1986).	2.7	9
3	POLYMICTIC DIAMICTITE: variegated but generally pale brown (5Y5/2); unbedded, large-scale flowage structures, one major fine-grained sandstone interbed, cobbles in sandy mudstone matrix up to 30 cm in diameter.	10.6	34.7
2	very fine- to fine-grained SANDSTONE: medium- to thick-bedded, common horizontal laminations, lenses of mud-clast conglomerate, flow rolls at upper contact with diamictite.	9.4	30.8
1	MUDSTONE with interbedded fine-grained SANDSTONE: greyish red (5R4/2); partial thickness, up to 1.5 m of erosional relief on upper contact.	23.5	77

dispersal systems. These systems became estuaries and barred-bay environments as a result of a major transgression in earliest Mississippian time.

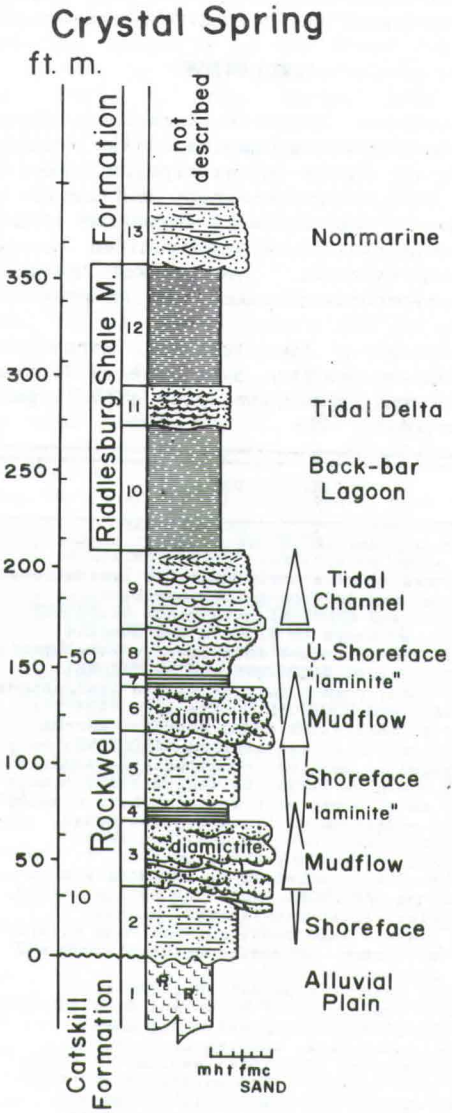


Figure 7. Columnar section of the lower part of the Rockwell Formation at Crystal Spring, along I-70 at the Fulton and Bedford county line. The section shows clearly interbedded diamictite, fine- to medium-grained sandstone and thin- to very thin-bedded, wave-rippled, siltstone and silty shale. This lithologic package is overlain by a thick sequence of olive grey to grey siltstone and silty shale referable to the Riddlesburg Shale Member. The Riddlesburg weathers to thin slivers and pencils and appears to contain no fossils, however, this lithofacies is poorly exposed compared to the Sideling Hill exposure. Lithologic symbols are as defined in Figures 3 and 5.

The Riddlesburg Shale Member at Sideling Hill contains a restricted fauna deposited in back-bar lagoon and marsh environments. This lithofacies

consists of black to dark grey, very fine-grained sandstone and siltstone with abundant thin coaly shales. Based on its fauna and lithology, exposures of the Riddlesburg Shale in the Broad Top Basin area and along the Structural Front at Altoona show that the Riddlesburg preserves a variety of depositional environments ranging from highly restricted back-bar lagoon and marsh facies to euryhaline and stenohaline nearshore and estuarine facies. The Riddlesburg Shale Member of the Price Formation in West Virginia and of the Rockwell Formation in Pennsylvania and Maryland are part of the same diachronous lithosome deposited in an eastwardly transgressing sea.

Based on position in vertical sequence, the basal Rockwell diamictites in south-central Pennsylvania and Maryland and by implication, the basal Spechty Kopf diamictites in the Anthracite region, are gross chronostratigraphic equivalents of the Riddlesburg Shale Member of the Rockwell Formation in Pennsylvania and Maryland. Riddlesburg Shale deposystems are nearshore equivalents of the basinal Sunbury Shale and were deposited during the same major transgression. These diverse and areally extensive facies preserve an environmental gradient from the basin-plain in Kentucky and Ohio (Sunbury) through shallow-shelf and nearshore environments in West Virginia (Riddlesburg), and estuarine, back-bar lagoon and marsh environments in Pennsylvania and western Maryland (Riddlesburg).

The upper Rockwell Formation preserves alluvial plain facies characterized by suspension-load fluvial systems, crevasse splay deposits, fine-grained overbank deposits and incipient paleosols. Red beds of the Patton Shale Member preserve well-drained interfluvial areas. The Purslane Sandstone unconformably overlies the Patton Shale Member and preserves bed-load fluvial deposystems and coal swamp environments deposited on a braidplain.

ACKNOWLEDGEMENTS

I thank Mr. Ronald Ritchey of Holloway Construction Company for allowing access to the Sideling Hill exposure while the final phases of highway preparation were still in progress. I thank Dr. Thomas M. Berg and Dr. William D. Sevon, Pennsylvania Geologic Survey and Dr. Frank R. Ettensohn, University of Kentucky for their extremely helpful suggestions during the review process.

This report was prepared during the course of field work for the author's Ph.D. dissertation at West Virginia University under the supervision of Thomas W. Kammer who also reviewed this manuscript. Support for this study was provided by a National Science Foundation (E.P.S.Co.R.) grant to T. W. Kammer.

The goals of this paper were focused considerably by observations made during a field trip to the Sideling Hill exposure in May, 1986 by the author and T. W. Kammer of West Virginia University, T. M. Berg, W. D. Sevon, J. H. Harper and C. H. Dodge of the Pennsylvania Geologic Survey, D. L. Woodrow of Hobart and William Smith Colleges, W. E. Edmunds, Consulting Geologist and J. R. Eggleston of the U.S. Geological Survey. These geologists made very significant observations and I deeply appreciated the opportunity to discuss stratigraphic and sedimentologic aspects of the Sideling Hill exposure with them.

REFERENCES

Berg, T.M. and Edmunds, W.E., 1979, The Huntley Mountain Formation: Catskill-to-Burgoon transition in north-central Pennsylvania: Pennsylvania Topographic and Geologic Survey, Information Circular 83, 79 p.

Berg, T.M. and others, 1983, Stratigraphic correlation chart of Pennsylvania: Pennsylvania Topographic and Geologic Survey, General geology report 75, 1 p.

Bjerstedt, T.W., 1986, Stratigraphy and deltaic depositional systems of the

Price Formation (Upper Devonian-Lower Mississippian) in West Virginia: unpublished Ph.D. dissertation, West Virginia University, Morgantown, WV, (in preparation)

Beuthin, J.D. and Dennison, J.M., 1985, The Late Devonian-Early Mississippian marine transgressive event in western Maryland and immediately surrounding areas: Appalachian Basin Industrial Associates Proceedings, v. 8, p. 77-104.

Boswell, R.M., 1985, Stratigraphy and sedimentation of the Upper Devonian-Lower Mississippian Acadian clastic wedge in northern West Virginia: unpublished master's thesis, West Virginia University, Morgantown, WV., 179 p.

Bull, W.B., 1972, Recognition of alluvial fan deposits in the stratigraphic record: *in* Hamblin, W.K. and Rigby, J.K., (eds.), Recognition of ancient sedimentary environments, Society of Economic Paleontologists and Mineralogists, Special Publication 16, p. 63-83.

Butts, C., 1905, Description of the Ebensburg quadrangle: U.S. Geological Survey, Geologic Atlas of the U.S., Folio 133, 9 p.

Campbell, M.R., 1904, Description of the Latrobe quadrangle: U.S. Geological Survey, Geologic Atlas of the U.S., Folio 110, 15 p.

Clifton, H.E., 1982, Estuarine deposits: *in* Scholle, P.A. and Spearing, D. (eds.), Sandstone depositional environments, American Association of Petroleum Geologists, Memoir 31, p. 179-189.

Collinson, J.D., 1978, Alluvial sediments: *in* Reading, H.G. (ed.), Sedimentary environments and facies, Elsevier, New York, p. 15-60.

Cotter, E., 1978, The evolution of fluvial style, with special reference to the central Appalachian Paleozoic: *in* Miall, A.D., (ed.), Fluvial sedimentology, Canadian Society of Petroleum Geologists, Memoir 5, p. 361-383, Calgary, Alberta.

Craig, L.C. and Connor, C.W. (eds.), 1979, Paleotectonic investigations of the Mississippian System in the United States: U. S. Geological Survey, Professional Paper 1010, 559 p.

Dally, J.E., 1956, Stratigraphy and paleontology of the Pocono Formation in West Virginia: unpublished Ph.D. dissertation, Columbia University, NY, 241 p.

Dennison, J.M. and Wheeler, W.H., 1975, Stratigraphy of Precambrian through Cretaceous strata of probable fluvial origin in southeastern United States and their potential as uranium source rocks: Southeastern Geology, Special Publication 5, 210 p.

Dewey, J.F. and Kidd, W.S.F., 1974, Continental collisions in the Appalachian-Caledonian orgenic belt: variations related to complete and incomplete suturing: Geology, v. 11, p. 543-546.

Edmunds, W.E., Berg, T.M., Sevon, W.D., Piotrowski, R.C., Heyman, L. and Rickard, L.V., 1979, The Mississippian and Pennsylvanian (Carboniferous) Systems in the United States - Pennsylvania and New York: U.S. Geological Survey, Professional Paper 1110B, p. B1-B33.

Ettensohn, F.R., 1985, The Catskill Delta complex and the Acadian Orogeny: a model: *in* Woodrow, D.L., and Sevon, W.D. (eds.), The Catskill Delta, Geological Society of America, Special Paper 201, p. 39-49.

Ettensohn, F.R. and Barron, L.S., 1981, Depositional model for the Devonian-Mississippian black shales of North America: A paleoclimatic-paleogeographic approach: *in* Roberts, T.G. (ed.), Geological Society of America Cincinnati Field Trip Guidebook, Volume 1: Stratigraphy and Sedimentology, p. 344-361.

Fettke, C.R. and Bayles, R.E., 1945, Conenaugh Gorge section of the Mississippian System southeast of Cramer, Pennsylvania, Proceedings of the Pennsylvania Academy of Science, v. 19, p. 86-95.

Fitzgerald, P.G. and Barrett, P.J., 1986, *Skolithos* in a Permian braided river deposit, southern Victoria Land, Antarctica: Paleogeography, Paleoclimatology, Paleoecology, v. 52, p. 237-247.

Folk, R.L., 1968, Petrology of sedimentary rocks: Hemphill's, Austin,

TX, 170 p.
Folk, R.L., Andrews, P.B. and Lewis, W.D., 1970, Detrital sedimentary rock classification and nomenclature for use in New Zealand: New Zealand Journal of Geology and Geophysics, v. 13, p. 937-968.
Girty, G.H., 1928, The Pocono fauna of the Broad Top coal field Pennsylvania: U.S. Geological Survey, Professional Paper 150-E, p. 111-127.
Goddard, E.N. and others, 1975, Rock color chart: Geological Society of America.
Grimsley, G.P., 1916, Jefferson, Berkeley, and Morgan Counties: West Virginia Geologic Survey, County Reports, 644 p.
Ingram, R.L., 1954, Terminology for the thickness of stratification and parting units in sedimentary rocks: Geological Society of America Bulletin, v. 65, p. 937-938.
Kammer, T.W. and Bjerstedt, T.W., 1986, Stratigraphic framework of the Price Formation in West Virginia: Southeastern Geology, (in press)
Laird, W.M., 1941, The Upper Devonian and Lower Mississippian of southwestern Pennsylvania: Pennsylvania Topographic and Geologic Survey, Progress Report 126, 23 p.
Laird, W.M., 1942, The stratigraphy of the Upper Devonian and Lower Mississippian of southwest Pennsylvania: unpublished Ph.D. dissertation, University of Cincinnati, Cincinnati, OH, 244 p.
Larese, R.E., 1974, Petrology and stratigraphy of the Berea Sandstone in the Gay Fink and Cabin Creek trends, West Virginia: unpublished Ph.D. dissertation, West Virginia University, Morgantown, WV, 387 p.
Lesley, J.P., 1985, A Summary description of the geology of Pennsylvania. Pennsylvania Geological Survey, v. 3, part 1, p. 1629-1789.
Miall, A.D., 1978, Lithofacies types and vertical profile models in braided river deposits: A summary: *in* Miall, A.D., (ed.), Fluvial Sedimentology, Canadian Society of Petroleum Geologists, Memoir 5, p. 597-604.
Middleton, G.V., 1973, Johannes Walther's law of the correlation of facies: Geological Society of America Bulletin, v. 84, p. 979-988.
Pelletier, B.R., 1958, Pocono paleocurrents in Pennsylvania and Maryland: Geological Society of America Bulletin, v. 69, p. 1033-1063.
Pettijohn, F.J. and Potter, P.E., 1964, Atlas and glossary of primary sedimentary structures: Springer-Verlag Inc., New York, 370 p.
Reinson, G.E., 1984, Barrier island and associated strand-plain systems: *in* Walker, R.G., Facies Models, Geological Association of Canada, 2nd edition, p. 119-140.
Reger, D.B., 1927, Pocono stratigraphy in the Broadtop Basin of Pennsylvania: Geological Society of America Bulletin, v. 38, p. 397-410.
Ross, C.A. and Ross, J.R., 1985, Late Paleozoic depositional sequences are synchronous and worldwide: Geology, v. 13, p. 194-197.
Rust, B.R. and Koster, E.H., 1984, Coarse alluvial deposits: *in* Walker, R.G., (ed.), Facies Models, Geological Association of Canada, 2nd edition, p. 53-89.
Sevon, W.D., 1969, The Pocono Formation in northeastern Pennsylvania: Guidebook for the 34th Annual Field Conference of Pennsylvania Geologists, Pennsylvania Geological Survey, 69 p.
Sevon, W.D., 1979, Polymictic diamictites in the Spechty Kopf and Rockwell Formations: Guidebook for the 44th Annual Field Conference of Pennsylvania Geologists, Devonian shales in south-central Pennsylvania and Maryland, Pennsylvania Topographic and Geologic Survey, p. 61-65.
Sevon, W.D., 1981, The Middle Devonian clastic wedge in northeastern Pennsylvanina: *in* Enos, P., (ed.), Guidebook for field trips in south-central New York, 53rd annual meeting, New York State Geological Association, Binghamton, NY., p. 31-63.
Sevon, W.D., 1985, Nonmarine facies of the Middle Devonian and Late Devonian

Catskill coastal alluvial plain: *in* Woodrow, D.L. and Sevon, W.D. (eds.), The Catskill Delta, Geological Society of America, Special Paper 201, p. 79-90.

Sevon, W.D., and Berg, T.M., 1986, Polymictic diamictites and associated lacustrine deposits at the Mississippian-Devonian boundary in Pennsylvania: Geological Society of America, Abstracts with Programs, v. 18 (1), p. 65.

Stose, G.W. and Swartz, C.K., 1912, Description of the Pawpaw-Hancock Quadrangles: U.S. Geological Survey, Geologic Atlas Folio 179, 24 p.

Tesmer, I.H., 1975, Geology of Cattaraugus County, New York: Buffalo Society of Natural Sciences Bulletin, v. 27, 105 p.

Thomas, W.A., 1977, Evolution of Appalachian-Quachita salients and recesses from reentrants and promontories in the continental margin: American Journal of Science, v. 277, p. 1233-1278.

Walker, R.G., and Cant, D.J., 1984, Sandy alluvial systems: *in* Walker, R.G., (ed.), Facies Models, Geological Association of Canada, 2nd edition, p. 71-89.

White, I.C., 1885, The geology of Huntingdon County: Pennsylvania Geological Survey, 2nd series, Report T3, 471 p.

Willard, B., 1939, The Devonian of Pennsylvania: Pennsylvania Geological Survey, 4th series, Bulletin G-19, 481 p.

Zaitlin, B.A. and Rust, B.R., 1983, A spectrum of alluvial deposits in the Lower Carboniferous Bonaventure Formation of western Chaleur Bay area, Gaspe and New Brunswick, Canada: Canadian Journal of Earth Sciences, v. 20, p. 1098-1110.

MISSISSIPPIAN REACTIVATION ALONG THE IRVINE-PAINT CREEK FAULT SYSTEM IN THE ROME TROUGH, EAST-CENTRAL KENTUCKY

GARLAND R. DEVER, JR. - *Kentucky Geological Survey*
University of Kentucky
Lexington, Kentucky 40506-0056

ABSTRACT

Extensional growth faulting dominated Paleozoic tectonic activity in the Rome Trough, a linear graben-like structure that extends eastward from central Kentucky into West Virginia. Units ranging in age from Cambrian to Pennsylvanian increase in thickness on the downthrown sides of bounding and interior faults.

Mississippian reactivation along the Irvine-Paint Creek Fault System in the interior of the Rome Trough was followed by erosion of the St. Louis Member of the Slade Formation (Mississippian). Erosional thinning was most pronounced along the border of the upthrown side of the Glencairn Fault. This pattern of erosion suggests that displacement was upward.

The occurrence of differential uplift and erosion contrasts sharply with the dominance of growth faulting and depositional thickening of units throughout most of Paleozoic time. It may be related to a mixed regime of compression and extension that apparently affected faults along the northern border of the trough during Devonian and Mississippian time.

INTRODUCTION

Normal faulting contemporaneous with deposition and influencing patterns of sediment thickness and distribution has been widely recognized in the geologic record. In Cenozoic deposits of the Gulf Coast region, growth faults, or contemporaneous normal faults, are characterized by an increase in the thickness of sediments on the downthrown sides and by an increased magnitude of displacement with depth (Hardin and Hardin, 1961; Ocamb, 1961).

In the Appalachian region, growth faulting dominated Paleozoic tectonic activity in the Rome Trough, a graben-like structure that extends east-northeastward from central Kentucky into West Virginia (Figure 1). Units ranging in age from Cambrian to Pennsylvanian thicken on the downthrown sides of faults (Ammerman and Keller, 1979).

The outcrop belt of Mississippian carbonate rocks of the Slade Formation crosses the Rome Trough along the western border of the Appalachian Plateau (Figures 1 and 2). Study of these rocks along the northern border of the trough, north of the present study area, has shown that units of the lower Slade and underlying Borden Formation have been extensively eroded on the upthrown side of the Kentucky River Fault System and that equivalent strata on the downthrown side show no unusual depositional thickening (Dever, 1977). The erosion is considered to be evidence of upward displacement which is in strong contrast to a general subsidence along the fault system during deposition of older and younger Paleozoic units noted in other studies.

South of the Kentucky River Fault System, the outcrop belt of the Slade Formation crosses the Irvine-Paint Creek Fault System in the interior of the Rome Trough (Figure 1). Previous studies have shown pre- and post-Mississippian growth faulting along this fault system, indicated by the increase in thickness of Cambrian, Silurian-Devonian, Devonian, and Pennsylvanian units on the downthrown side (Weaver and McGuire, 1973; Horne and Ferm, 1978; Dillman, 1980; Webb, 1980).

The purpose of this report is to show that the lower part of the Slade Formation was affected by Mississippian movement along the Irvine-Paint Creek Fault System. Another objective is to compare the Mississippian reactivation with earlier and later fault movements. Depositional thickening in the Slade would be an indication of subsidence, as affected older and younger Paleozoic

units along the fault system. Erosional thinning such as occurred to the north along the Kentucky River Fault System would suggest upward movement.

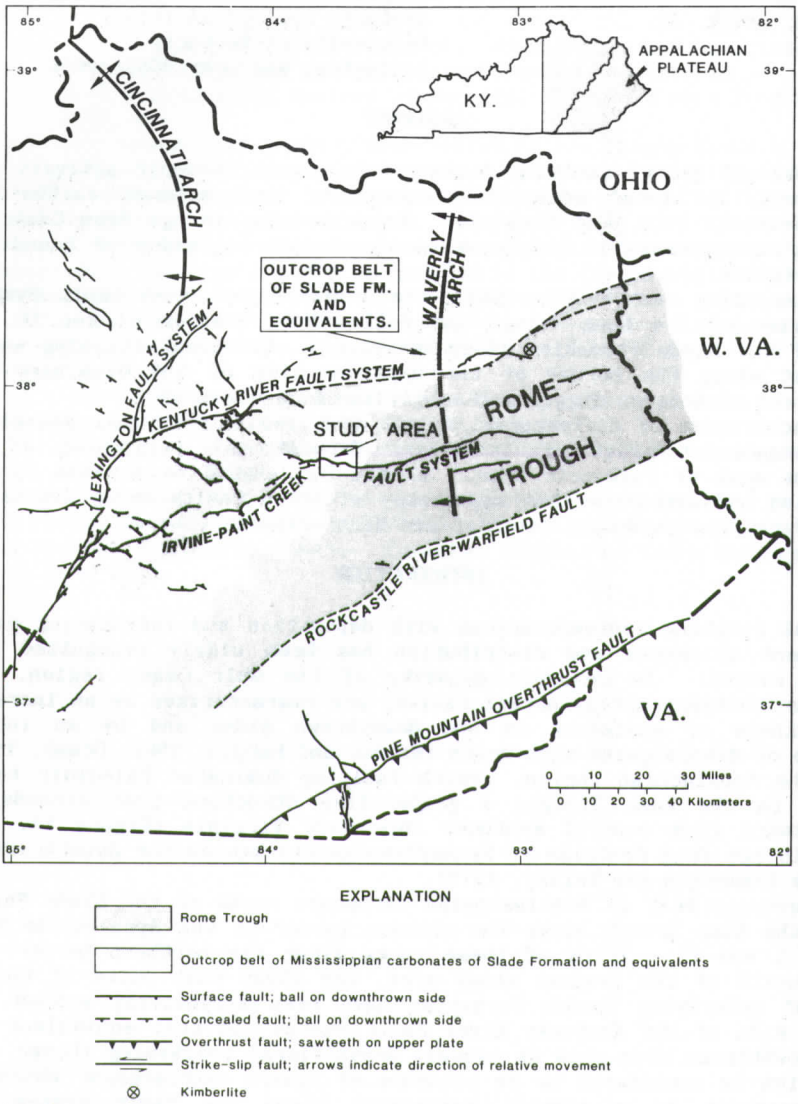


Figure 1. Structural setting of study area in Rome Trough. Surface faults from compilation by Roger Potts. Concealed faults compiled from Ammerman and Keller (1979), Webb (1980), Sutton (1981), Maynor (1984), and data reported by Dever and others (1984). Mississippian axis of Waverly Arch modified from Ettensohn (1977, and MacQuown and Pear (1983).

STRUCTURAL SETTING

The Rome Trough is a linear graben-like structure in the subsurface of eastern and central Kentucky, bounded on the north by the Kentucky River Fault System and on the west by the Lexington Fault System (Figure 1). The southern boundary is complex, with bordering uplifts and basins, but generally is drawn at the subsurface Rockcastle River-Warfield Fault

	WEIR (1974a)	ETTENSOHN AND OTHERS (1984b)
PENNSYLVANIAN	UPPER MBR. OF BREATHITT FM.	
	CORBIN SANDSTONE MBR. OF LEE FM.	
	LOWER TONGUE OF BREATHITT FM.	
MISSISSIPPIAN	PENNINGTON (?) FORMATION	PARAGON FORMATION
	NEWMAN LIMESTONE	SLADE FORMATION
	BORDEN FORMATION	BORDEN FORMATION
DEV.	NEW ALBANY SHALE	

Figure 2. Stratigraphic nomenclature for surface rocks in study area. Shaded interval represents position of the Renfro, St. Louis, Ste. Genevieve, Warix Run, and Mill Knob Members of the Slade Formation.

(Ammerman and Keller, 1979; Webb, 1980; Sutton, 1981) (Figure 1). The trough represents a continental rift zone (Ammerman and Keller, 1979; Webb, 1980). Rifts are prime sites for reactivation by younger, but not necessarily extensional, tectonic regimes (Keller and others, 1983).

The Rome Trough was formed mainly by Cambrian faulting, which may have been partly controlled by Precambrian zones of weakness or faults along the 38th Parallel Lineament of Heyl (1972) (Ammerman and Keller, 1979). Greatest subsidence occurred during Cambrian time (Thomas, 1960; Webb, 1969, 1980; Avila, 1971; Silberman, 1972). Exploratory drilling in eastern Kentucky has shown that the Rome Formation (Cambrian) increases in thickness from 102 to 944 meters (334 to 3,097 feet) southward across the Kentucky River Fault System along the northern boundary of the trough; the formation further thickens to at least 1,919 meters (6,295 feet) in the deepest part of the trough on the downthrown side of the Irvine-Paint Creek Fault System (M. C. Noger, written communication, 1986).

Post-Cambrian faulting of decreased magnitude occurred intermittently along bounding and interior faults, resulting in the thickening of Paleozoic sediments on the downthrown sides: Ordovician (Silberman, 1972; Harris, 1978; Price, 1981); Silurian (Watson, 1983); Silurian-Devonian (Miles, 1972; Weaver and McGuire, 1973; Currie and MacQuown, 1984); Devonian (Dillman, 1980); Mississippian (Ettensohn, 1977); and Pennsylvanian (Haney and others, 1975; Horne and Ferm, 1978; Sergeant, 1979). Basement faulting along the northern boundary of the trough provided the pathway for intrusion of Permian kimberlite in northeastern Kentucky (Silberman, 1972; Bolivar, 1982)(Figure 1).

The Waverly Arch, a low broad arch extending from Ohio southward into eastern Kentucky, was tectonically active in Cambrian and Ordovician time (Woodward, 1961) (Figure 1). Recurrent uplift during Mississippian time affected units in the Slade Formation and its subsurface equivalent, the "Big

Lime" (Dever, 1977; Ettensohn, 1977; MacQuown and Pear, 1983).

LITHOLOGIC CHARACTERISTICS OF THE LOWER PART OF THE SLADE FORMATION

The Slade Formation (Mississippian) consists of carbonate and shale units that formerly were assigned to the Newman Limestone and the upper part of the underlying Borden Formation (Ettensohn and others, 1984b) (Figure 2). The Slade is composed mainly of limestone, with lesser amounts of dolomite and shale. The contacts of the Slade with the underlying Borden Formation and overlying Paragon Formation are gradational. Borden shale grades upward into nodular dolomite and shale of the basal Slade. Limestone of the upper Slade grades upward through interbedded shale and limestone into shale of the basal Paragon.

The lower part of the Slade Formation was studied at 10 sites along the Glencairn Fault, a segment of the Irvine-Paint Creek Fault System in Powell and Wolfe Counties of east-central Kentucky (Figure 1). The lower Slade consists of five members, in ascending order, the Renfro, St. Louis, Ste.

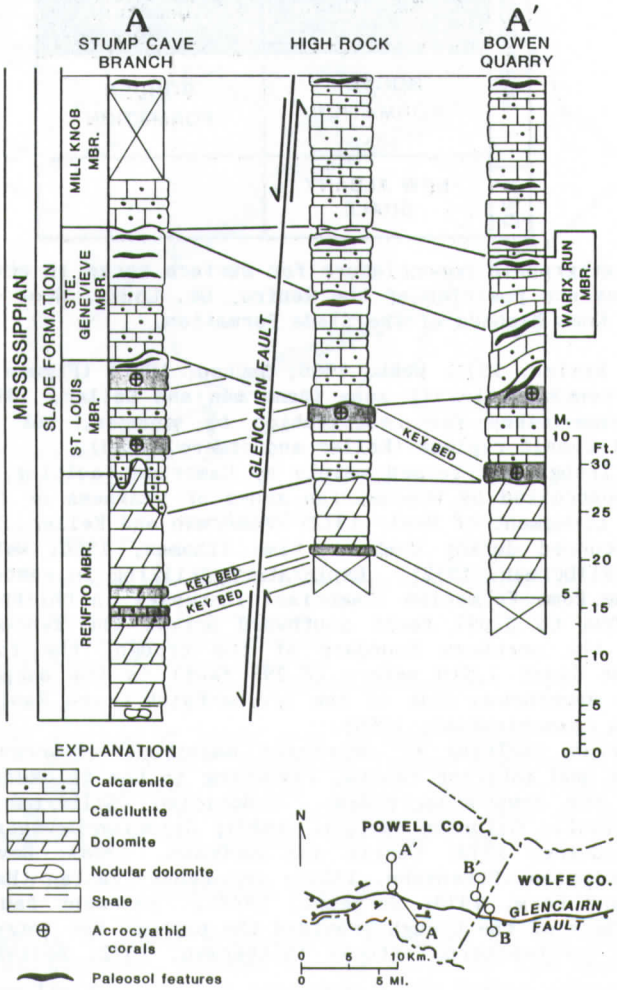


Figure 3. Northwest-southeast stratigraphic cross section across Glencairn Fault in western part of study area. Datum is top of Mill Knob Member.

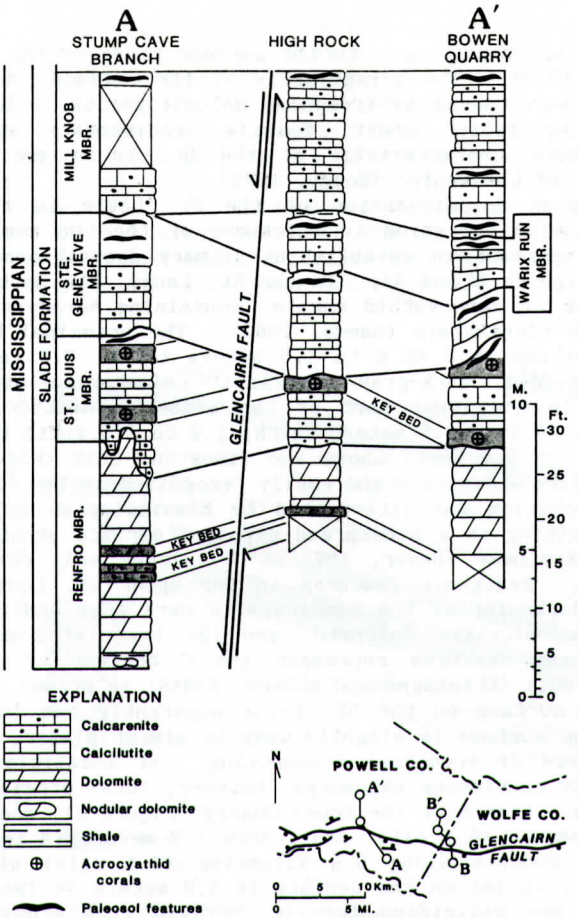


Figure 4. North-south stratigraphic cross section across Glencairn Fault in eastern part of study area. Datum is top of Mill Knob Member.

Genevieve, Warix Run, and Mill Knob Members (Ettensohn and others, 1984b). General lithologic characteristics and stratigraphic units of the lower Slade at seven sites are shown in Figures 3 and 4.

The Renfro and St. Louis Members in the basal part of the Slade Formation are the focus of this report. The Renfro is composed principally of light-gray, microcrystalline to very finely crystalline dolomite, in thick to thin beds. The rocks weather to yellowish orange or grayish orange. The overlying St. Louis Member mainly consists of light-olive-gray to light-gray cherty, bioclastic calcarenite, in thin to medium beds. The calcarenite is interbedded with calcilutite, argillaceous limestone, and shale. The St. Louis is very fossiliferous, containing brachiopods, bryozoans, pelmatozoans, echinoids, and corals. The Renfro is believed to be mainly supratidal to intertidal in origin; the St. Louis is composed of subtidal deposits (Dever, 1977; Moody, 1982).

The contact between dolomite of the Renfro Member and limestone of the overlying St. Louis Member is marked by a sharp but slightly wavy to very irregular surface, and a thin clay parting along the surface separates dolomite from limestone (Figures 3 and 4). The upper surface of the Renfro in its type area, south of the present study area, was considered to be a diastem (Weir and others, 1966). In the area of this study, the surface has been interpreted to be a minor unconformity between supratidal and subtidal

deposits, with narrow depressions in the surface representing littoral surge channels (Dever, 1973). The irregularity of the surface, however, is now believed to have been caused by irregular dolomitization of basal St. Louis limestones, because relict chert, fossils, sedimentary structures, and pockets of limestone characteristic of the St. Louis are found in the uppermost dolomite of the Renfro (Moody, 1982).

As the irregular dolomitization of the St. Louis in the study area results in substantial variation in thickness of the two members, key beds within them are important in establishing primary depositional thickness of the rock units (Figures 3 and 4). In the St. Louis, the most important key beds are two zones of acrocyathid corals, containing *Acrocyathus proliferus* and *A. floriformis floriformis* (Sando, 1983). The principal key bed in the Renfro is a widespread unit (0.6 to 1.5 meters thick; 2.0 to 4.8 feet) of thin- to medium-bedded, dark-gray bioclastic calcarenite with interbedded shale, occurring in the lower part of the member. Another key bed is a resistant dolomite (0.3 to 0.5 meters thick; 1.0 to 1.5 feet) that occurs 0.3 to 0.6 meter (1.0 to 2.0 feet) above the limestone unit (Figures 3 and 4). This resistant bed is widespread and easily recognized in the field.

St. Louis deposition was interrupted by Mississippian uplift along the Waverly Arch, resulting in a widespread exposure surface across northeastern and east-central Kentucky (Dever, 1977). A caliche soil developed on the exposed sediments. Pedogenic features in the upper St. Louis include (1) melanization, or darkening of the limestone to dark gray and dusky yellowish brown, (2) laminar micritic calcrete, and (3) brecciated calcilutite and calcisiltite. These features represent the C horizon of an erosionally truncated caliche soil (Ettensohn and others, 1984A; in press).

The subaerial surface on the St. Louis apparently had low relief. In most exposures, the surface is slightly wavy to almost planar, with little or no apparent evidence of erosional downcutting. At localities with closely spaced or laterally continuous exposures, however, local erosional relief can be determined. In a trench at the Bowen Quarry (Figure 3), sections measured less than 0.2 kilometer (0.1 mile) apart show 1.2 meters (4 feet) of relief. In discontinuous roadcuts along 0.6 kilometer (0.4 mile) of the Mountain Parkway (Figure 4), relief on the surface is 1.8 meters (6 feet). Clasts of chert, limestone, and silicified fossils, derived from erosion of the St. Louis, are present in the basal 2 to 68 centimeters (1 to 27 inches) of the overlying Ste. Genevieve Member; clasts locally occur 2.1 to 2.7 meters (7 to 9 feet) above the base.

The Ste. Genevieve, Warix Run, and Mill Knob Members of the lower Slade mainly consist of light-olive-gray to light-gray, bioclastic and oolitic calcarenite, with lesser amounts of light-olive-gray calcilutite (Figures 3 and 4). Detrital quartz sand is common in the Warix Run calcarenite. Remnant paleosols cap each of the units in the study area. These three members thin and pinch out eastward toward the axis of the Waverly Arch, which remained a positive feature following the uplift that interrupted St. Louis deposition (Dever, 1977).

RELATIONSHIP OF EROSION TO STRUCTURE

The pattern of intra-Mississippian erosion within the lower Slade Formation along the Glencairn Fault indicates that the fault was reactivated during Mississippian time. Earlier Paleozoic movement along the fault had resulted in an increase in thickness of Cambrian, Silurian-Devonian, and Devonian units on the downthrown side (Weaver and McGuire, 1973; Dillman, 1980; Webb, 1980). Erosional thinning of the St. Louis Member is most pronounced adjacent to the fault along the border of the upthrown (north) side. Reactivation occurred after uplift along the Waverly Arch interrupted St. Louis deposition and created an extensive exposure surface, but before St. Genevieve deposition.

The pattern of erosion, outlined below, suggests that displacement during

Mississippian reactivation was primarily upward and was accompanied by slight northward tilting (0.7 meter per kilometer; 4 feet per mile) of the upthrown block (Figure 5). Northward tilting of Renfro and St. Louis strata also is suggested by the thickness of sediments that subsequently accumulated during deposition of the Ste. Genevieve, Warix Run, and Mill Knob Members (Figure 4). Progradation at the end of Mill Knob deposition left a subaerial surface of low relief, and a useful stratigraphic datum, capping this sequence of sediments (Dever, 1977). The present dip of strata along the north side of the Glencairn Fault in the study area commonly is northward at a rate of 15 to 24 meters per kilometer (80 to 125 feet per mile) for a distance of 0.7 to 1.3 kilometers (0.4 to 0.8 mile) north of the fault (Weir, 1974a, 1974b), reflecting the effects of post-Misasissippian fault movement. Northward away from the fault, strata generally dip east-southeastward.

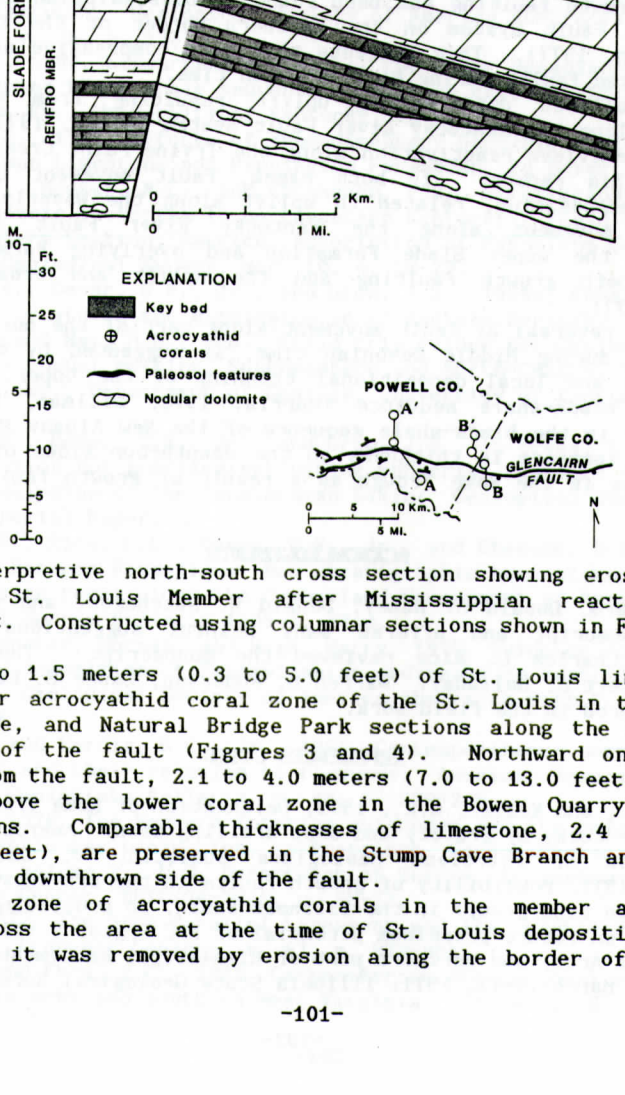


Figure 5. Interpretive north-south cross section showing erosional surface developed on St. Louis Member after Mississippian reactivation along Glencairn Fault. Constructed using columnar sections shown in Figure 4.

Only 0.1 to 1.5 meters (0.3 to 5.0 feet) of St. Louis limestone remain above the lower acrocyathid coral zone of the St. Louis in the High Rock, Mill Creek Mine, and Natural Bridge Park sections along the border of the upthrown side of the fault (Figures 3 and 4). Northward on the upthrown block, away from the fault, 2.1 to 4.0 meters (7.0 to 13.0 feet) of limestone are present above the lower coral zone in the Bowen Quarry and Mountain Parkway sections. Comparable thicknesses of limestone, 2.4 to 3.5 meters (8.0 to 11.5 feet), are preserved in the Stump Cave Branch and Middle Fork sections on the downthrown side of the fault.

The upper zone of acrocyathid corals in the member apparently was continuous across the area at the time of St. Louis deposition, but after fault movement it was removed by erosion along the border of the upthrown

side. The upper coral zone is preserved only on the downthrown side of the fault and in sections on the upthrown block that are some distance away from the fault (Figures 3, 4, and 5).

The caliche soil that developed on the St. Louis during subaerial exposure was partly eroded across the entire area, but deepest erosion was along the Glencairn Fault. The upper 0.3 to 0.9 meter (1.0 to 3.0 feet) of the St. Louis on the downthrown side of the fault and in sections some distance away from it on the upthrown block contains pedogenic features such as melanized (darkened) limestone, laminar calcrete, and brecciated limestone. Only 2 to 20 centimeters (1 to 8 inches) of similarily altered limestone (melanized and locally, slightly brecciated) remain in the sections along the border of the upthrown side.

DISCUSSION

The erosion that resulted from movement along the Irvine-Paint Creek Fault System in Mississippian time contrasts strongly with the dominance of extensional growth faulting and depositional thickening of units during earlier and later Paleozoic fault movements in the Rome Trough, but it is not an isolated event. Evidence for both differential uplift, followed by erosion, and growth faulting has been found in Mississippian units along the Kentucky River Fault System on the northern border of the trough (Dever, 1977; Ettensohn, 1977). This suggests that both compressive and extensional stresses affected faults during Mississippian time.

Erosion followed differential uplift resulting from pre-Warix Run reactivation along the Kentucky River Fault System (Dever, 1977), as well as the pre-Ste. Genevieve reactivation along the Irvine-Paint Creek Fault System outlined in this report. In both cases, fault movement coincided with periods of nondeposition, related to uplift along the Waverly Arch. Later Mississippian movement along the Kentucky River Fault System during deposition of the upper Slade Formation and overlying Paragon Formation resulted in both growth faulting and the uplift and erosion of units (Ettensohn, 1977).

An earlier reversal of fault movement along part of the northern boundary of the trough during Middle Devonian time, is suggested by the erosion of Silurian units and local depositional thinning of the Upper Devonian-Lower Mississippian black-shale sequence (Harris, 1978; Dillman, 1980). Upper Devonian units in the black-shale sequence of the New Albany and Ohio Shales more commonly increase in thickness on the downthrown sides of bounding and interior faults in the Rome Trough as a result of growth faulting (Dillman, 1980).

ACKNOWLEDGEMENTS

John C. Ferm, Donald C. Haney, Donald W. Hutcheson, and John D. Kiefer read the manuscript and offered many helpful suggestions. Robert C. McDowell and Charles L. Rice reviewed the manuscript. The figures were drafted by Robert C. Holladay. Warren H. Anderson, Lance S. Barron, and Jack R. Moody assisted in the field work.

REFERENCES CITED

Ammerman, M.L., and Keller, G.R., 1979, Delineation of Rome Trough in eastern Kentucky by gravity and deep drilling data: American Association of Petroleum Geologists Bulletin, v. 63, p. 341-353.

Avila, John, 1971, Possibility of growth faults in the pre-Knox interval (Montevallo Supergroup) in the Illinois Basin, *in* proceedings of symposium on future petroleum potential of NPC Region 9 (Illinois Basin, Cincinnati Arch, and northern part of Mississippi Embayment), Champaign, Illinois, March 11-12, 1971: Illinois State Geological Survey Illinois

Petroleum 95, p. 50-57.
Bolivar, S.L., 1982, Kimberlite of Elliott County, Kentucky: Kentucky Geological Survey, ser. 11, Thesis Series 2, 37 p.
Currie, M.T., and MacQuown, W.C., 1984, Subsurface stratigraphy of the Corniferous (Silurian-Devonian) of eastern Kentucky, *in* Luther, M.K., ed., Proceedings of the technical sessions, Kentucky Oil and Gas Association forty-fifth annual meeting, June 10-12, 1981: Kentucky Geological Survey, ser. 11, Special Publication 11, p. 1-21.
Dever, G.R., Jr., 1973, Stratigraphic relationships in the lower and middle Newman Limestone (Mississippian), east-central and northeastern Kentucky (M.S. thesis): Lexington, Ky., University of Kentucky, 121 p.; 1980, Kentucky Geological Survey, ser. 11, Thesis Series 1, 49 p.
Dever, G.R., Jr., 1977, The lower Newman Limestone: Stratigraphic evidence for Late Mississippian tectonic activity, *in* Dever, G.R., Jr., Hoge, H.P., Hester, N.C., and Ettensohn, F.R., Stratigraphic evidence for late Paleozoic tectonism in northeastern Kentucky. Field trip guidebook, Eastern Section, American Association of Petroleum Geologists: Kentucky Geological Survey, ser. 10, p. 8-18.
Dever, G.R., Jr., Ettensohn, F.R., and Haney, D.C., 1984, Relationship between the Kentucky River Fault System and Paleozoic sedimentation in eastern Kentucky: Geological Society of America Abstracts with Programs, v. 16, p. 133.
Dillman, S.B., 1980, Subsurface geology of the Upper Devonian-Lower Mississippian black-shale sequence in eastern Kentucky (M.S. thesis): Lexington, Ky., University of Kentucky, 72 p.
Ettensohn, F.R., 1977, Effects of synsedimentary tectonic activity on the upper Newman Limestone and Pennington Formation, *in* Dever, G.R., Jr., Hoge, H.P., Hester, N.C., and Ettensohn, F.R., Stratigraphic evidence for late Paleozoic tectonism in northeastern Kentucky. Field trip guidebook, Eastern Section, American Association of Petroleum Geologists: Kentucky Geological Survey, ser. 10, p. 18-29.
Ettensohn, F.R., Dever, G.R., Jr., and Grow, J.S., 1984a, Fossil soils and subaerial crusts in the Mississippian of eastern Kentucky, *in* Rast, Nicholas, and Hay, H.B., eds., Field trip guides for Geological Society of America annual meeting, Southeastern and North-Central Sections: Lexington, Ky., Department of Geology, University of Kentucky, p. 84-105.
Ettensohn, F.R., Dever, G.R., Jr., and Grow, J.S., in press, A paleosol interpretation for profiles exhibiting subaerial exposure "crusts" from the Mississippian of the Appalachian Basin: Geological Society of America Special Paper.
Ettensohn, F.R., Rice, C.L., Dever, G.R., Jr., and Chesnut, D.R., 1984b, Slade and Paragon Formations--New stratigraphic nomenclature for Mississippian rocks along the Cumberland Escarpment of Kentucky: U.S. Geological Survey Bulletin 1605-B, 37 p.
Haney, D.C., Hester, Norman, and Hoge, Harry, 1975, Evidence for fault-controlled deposition of Pennsylvanian sediments in central eastern Kentucky: Geological Society of America Abstracts with Programs, v. 7, p. 495-496.
Hardin, F.R., and Hardin, G.C., Jr., 1961, Contemporaneous normal faults of Gulf Coast and their relation to flexures: American Association of Petroleum Geologists Bulletin, v. 45, p. 238-248.
Harris, L.D., 1978, The Eastern Interior aulacogen and its relation to Devonian shale-gas production, *in* Preprints, Second Eastern Gas Shales Symposium, volume 2: U.S. Department of Energy, Morgantown Energy Technology Center, METC/SP-78/6, p. 55-72.
Heyl, A.V., 1972, The 38th Parallel Lineament and its relationship to ore deposits: Economic Geology, v. 67, p. 879-894.
Horne, J.C., and Ferm, J.C., 1978, Carboniferous depositional environments: Eastern Kentucky and southern West Virginia: Columbia, S.C., Department

of Geology, University of South Carolina, 151 p.
Keller, G.R., Lidiak, E.G., Hinze, W.J., and Braile, L.W., 1983, The role of rifting in the tectonic development of the Midcontinent, U.S.A., *in* Morgan, Paul, and Baker, B.H., eds., Processes of continental rifting: Tectonophysics, v. 94, p. 391-412.
MacQuown, W.C., and Pear, J.L., 1983, Regional and local geologic factors control Big Lime stratigraphy and exploration for petroleum in eastern Kentucky, *in* Luther, M.K., ed., Proceedings of the technical sessions, Kentucky Oil and Gas Association forty-fourth annual meeting, June 11-13, 1980: Kentucky Geological Survey, ser. 11, Special Publication 9, p. 1-20.
Maynor, G.K., 1984, Middle Paleozoic subsurface geology of the Rockcastle River Uplift area, southeastern Kentucky (M.S. thesis): Lexington, Ky., University of Kentucky, 126 p.
Miles, P.M., 1972, Notes on "Corniferous" production in eastern Kentucky, *in* Hutcheson, D.W., ed., Proceedings of the technical sessions, Kentucky Oil and Gas Association thirty-fourth and thirty-fifth annual meetings, 1970 and 1971: Kentucky Geological Survey, ser. 10, Special Publication 21, p. 46-49.
Moody, J.R., 1982, Dolomite bodies in the St. Louis Limestone Member of the Newman Limestone (Mississippian) of east-central Kentucky (M.S. thesis): Richmond, Ky., Eastern Kentucky University, 82 p.
Ocamb, R.D., 1961, Growth faults of south Louisiana: Transactions, Gulf Coast Association of Geological Societies, v. 11, p. 139-175.
Price, M.L., 1981, A regional study of the St. Peters Sandstone in eastern Kentucky, *in* Luther, M.K., ed., Proceedings of the technical sessions, Kentucky Oil and Gas Association thirty-eighth annual meeting, June 6-7, 1974: Kentucky Geological Survey, ser. 11, Special Publication 3, p. 1-19.
Sando, W.J., 1983, Revision of *Lithostrotionella* (Coelenterata, Rugosa) from the Carboniferous and Permian: U.S. Geological Survey Professional Paper 1247, 52 p.
Sergeant, R.E., 1979, Stratigraphic correlations of selected Alleghenian coals of the Princess Reserve District, Kentucky (M.S. thesis): Richmond, Ky., Eastern Kentucky University, 94 p.
Silberman, J.D., 1972, Cambro-Ordovician structural and stratigraphic relationships of a portion of the Rome Trough, *in* Hutcheson, D.W., ed., Proceedings of the technical sessions, Kentucky Oil and Gas Association thirty-fourth and thirty-fifth annual meetings, 1970 and 1971: Kentucky Geological Survey, ser. 10, Special Publication 21, p. 35-45.
Sutton, E.M., 1981, Deep exploration in eastern Kentucky by the SCLAW Group during the early seventies, *in* Luther, M.K., ed., Proceedings of the technical sessions, Kentucky Oil and Gas Association thirty-eighth annual meeting, June 6-7, 1974: Kentucky Geological Survey, ser. 11, Special Publication 3, p. 31-44.
Thomas, G.R., 1960, Geology of recent deep drilling in eastern Kentucky, *in* McGrain, Preston, and Crawford, T.J., eds., Proceedings of the technical sessions, Kentucky Oil and Gas Association twenty-fourth annual meeting, June 3, 1960: Kentucky Geological Survey, ser. 10, Special Publication 3. p. 10-28.
Watson, W.A., Jr., 1983, Boyd County Clinton gas field, *in* Luther, M.K., ed., Proceedings of the technical sessions, Kentucky Oil and Gas Association forty-third annual meeting, June 13-15, 1979: Kentucky Geological Survey, ser. 11, Special Publication 7, p. 31-43.
Weaver, O.D., and McGuire, W.H., 1973, Hunton exploration offers multiple objectives in eastern Kentucky: Oil and Gas Journal, v. 71, no. 5, p. 139-145.
Webb, E.J., 1969, Geologic history of the Cambrian System in the Appalachian Basin, *in* Rose, W.D., ed., Proceedings of the technical sessions, Kentucky Oil and Gas Association thirty-third annual meeting, June 5-6,

1969: Kentucky Geological Survey, ser. 10, Special Publication 18, p. 7-15.

Webb, E.J., 1980, Cambrian sedimentation and structural evolution of the Rome Trough in Kentucky (Ph.D. dissertation): Cincinnati, Ohio, University of Cincinnati, 98 p.

Weir, G.W., 1974a, Geologic map of the Stanton quadrangle, Powell and Estill Counties, Kentucky: U.S. Geological Survey Geologic Quadrangle Map GQ-1182.

Weir, G.W., 1974b, Geologic map of the Slade quadrangle, east-central Kentucky: U.S. Geological Survey Geologic Quadrangle Map GQ-1183.

Weir, G.W., Gualtieri, J.L., and Schlanger, S.O., 1966, Borden Formation (Mississippian) in south- and southeast-central Kentucky: U.S. Geological Survey Bulletin 1224-F, 38 p.

Woodward, H.P., 1961, Preliminary subsurface study of southeastern Appalachian Interior Plateau: American Association of Petroleum Geologists Bulletin, v. 45, p. 1634-1655.

MINERALOGY OF ULTRAMAFIC CHLORITE-AMPHIBOLE SCHISTS, INNER PIEDMONT BELT, SOUTH CAROLINA

RICHARD D. WARNER
VILLARD S. GRIFFIN, JR.
ROBERT C. HOOVER

Department of Earth Science
Clemson University
Clemson, SC 29634-1908

J. GREGORY BRYAN

Comap Exploration Services, Inc.
Lakewood, CO 80228

STEVEN H. POE

W. A. L., Inc.
Golden, CO 80403

JEFFREY C. STEINER

Department of Earth and
Planetary Sciences
City College (CUNY)
New York, NY 10031

ABSTRACT

Two small ultramafic schist bodies that occur in the Inner Piedmont belt of northwestern South Carolina contain abundant chlorite and calcic amphibole; olivine or serpentine or talc are also important constituents. Accessory minerals include anthophyllite, Cr-bearing magnetite, ilmenite, calcite and dolomite. Mineral compositions are compatible with amphibolite facies regional metamorphism. Equilibration temperatures estimated from coexisting magnetite-ilmenite mineral pairs are 570 - 580°C. Bulk chemical analyses indicate that the ultramafic bodies are relatively Al_2O_3-rich (~5-7 weight percent).

The Piedmont ultramafic rocks herein described seem to be distinctly different in chemical composition and mineralogy from dunitic-harzburgitic rocks typical of the Blue Ridge ultramafic province further to the west. The protolith of the Piedmont ultramafic rocks might have been a lherzolite or hornblende peridotite, or it might have been plagioclase peridotite dismembered from an ophiolite cumulate complex. Regional metamorphism caused a thorough recrystallization of the ultramafic rocks, including the growth of poikiloblastic olivine porphyroblasts. The ultramafic rocks acquired a schistose fabric and were converted to mineral assemblages (chlorite + tremolite + Ca-poor amphibole + olivine; chlorite + tremolite + Ca-poor amphibole + talc) in equilibrium with amphibolite facies P, T conditions.

INTRODUCTION

Numerous small bodies of ultramafic rocks are found in the southern Appalachians. A narrow, well-defined belt of ultramafic bodies occurs in the Blue Ridge province, while further east isolated ultramafic bodies are irregularly distributed throughout the Piedmont belt (Larrabee, 1966; Misra and Keller, 1978). Investigations of the Blue Ridge ultramafic bodies have shown that to be of the alpine type (e.g., Neathery, 1968; Bentley and Neathery, 1970; Condie and Madison, 1969; Hartley, 1973; Dribus and others, 1976; Honeycutt and Heimlich, 1980; Swanson, 1981). Mineralogically, the Appalachian alpine ultramafic rocks are characterized by olivine and associated pyroxene with high (~0.9) $Mg/(Mg+Fe^{2+})$, and by widely varying amounts of hydrous minerals, principally including serpentine, talc, chlorite, and/or amphibole.

The ultramafic rocks occurring in the Piedmont belt have received far less study than those of the Blue Ridge belt. A better understanding of the nature of the Piedmont ultramafic rocks is needed to allow regional comparisons with the better known Blue Ridge occurrences and in turn provide constraints on the geologic evolution of the southern Appalachians. A major

unresolved problem in this regard has to do with the possibility that at least some southern Appalachian ultramafic bodies may represent remnants of ophiolites that have been tectonically dismembered and subsequently metamorphosed during the southern Appalachian orogen (Misra and Keller, 1978).

Bryan and Griffin (1981) described the field relationships and provided preliminary summaries of the mineralogic characteristics of three small ultramafic bodies that outcrop in the South Carolina Piedmont. The bodies occur in terrane that has undergone amphibolite facies regional metamorphism. Based on their observations, Bryan and Griffin concluded that the Piedmont ultramafic rocks are compositionally distinct from alpine ultramafics occurring elsewhere in the southern Appalachian belt. In this paper we report electron microprobe mineral analyses and bulk rock chemical analyses for two of the ultramafic bodies mapped by Bryan and Griffin (1981). The principal goals of our study are to (1) characterize the mineral assemblage(s) developed during regional metamorphism of the ultramafic rocks and (2)elucidate the nature of the protolith(s) of these enigmatic bodies.

METHODS OF STUDY

Standard rectangular thin sections and circular 1" diameter polished microprobe sections were made of representative ultramafic samples. These, augmented by petrographic thin sections previously utilized by Bryan and Griffin (1981), were examined with transmitted and reflected light microscopy. Modes of the ultramafic rocks were determined using a Swift automatic point counter. Mineral compositions were obtained with an automated MAC-400 electron microprobe at NASA, Goddard Space Flight Center. The analyses were made using a beam size of ~ 2 μ m at an accelerating potential of 15 kv. The mineral analyses were corrected for differential matrix effects following the procedure of Bence and Albee (1968) and Albee and Ray (1970).

In addition to the microprobe work, representative samples from the Piedmont ultramafic bodies were analyzed by wavelength dispersive, X-ray fluorescence spectroscopy. Samples were analyzed as finely ground solids fused into disks. Cut chips sampled from various parts of the two ultramafic bodies were ground to a -100 mesh and then fused at 1000°C, with the addition of a lithium tetraborate/lithium carbonate flux. Qualitative and quantitative analyses of the elemental constituency were reported as oxides. Sample analyses were performed at W.A.L., Inc. a petrographic laboratory in Golden, Colorado, and at City College, City University of New York.

GEOLOGIC SETTING

The ultramafic rocks described in this paper occur in the Inner Piedmont Belt in northwestern South Carolina (Figure 1). One of the ultramafic bodies is located about 2 km north of Clemson in the Clemson 7.5 minute quadrangle. Its exposed dimensions are 133 m by 76 m, but part is covered by Lake Hartwell (Bryan and Griffin, 1981). The Clemson ultramafic body is surrounded by migmatitic amphibolite and granitoid gneiss, with mica schist occurring along the southern contact (Bryan and Griffin, 1981). The second ultramafic body is situated approximately 6 km east of Seneca in the Seneca 7.5 minute quadrangle. The Seneca ultramafic body is about 66 m long by 30 m wide, and occurs in a mica schist - biotite gneiss complex, with scattered migmatitic material (Bryan and Griffin, 1981). Quartz-rich, epidote-clinozoisite-bearing rocks are present along at least a portion of the contacts between both ultramafic bodies and the host rocks (Bryan and Griffin, 1981).

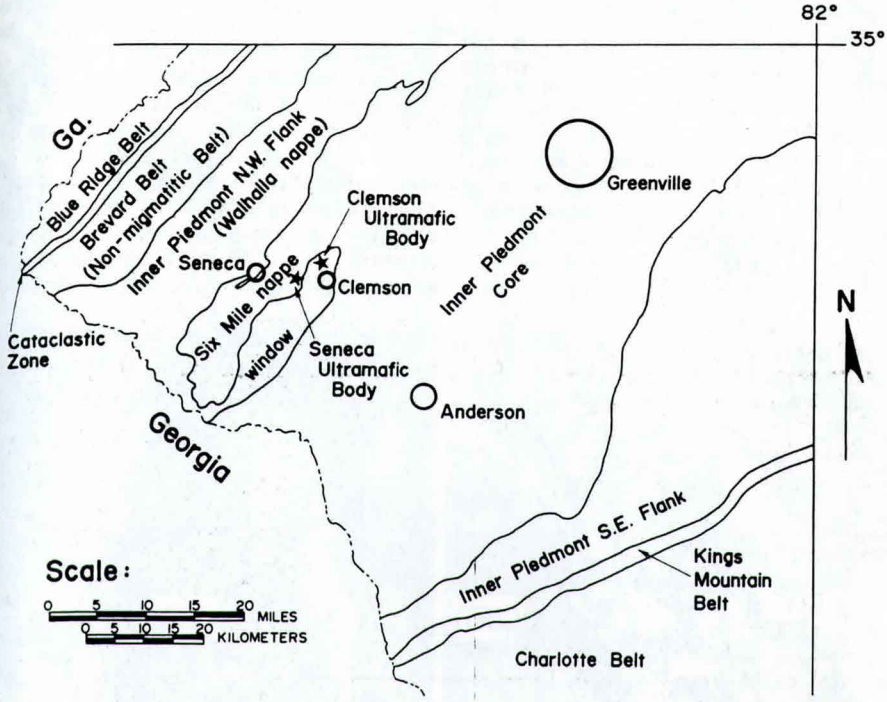


Figure 1. Geologic map of a portion of western South Carolina showing locations of the ultramafic bodies studied in this investigation. Boundaries between metamorphic belts are from Griffin (1973).

MINERALOGY

The principal minerals in the ultramafic bodies are calcic amphibole and chlorite; together these constitute from 53 to more than 95 percent of the rocks (Table 1). Olivine is an abundant mineral in most of the thin sections

Table 1. Modal abundances of ultramafic rocks.

	Clemson						
	WGC-1	WGC-1A	WGC-2	WGC-3	WGC-4	C-6	C-10
Amphibole	45.7	42.4	39.8	45.8	69.7	55.4	30.8
Chlorite	39.9	36.6	31.1	38.5	25.4	39.2	38.0
Olivine	9.9	16.1	26.1	11.9	--	--	--
Serpentine	--	--	--	Tr	--	--	--
Talc	--	--	--	--	2.6	1.5	26.0
Opaques	4.3	4.9	2.7	3.4	2.0	3.5	4.4
Carbonate	0.1	--	0.2	0.3	--	--	--
Other	0.2	--	--	0.1	0.3	0.4	0.8

	Seneca							
	WGS-1	WGS-2	WGS-3	WGS-4	WGS-5	S-1	S-5	A281
Amphibole	34.0	28.3	23.9	36.8	45.4	41.2	38.9	61.1
Chlorite	38.8	24.7	29.4	34.4	34.6	41.2	39.0	31.4
Olivine	21.2	--	--	--	10.4	13.5	15.0	--
Serpentine	--	40.7	40.4	21.1	2.6	Tr	Tr	--
Talc	--	0.4	--	1.1	--	--	0.3	7.0
Opaques	3.6	4.6	5.4	5.9	3.5	3.5	5.2	0.5
Carbonate	2.0	--	--	--	1.3	--	1.1	--
Other	0.5	1.4	0.8	0.8	2.1	0.5	0.4	--

examined, comprising as much as 26 percent modally. A few samples lack olivine, and in these serpentine or talc is an important constituent. Serpentine principally occurs in samples of the Seneca ultramafic body, whereas talc is more abundant in the Clemson ultramafic samples. In addition, the ultramafic rocks contain minor amounts of Ca-poor amphibole, generally 2 - 5 volume percent opaque minerals (principally magnetite but including ilmenite and sulfide minerals) and variable amounts (up to 2 percent modally) of dolomite and calcite.

The ultramafic rocks have a schistose texture. This texture is imparted primarily by chlorite, which occurs as thin laths or sprays of needles, often in felty aggregates (Figure 2A). The chlorite is pleochroic from nearly colorless to pale green. Amphibole is present as idioblastic crystals, often bounded by {110} faces (Figure 2B), as elongate laths up to 4 mm long, as

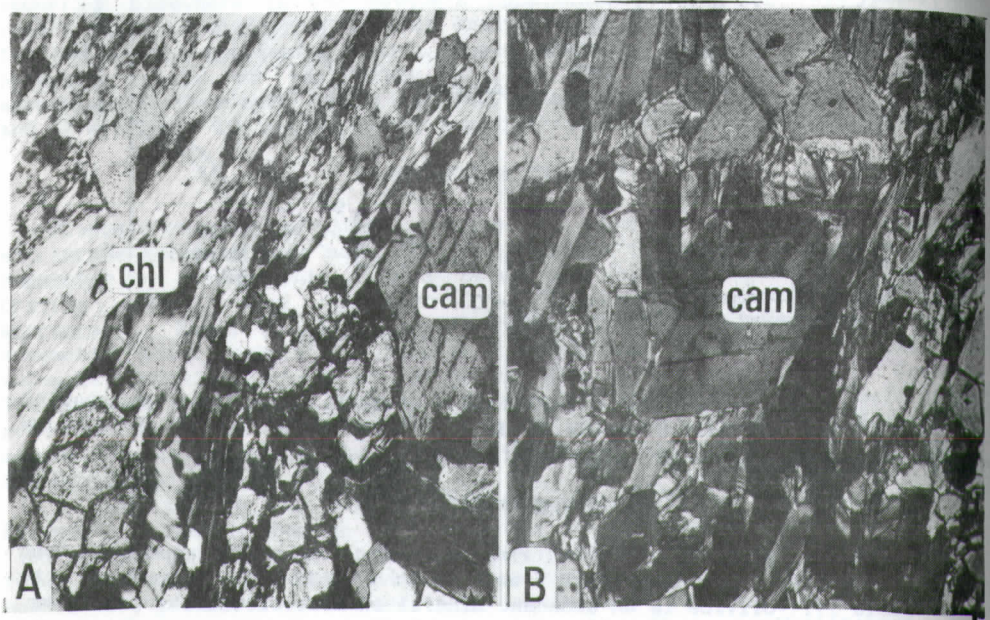


Figure 2. Photomicrographs (transmitted light, X nicols) of ultramafic rock samples. A. Schistose texture imparted by oriented chlorite laths, WGC-1. Long dimension: 0.8 mm. B. Idioblastic calcic amphibole crystal, bounded by {110} faces, in matrix of fine-grained chlorite plus calcic amphibole, WGC-3. Long dimension: 0.8 mm. Abbreviations: cam - calcic amphibole; chl - chlorite.

broad, irregular plates, and in clusters of medium-to fine- grained crystals, some of which appear to be polygonized remnants of larger crystals. The lath- shaped amphibole crystals are for the most part randomly oriented with respect to the foliation. In most thin sections the amphibole is sensibly colorless, but in a few thin sections it is pale green and pleochroic. Olivine is by far the coarsest-grained mineral in the ultramafic schists. It occurs in irregular crystals (Figure 3A) which tend to be elongated more or less parallel to the foliation, and are as much as 2 cm long by 0.5 cm wide. Frequently, the olivine porphyroblasts exhibit sieve texture, with abundant inclusions of amphibole, chlorite and opaques. This feature is most prominent in the Clemson ultramafic samples. The inclusions are of similar orientation to that of these minerals elsewhere in the rock. Some olivine crystals in both ultramafic bodies (particularly the Seneca body) are polygonized. The habit of serpentine is very similar to that of chlorite,

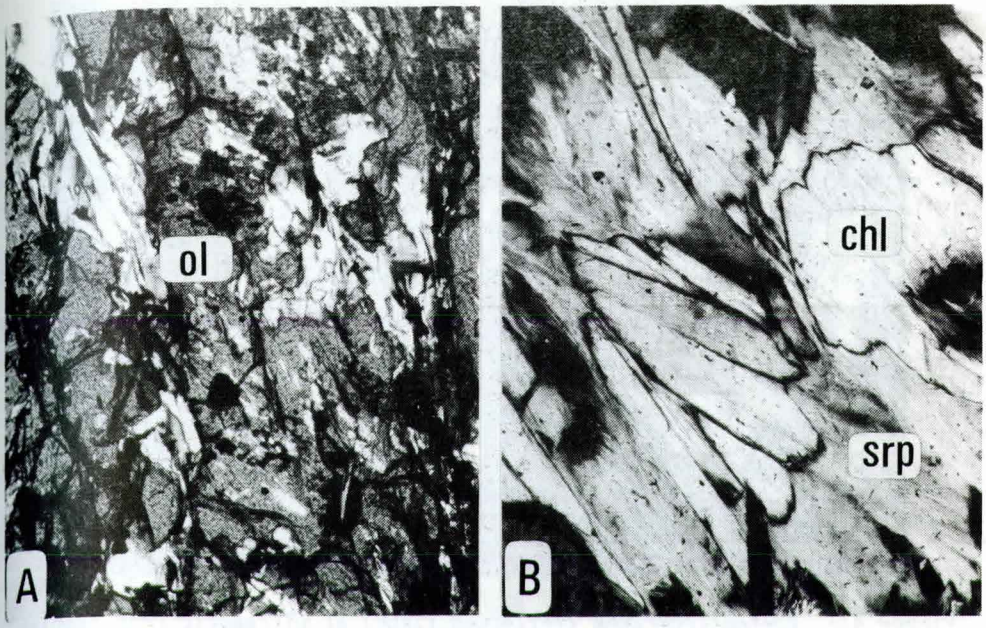


Figure 3. Photomicrographs (transmitted light, X nicols) of ultramafic rock samples. A. Olivine porphyroblast, WGC-2. Field of view consists of single olivine crystal with poikiloblastic texture. Inclusions are chlorite (laths), calcic amphibole and magnetite (opaque). Long dimension: 0.8 mm. B. Chlorite laths surrounded by serpentine, WGS-2. Both minerals are in optical continuity. Long dimension: 0.8 mm. Abbreviations: chl - chlorite, ol - olivine, srp - serpentine.

with which mineral it is usually associated. Laths of serpentine characteristically appear to wrap around and are in similar optical orientation with chlorite laths (Figure 3B). Talc, when present, occurs as laths or broad flakes; the crystals are often in random orientation. The talc is usually associated with amphibole, which it appears to replace. Magnetite and subordinate ilmenite are present as anhedral crystals about 0.05 to 0.5 mm across; in a number of instances the two minerals occur together (Figure 4A). In serpentine-rich samples (e.g., WGS-2) and also in WGC-4 (talc-bearing), much of the magnetite exhibits abundant fine-scale exsolution lamellae, presumably of hematite (Figure 4B). Magnetites in the olivine-bearing rocks are homogeneous throughout. In the latter rocks magnetite is most abundant in olivine-rich areas and commonly is included within olivine; some irregular, platy amphibole crystals are also clouded by myriads of tiny magnetite inclusions. Thin stringers of magnetite, associated with serpentine and/or chlorite, are frequently dispersed along fractures in olivine crystals and have formed by alteration of olivine. Carbonate, both dolomite and calcite, is present as scattered anhedral crystals 0.5 - 1 mm wide.

Compositions of minerals in the two ultramafic bodies are summarized in Tables 2 and 3. The minerals present show little grain to grain compositional variation in individual samples; the analyses shown in Tables 2 and 3 represent average mineral compositions in a given thin section. Analyses of the calcic amphibole reveal it to be an Al-bearing tremolite containing several weight percent Al_2O_3. Other important aspects of the mineral chemistry of the ultramafic schists include: 1) chlorite is Al-rich ($\geqslant$ 17.5 weight percent Al_2O_3); 2) serpentine, chlorite and calcic amphibole have the highest Mg' (= $Mg/(Mg + Fe^{2+})$) of the ultramafic minerals analyzed,

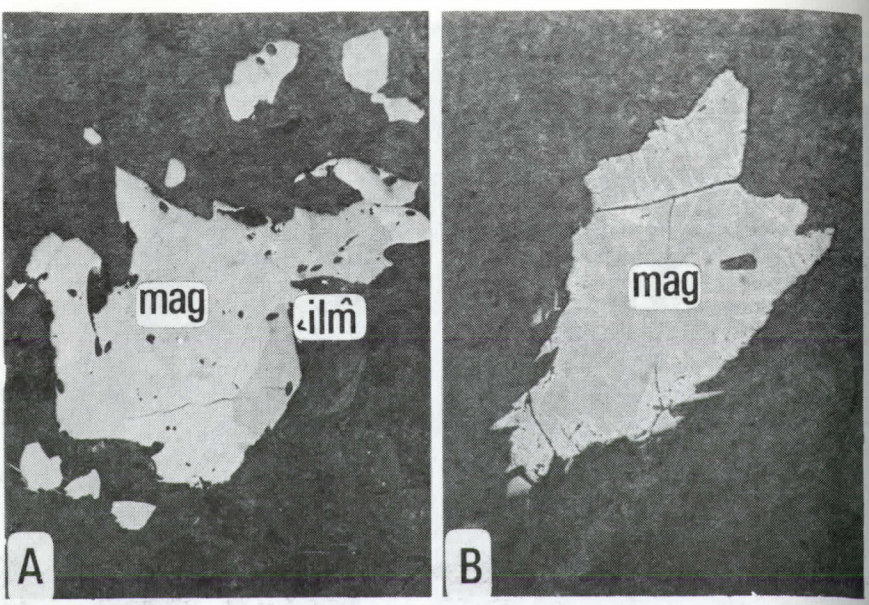


Figure 4. Photomicrographs (reflected light) of ultramafic rock samples. A. Composite magnetite - ilmenite grain, WGS-1. Long dimension: 0.6 mm. B. Magnetite with exsolution lamellae of hematite, WGS-3. Long dimension: 0.6 mm. Abbreviations: ilm - ilmenite; mag - magnetite.

Table 2. Representative electron microprobe analyses (wt. %) of minerals from Clemson ultramafic body.

	calcic amphibole			Ca-poor amphibole		chlorite		olivine		magnetite*		ilmenite*
	1	2	3	4	5	6	7	8	9	10	11	12
SiO_2	53.28	54.18	53.54	58.69	57.41	29.87	29.92	38.13	37.83			
TiO_2	0.16	0.14				0.06	0.05			0.30	0.52	49.15
Al_2O_3	4.29	3.93		0.14		20.63	19.22			0.20	0.20	0.19
Cr_2O_3	0.14	0.09				0.33	0.26			3.14	2.91	0.11
Fe_2O_3										66.00	65.61	8.00
FeO	5.77	5.20	5.87	13.02	13.47	7.36	7.39	21.70	22.57	29.90	30.40	37.86
MnO	0.17	0.20				0,04	0.05			0.08	0.05	1.36
MgO	21.50	21.73	21.64	26.21	26.16	30.05	31.22	39.46	39.36	1.06	0.85	2.75
CaO	12.13	12.34	12.24	0.82	0.63	0.08	0.14	0.01	0.06			
Na_2O	0.57	0.39				0.04	0.00					
Total	98.01	98.20	93.29	98.88	97.67	88.46	88.25	99.30	99.62	100.68	100.54	99.42
Number of oxygens =	23	23	23	23	23	14	14	4	4	4	4	3
Si	7.422	7.500	7.825	8.012	7.964	2.830	2.844	0.995	0.984			
Ti	0.016	0.012				0.002	0.002			0.009	0.015	0.921
Al	0.703	0.638		0.020		2.302	2.154			0.009	0.009	0.006
Cr	0.012	0.008				0.022	0.020			0.094	0.087	0.002
Fe^{3+}										1.880	1.874	0.150
Fe^{2+}	0.670	0.601	0.720	1.482	1.558	0.581	0.586	0.473	0.494	0.946	0.965	0.789
Mn	0.016	0.020				0.002	0.002			0.003	0.002	0.029
Mg	4.463	4.485	4.716	5.332	5.412	4.243	4.426	1.535	1.535	0.060	0.048	0.104
Ca	1.811	1.828	1.914	0.120	0.092	0.005	0.011	0.000	0.001			
Na	0.151	0.105				0.005	0.000					
Sum	15.264	15.197	15.175	14.966	15.026	9.992	10.045	3.003	3.014	3,001	3.000	2.001
$Mg/Mg+Fe^{2+}$	0.87	0.88	0.87	0.78	0.78	0.88	0.88	0.76	0.76			

*Fe_2O_3 and FeO calculated on the basis of oxide stoichiometry

Columns: 1 - average of 10 analyses, WGC-2; 2 - average of 10 analyses, WGC-3; 3 - average of 3 analyses, WGC-4; 4 - single analysis, WGC-1; 5 - average of 3 analyses, WGC-4; 6- average of 6 analyses, WGC-2; 7 - average of 8 analyses, WGC-3; 8 - average of 10 analyses, WGC-1; 9 - average of 6 analyses, WGC-3; 10 - average of 7 analyses, WGC-1; 11 - average of 6 analyses, WGC-3; 12 - average of 3 analyses, WGC-1.

with olivine and Ca-poor amphibole (anthophyllite) being about 10 mol % richer in Fe^{2+}; and 3) magnetite in both ultramafic bodies contains appreciable Cr (about 3 weight percent Cr_2O_3 on the average).

Comparatively, silicate minerals in the Seneca ultramafic body are more magnesian: Mg' in both chlorite and calcic amphibole is 0.91 in the Seneca samples vs about 0.88 in th Clemson samples, while Mg' of anthophyllite is 0.81 vs 0.78 and olivine is Fo_{82-85} in the Seneca ultramafic schist vs Fo_{76} in the Clemson body. Chlorite and tremolite are distinctly more Al-rich in the Clemson ultramafic rocks.

Table 3. Representative electron microprobe analyses (wt. %) of minerals from Seneca ultramafic body.

	calcic amphibole			Ca-poor amphibole	chlorite	serpentine		olivine		magnetite*	ilmenite*
	1	2	3	4	5	6	7	8	9	10	11
SiO_2	56.59	55.56	56.34	56.40	31.99	43.30	40.37	39.45	40.07		
TiO_2	0.12				0.10					0.18	49.25
Al_2O_3	2.17			0.13	17.55					0.19	0.18
Cr_2O_3										2.81	0.16
Fe_2O_3										66.81	8.65
FeO	3.94	3.72	4.14	11.58	5.74	7.04	4.41	16.86	14.75	29.70	35.93
MnO	0.17				0.03					0.08	2.02
MgO	23.04	21.97	23.26	27.47	32.21	37.90	40.79	43.61	46.29	1.15	3.54
CaO	12.48	12.81	13.31	0.91	0.07	0.03	0.05	0.04	0.08		
Na_2O	0.33				0.00						
K_2O	0.06				0.00						
Total	98.90	94.06	97.05	96.49	87.69	88.27	85.62	99.96	101.19	100.92	99.73
Number of oxygens =	23	23	23	23	14	7	7	4	4	4	3
Si	7.713	7.956	7.854	7.875	3.025	2.035	1.945	0.998	0.992		
Ti	0.012				0.007					0.005	0.915
Al	0.349			0.020	1.956					0.008	0.005
Cr										0.084	0.003
Fe^{3+}										1.897	0.161
Fe^{2+}	0.449	0.447	0.486	1.352	0.454	0.277	0.177	0.356	0.305	0.937	0.743
Mn	0.020				0.002					0.002	0.042
Mg	4.680	4.688	4.831	5.716	4.540	2.654	2.930	1.645	1.708	0.065	0.130
Ca	1.822	1.961	1.984	0.135	0.007	0.001	0.003	0.000	0.002		
Na	0.087				0.000						
K	0.011				0.000						
Sum	15.143	15.052	15.155	15.098	9.991	4.967	5.055	2.999	3.007	2.998	1.999
$Mg/Mg+Fe^{2+}$	0.91	0.91	0.91	0.81	0.91	0.91	0.94	0.82	0.85		

*Fe_2O_3 and FeO calculated on the basis of oxide stoichiometry

Columns: 1 - average of 12 analyses, WGS-1; 2 - average of 6 analyses, WGS-3; 3 - average of 2 analyses, WGS-4; 4 - single analysis, WGS-1; 5 - average of 13 analyses, WGS-1; 6 - average of 6 analyses, WGS-4; 7 - average of 2 analyses, WGS-5; 8 - average of 10 analyses, WGS-1; 9 - single analysis, WGS-5; 10 - average of 11 analyses, WGS-1; 11 - average of 3 analyses, WGS-1.

CHEMICAL COMPOSITION

Bulk chemical analyses of representative samples of the Clemson and Seneca ultramafic schists are reported in Table 4. The analyses of the Clemson samples, taken from widely separated portions of the body, show little variation in most major elements, with the exception that the analysis of sample WGC-4 contains about 2 weight percent higher SiO_2 and approximately 4 - 5 weight percent lower MgO than the other analyses. This is consistent with the modal mineralogy (Table 1), wherein WGC-4 lacks olivine and contains significantly more amphibole than WGC-1, WGC-1A, WGC-2 and WGC-3, imparting a higher Si/Mg ratio to WGC-4. The analyses from the Seneca ultramafic body are similar to those of the Clemson samples, with the most notable departure being lower CaO abundances (3.8 ± 0.9 weight percent in the Seneca vs 5.8 ± 0.4 weight percent in the Clemson samples). Al_2O_3 also is slightly lower (5.4 ± 1.1 vs 6.8 ± 0.9 weight percent) in the Seneca ultramafic samples and MgO and Fe_2O_3 each average about 2 weight percent higher.

Table 4. Chemical analyses of ultramafic schists.

	Clemson					Seneca			
	WGC-1	WGC-1A	WGC-2	WGC-3	WGC-4	WGS-1	WGS-2	WGS-4	WGS-5
SiO_2	42.34	42.47	42.74	42.50	44.76	42.59	40.68	41.59	41.68
TiO_2	0.30	0.29	0.34	0.31	0.32	0.31	0.28	0.30	0.24
Al_2O_3	6.98	7.68	6.44	6.39	6.43	6.54	5.02	4.81	5.11
Cr_2O_3	0.14	0.15	0.14	0.13	NA	0.15	NA	NA	NA
*Fe_2O_3	13.11	13.35	13.83	13.63	13.21	14.90	15.10	16.65	15.48
MnO_2	0.26	0.24	0.29	0.29	NA	0.30	NA	NA	NA
MgO	27.42	26.92	29.03	27.98	23.38	29.59	28.88	28.01	29.35
CaO	5.70	5.94	5.43	5.93	6.12	4.42	2.87	3.66	4.34
SrO	0.02	0.01	0.01	0.01	NA	0.03	NA	NA	NA
BaO	≤0.01	≤0.01	≤0.01	≤0.01	NA	≤0.01	NA	NA	NA
Na_2O	0.20	0.36	0.39	0.20	<0.1	0.29	<0.1	<0.1	<0.1
K_2O	0.01	0.01	0.01	0.01	≤0.01	0.03	≤0.01	≤0.01	≤0.01
P_2O_5	≤0.01	≤0.01	≤0.01	≤0.01	NA	≤0.01	NA	NA	NA
SO_3	0.22	0.24	0.11	0.07	NA	0.13	NA	NA	NA
H_2O^+	NA	NA	NA	NA	5.95	NA	7.79	5.36	4.27
Total	96.70	97.66	98.76	97.45	100.17	99.28	100.62	100.38	100.47

*All iron determined as Fe_2O_3

NA = not analyzed

As a whole, the rocks are clearly ultrabasic in composition, containing between 40 and 45 weight percent SiO_2. They contain 23 - 30 weight percent MgO and 13 - 16 weight percent Fe_2O_3. The ultramafic schists are also characterized by relatively high (for ultramafics) Al_2O_3, 4.8 - 7.7 weight percent, and moderate CaO. Alkali contents are low, particularly K_2O (0.01 - 0.03 weight percent). P_2O_5 is also extremely low, <0.01 weight percent. Finally, the ultramafic rocks contain 0.13 - 0.15 weight percent Cr_2O_3, about 0.3 weight percent TiO_2, and up to 7.8 weight percent H_2O.

CONDITIONS OF METAMORPHISM

Rocks of the Inner Piedmont belt have been subjected to amphibolite facies regional metamorphism. In the vicinity of the ultramafic schists herein described, conditions of the sillimanite isograd were attained. As noted, magnetite and ilmenite occasionally coexist (Figure 4A), permitting estimation of the temperatures at which the ultramafic rocks last equilibrated. We analyzed several coexisting magnetite - ilmenite pairs from both ultramafic bodies and, using the model of Spencer and Lindsley (1981) and the minor element correction scheme recommended by Stormer (1983), have calculated equilibration temperatures of 570 ± 20°C for the Clemson ultramafic rock and 580 ± 20°C for the Seneca body. Peak temperatures during metamorphism must have been at least this high.

The schistose texture (Figure 2A) and presence of chlorite, tremolite, talc and anthophyllite can be attributed to the regional metamorphism experienced by the ultramafic rocks. The origin of the olivine and serpentine in the rocks is more problematic. Olivine in altered ultramafic rocks is generally assumed to be relict from the protolith. However, the sieve texture (Figure 3A) exhibited by olivine in our samples leads us to believe that the olivine now present in the rocks crystallized during metamorphism. Similar occurrences in ultramafic rocks from western Australia that also have been metamorphosed to upper amphibolite facies conditions were described by Oliver and others (1972). Olivine in the Australian ultramafic rocks occurs as megacrysts up to 2 cm in diameter which contain abundant tremolite inclusions: the authors interpret the olivine as being of metamorphic origin. In their studies of alpine metamorphism of peridotite

rocks Evans (1977) and Trommsdorff and Evans (1974) concluded that a number of ultramafic bodies in the Alps were serpentinites prior to metamorphism and that olivine-bearing assemblages were developed during medium and high grade regional metamorphism of the serpentinized bodies. We note further that Carpenter and Phyfer (1969) have argued for a metamorphic origin of olivine in many southern Appalachian ultramafic bodies: they envisage that the bodies were emplaced as solid serpentine and during subsequent metamorphism were dehydrated to olivine-rich rocks. In the case of the ultramafic bodies under consideration here, we deem it likely that they were serpentinized prior to their emplacement, and that olivine crystallized during regional amphibolite grade metamorphism. The serpentine that is now present in some of the Seneca ultramafic rock samples we do not attribute to this prior serpentinization, however, but rather to retrograde metamorphism. The textural relation of the serpentine to chlorite, wherein serpentine appears to wrap around chlorite laths (Figure 3B) suggests that the serpentine is of late-stage derivation.

Textural evidence suggests that a significant amount of metamorphic crystallization was post-tectonic. For example, the abundance of randomly oriented tremolite and talc laths indicates that much of the amphibole and talc formed after the rocks had already acquired their foliation (Spry, 1969). The polygonization present in olivine (and tremolite) implies some post-deformational crystallization of both these minerals (Spry, 1969). The olivine porphyroblasts, which enclose chlorite and other minerals oriented parallel to the foliation, must also primarily be post-tectonic.

In summary, our textural observations indicate the following sequence of metamorphic crystallization. Chlorite and some of the amphibole were the earliest minerals to form, followed by the growth of olivine (often as porphyroblasts) and the randomly oriented laths of amphibole, with talc and serpentine being the last minerals to appear. Polygonization of olivine and tremolite was also late.

The mineral assemblages observed in the Clemson and Seneca ultramafic bodies can be compared to those delineated by Evans (1977) as being stable in ultramafic rock compositions at various metamorphic facies. According to Evans (1977), chlorite is stable over a wide range of metamorphic conditions, and is in equilibrium with the assemblages tremolite + olivine + Ca-poor amphibole and tremolite + talc + Ca-poor amphibole at amphibolite facies conditions (assemblages involving serpentine are also stable in the amphibolite facies but at a slightly lower grade). Compositions of the minerals present in these assemblages can be correlated with metamorphic grade. For example, Evans (1977) points out that the composition of chlorite coexisting with two other Mg-silicates becomes increasingly rich in Al with increasing metamorphic grade. The chlorite analyses (Tables 2 and 3) indicate between 1.96 and 2.3 Al per formula unit, close to the maximum of 2.4 Al at chlorite's breakdown temperature (Frost, 1975).

Evans (1977) also notes that the calcic amphibole in chlorite - olivine - Mg-silicate rocks systematically changes composition across the $Ca_2Mg_5Si_8O_{22}(OH)_2$ - $Mg_7Si_8O_{22}(OH)_2$ join, from about 96% tremolite in serpentine-bearing assemblages (stable at lower metamorphic grades) to less than 80% tremolite in higher temperature parageneses involving enstatite. The calcic amphibole in the Clemson and Seneca ultramafic schists contains 85 - 95% tremolite (average, 92%), within the range indicated by Evans (1977). Data of Trommsdorff and Evans (1974) show that Mg preferentially partitions into tremolite, antigorite and chlorite as opposed to olivine and anthophyllite; our mineral analyses are consistent with this as they indicate higher Mg' in calcic amphibole, serpentine and chlorite than in Ca- poor amphibole and olivine (Tables 2 and 3).

The composition of the spinel group mineral (here a Cr-bearing magnetite) coexisting with chlorite and two Mg-silicate minerals is also a function of metamorphic grade. Data of Evans and Frost (1975) show that in low-grade serpentinites Cr-magnetite is stable, with more chromiferous

compositions encountered with increasing metamorphic grade, and Al-rich spinels only appearing in high grade assemblages containing enstatite. The magnetite analyses (Tables 2 and 3) fall within the range (but near the low-Cr limit) indicated by Evans and Frost (1975) for Ca-poor amphibole + olivine + tremolite assemblages. Thus the mineral assemblages and mineral compositions observed in the Clemson and Seneca ultramafic schists are compatible with amphibolite facies metamorphism.

SPECULATIONS CONCERNING PARENT ROCKS

Textural and mineralogical data summarized in the preceding section indicate that the original texture and mineralogy of the ultramafic protolith have been obliterated by amphibolite facies regional metamorphism. An important question is whether or not the chemical composition of the ultramafic parent bodies also was changed as a consequence of the regional metamorphism. The juxtaposition of high-Mg, low-Si ultramafic rock with sialic country rock commonly results in the growth of ultramafic reaction zones, often nearly monomineralic, at their contact. The metasomatic diffusion of various elements is controlled by chemical potential gradients between the ultramafic and surrounding rocks (Brady, 1977; Sanford, 1982). Thus, the principal mass transfer generally involves Mg (lost from ultramafic body) and Si (added to ultramafic body). Migrations involving other components are not so readily characterized. For example, Sanford (1982) concluded that at low metamorphic grade Ca is introduced to the ultramafic, whereas at high grade Ca is removed from the ultramafic body. Also, where Al transfer has occurred, there is not universal agreement in the direction of Al movement: according to Jahns (1967), Al moves outward from ultramafic rock to host rock, but reaction zones at four localities studied by Sanford (1982) require an influx of Al from the country rock.

Certain aspects concerning the chemistry of the Clemson and Seneca ultramafic bodies lead us to believe that metasomatic exchange with the host rocks has not been extensive. First, there is no evidence that a significant amount of Si has been added. Rather, SiO_2 abundances of the ultramafic schist samples are uniformly low (Table 4), and are nearly identical to the worldwide average value for peridotite (Table 5, column 1) reported by Le Maitre (1976). Nor is there evidence for appreciable migration of Mg outward from the ultramafic rocks into the host rocks, as the quartz-epidote-clinozoisite border rocks are essentially devoid of Mg- rich silicates. Second, chemical potential gradients ought to have been such as to cause K, which is generally considered to be a mobile element during metamorphism, to diffuse inward from K-bearing neighboring rocks (mica schist, biotite gneiss, granitoid gneiss) into the much lower-K ultramafic rocks. Instead, the chemical analyses of the ultramafic rock samples show consistent, extremely low K_2O abundances (Table 4), indicating such transfer has not occurred. Third, as pointed out by Bryan and Griffin (1981), the quartz-epidote-clinozoisite border rocks are not particularly deficient in Al (or Ca) and are of relatively small volume (compared to that of the ultramafic bodies), so that significant addition of Al and Ca to the ultramafic rocks via their extraction from the immediately adjacent boundary zones does not appear to have occurred.

If, as we infer, the chemical composition of the ultramafic rocks has not been much modified during metamorphism, their high Al_2O_3 concentrations preclude protoliths dominantly composed of olivine and/or orthopyroxene (i.e., dunite, harzburgite, orthopyroxenite). SiO_2 and CaO are too low (and MgO much too high) for the protolith to have been a clinopyroxene-rich ultramafic such as a websterite, and alkalis (especially K_2O) are too low for the parent bodies to have been mica peridotite or kimberlite. Some ultramafic rocks that compositionally seem the most feasible candidates are listed in Table 5. The lherzolite composition (column 2) represents a worldwide average of 179 analyses (LeMaitre, 1976); with the exception of

Table 5. Composition of representative aluminous ultramafic rocks.

	Peridotite	Lherzolite	Hornblende Peridotite	Ophiolite Cumulate Complexes	
	1	2	3	4[a]	5[a]
SiO_2	42.26	42.52	42.43	41.85	42.69
TiO_2	0.63	0.42	0.96	0.28	0.10
Al_2O_3	4.23	4.11	5.50	6.82	7.82
Cr_2O_3	NG	NG	NG	0.60	0.13
Fe_2O_3	3.61	4.82	2.62	--	--
FeO	6.58	6.96	9.90	8.95	11.95
MnO	0.41	0.17	NG	0.14	0.20
MgO	31.24	28.37	30.79	36.73	28.46
CaO	5.05	5.32	4.78	4.21	8.02
Na_2O	0.49	0.55	0.81	0.16	0.40
K_2O	0.34	0.25	0.44	--	--
H_2O^+	3.91	1.07	1.69	--	--
P_2O_5	0.10	0.11	0.08	0.01	NG

[a]Analyses normalized to 100% after removal of H_2O, CO_2, and all iron recalculated as FeO

NG - not given

Columns: 1 - Average of 108 peridotite analyses (Le Maitre, 1976); 2 - Average of 179 lherzolite analyses (Le Maitre, 1976); 3 - Average of 5 hornblende peridotite analyses (Nockolds, 1954); 4 - Average chemical composition of feldspar dunite cumulates, Bay of Islands ophiolite, New Foundland (Coleman, 1977, based on earlier sources); 5 - Average chemical composition of cumulate gabbros, Semail ophiolite, Oman (Coleman, 1977, based on earlier sources).

Al_2O_3 (about 3 weight percent too low), it provides a close fit to the composition of the ultramafic schists (Table 4). The standard deviation given by Le Maitre (1976) for Al_2O_3 is 2.70 weight percent, indicating that some individual lherzolites have Al_2O_3 contents as high as those of the Clemson and Seneca ultramafic rocks. In terms of Al_2O_3, a somewhat better match would be hornblende peridotite (Table 5, column 3). The abundances of TiO_2, Na_2O, K_2O and P_2O_5 in the ultramafic schists are considerably below the average for hornblende peridotite, however.

It has been suggested that ultramafic rocks from the Piedmont of Wake County, North Carolina may be dismembered ophiolite fragments (Spence and Carpenter, 1976; Horton and others, 1985). According to Coleman (1977), an idealized ophiolite sequence consists of basal metamorphic peridotite (dominantly harzburgite or dunite and more or less serpentinized) overlain by cumulate complexes which grade upward from peridotitic to gabbroic rocks, with mafic sheeted dikes and mafic extrusives (often pillowed) as upper units. The Clemson and Seneca ultramafic rocks do not correspond in composition to ophiolite metamorphic peridotites, but reasonable compositional analogues are to be found among plagioclase-bearing peridotitic rocks from ophiolite cumulate zone complexes. The analyses given in Table 5 (columns 4, 5) are of feldspar dunite cumulate from the Transition Zone in the Bay of Islands ophiolite, New Foundland and of cumulate gabbro within the Semail ophiolite, Oman, respectively. Except for CaO (too high), the latter closely matches the ultramafic schist compositions. We therefore speculate that the protoliths of the Clemson- Seneca ultramafic bodies may have been fragments of plagioclase peridotite torn from an ophiolite cumulate complex.

Despite uncertainty about the exact nature of the ultramafic protolith, a

general chemical affinity to lherzolite, hornblende peridotite and/or plagioclase peridotite clearly marks these bodies as distinct from the dunites, harzburgites and pyroxenites typical of the Blue Ridge belt. The Blue Ridge ultramafics are much more olivine-rich and are characterized by more magnesian mineral compositions. For example, Swanson (1981) reported olivine compositions of Fo_{92-95} in the Day Book dunite of western North Carolina, while Carpenter and Phyfer (1975) found similar olivine compositions for other southern Appalachian ultramafic bodies. In contrast, olivine compositions in the two Piedmont bodies described herein are Fo_{76} and Fo_{82-85}. Another difference is in the composition of the spinel group mineral: whereas ultramafic rocks in the Blue Ridge belt contain chromite as a common constituent, we find instead Cr-bearing magnetite in both the Clemson and Seneca bodies.

Whatever the provenance of the ultramafic protoliths, the bodies were emplaced prior to (or during) amphibolite facies metamorphism that affected the Inner Piedmont belt. The mode of emplacement has largely been obscured by the subsequent metamorphism. The prevalence of hydrous minerals in the metamorphosed ultramafic bodies taken in conjunction with the observation that progressive alpine metamorphism consists primarily of a sequence of dehydration reactions (Evans, 1977), lead us to conjecture that the original ultramafic rocks were hydrated before their emplacement. That is, the protoliths may have been partly or wholly serpentinized, in keeping with the emplacement model elaborated by Carpenter and Phyfer (1969).

CONCLUSIONS

Our mineralogical-chemical investigation of two small metamorphosed ultramafic bodies situated in the Inner Piedmont belt of northwestern South Carolina leads us to conclude that the original parent bodies were relatively Al-rich and distinctly different in composition from dunitic- harzburgitic rocks typical of the Blue Ridge ultramafic province. We speculate that the ultramafic protolith might have been of lherzolite composition or a hornblende- or plagioclase-bearing peridotite. The ultramafic bodies may have been serpentinized prior to their emplacement. Subsequent to their emplacement, the ultramafic rocks were subjected to amphibolite facies regional metamorphism. The metamorphism caused a thorough recrystallization of the ultramafic rocks, whereby they acquired a schistose fabric and were converted to mineral assemblages in equilibrium with amphibolite facies P, T conditions. Equilbration temperatures estimated from coexisting magnetite-ilmenite mineral pairs are 570 -580°C.

ACKNOWLEDGEMENTS

The authors wish to thank Pete Wasilewski for providing access to the automated MAC-400 electron microprobe at NASA, Goddard Space Flight Center. The assistance of Chuck Gill in writing computer programs to obtain equilibration temperatures from coexisting magnetite-ilmenite mineral pairs is gratefully acknowledged. We wish also to thank Barbara Peters for typing the manuscript, and Dave Snipes and A. A. Roberts for their helpful editorial remarks. The manuscript has benefitted from constructive comments supplied by reviewer Sam Swanson.

REFERENCES

Albee, A.L., and Ray, L., 1970, Correction factors for electron probe microanalyses of silicates, oxides, carbonates, phosphates and sulfates: Anal. Chem., v. 42, p. 1408-1414.

Bence, A.E., and Albee, A.L., 1968, Empirical correction factors for the electron microanalysis of silicates and oxides: Jour. Geol., v. 76, p. 382-403.

Bentley, R.D., and Neathery, T.L., 1970, Geology of the Brevard fault zone and related rocks of the Inner Piedmont of Alabama: Alabama Geol. Society Guidebook, 8th Annual Field Trip, 119 p.

Brady, J.B., 1977, Metasomatic zones in metamorphic rock: Geochim. Cosmochim. Acta, v. 41, p. 113-126.

Bryan, J.G., and Griffin, V.S., Jr., 1981, Ultramafic bodies in the Piedmont of northwestern South Carolina: South Carolina Geol., v. 25, p. 69-78.

Carpenter, J.R., and Phyfer, D.W., 1969, Proposed origin for the alpine type ultramafics of the Appalachians (abs): Geol. Soc. Amer. Abs. with Programs, v. 1, p. 261-263.

Carpenter, J.R. and Phyfer, D.W., 1975, Olivine compositions from southern Appalachian ultramafics: Southeast. Geol., v. 16, p. 169-172.

Coleman, R.G., 1977, Ophiolites: Berlin, Springer - Verlag, 229 p.

Condie, K., and Madison, J., 1969, Composition and volume change accompanying progressive serpentinization of dunites from the Webster-Addie ultramafic body, North Carolina: Amer. Mineralogist, v. 54, p. 1173-1179.

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

Evans, B.W., 1977, Metamorphism of alpine peridotite and serpentinite: Ann. Rev. Earth Sci., v. 5, p. 397-447.

Evans, B.W., and Frost, B. R., 1975, Chrome-spinel in progressive metamorphism - a preliminary analysis. Geochim. Cosmochim. Acta, v. 39, p. 959-972.

Frost, B.R., 1975, Contact metamorphism of serpentinite, chloritic blackwall and rodingite at Paddy-Go-Easy Pass, Central Cascades, Washington: Jour. Petrol., v. 16, p. 272-313.

Griffin, V.S., Jr., 1973, Geology of the Old Pickens Quadrangle, South Carolina: S.C. State Devel. Bd., Div. of Geol., MS-18, 54 p. (+ 1:24,000 maps).

Hartley, M.E., 1973, Ultramafic and related rocks in the vicinity of Lake Chatuge: Georgia Geol. Survey Bull., v. 85, 61 p.

Honeycutt, F.M., and Heimlich, R.A., 1980 Petrology of the Balsam Gap dunite, Jackson County, North Carolina: Southeastern Geol., v. 21, p. 251-260.

Horton, J.W., Jr., Stoddard, E.F., Wylie, A.S., Jr., Moye, R.J., Jr. and Blake, D.E., 1985, Melange terrane in the eastern Piedmont, North Carolina (abs): Geol. Soc. Amer. Abs. with Programs, v. 17, p. 96.

Jahns, R.H., 1967, Serpentinites of the Roxbury district, Vermont: *in* Wyllie, P.J., ed., Ultramafic and related rocks, New York, John Wiley and Sons, p. 137-160.

Larrabee, D.M., 1966, Map showing distribution of ultramafic and intrusive mafic rocks from northern New Jersey to eastern Alabama: U.S. Geol. Survey, Misc. Geol. Inv. Map I-476, 1:500,000.

LeMaitre, R.W., 1976, The chemical variability of some common igneous rocks: Jour. Petrol., v. 17, p. 589-637.

Misra, K.C., and Keller, F.B., 1978, Ultramafic bodies in the southern Appalachians: A review: Amer. Jour. Sci., v. 278, p. 389-418.

Neathery, T.L., 1968, Talc and anthophyllite asbestos deposits in Tallapoosa and Chambers Counties, Alabama: Geol. Survey of Alabama, Bull., v. 90, 98 p.

Nockolds, S.R., 1954, Average chemical composition of some igneous rocks: Geol. Soc. America Bull., v. 65, p. 1007-1032.

Oliver, R.L, Nesbitt, R.W., Hausen, D.M., and Franzen, N., 1972, Metamorphic olivine in ultramafic rocks from western Australia: Contrib. Mineral. Petrol., v. 36, p. 335-342.

Sanford, R.F., 1982, Growth of ultramafic reaction zones in greenschist to amphibolite facies metamorphism: Amer. Jour. Sci., v. 282, p. 543-616.

Spence, W.H., and Carpenter, P.A., III, 1976, Metallogenic zonation and a model for the development of the Piedmont of eastern North Carolina (abs): Geol. Soc. Amer. Abs. with Programs, v. 8, p. 272-274.

Spencer, K.J., and Lindsley, D.H., 1981, A solution model for coexisting iron-titanium oxides: Amer. Mineralogist, v. 66, p. 1189-1201.

Spry, A., 1969, Metamorphic Textures: Oxford, Pergamon Press, 350 p.

Stormer, J.C., Jr., 1983, The effects of recalculation on estimates of temperature and oxygen fugacity from analyses of multicomponent iron-titanium oxides: Amer. Mineralogist, v. 68, p. 586-594.

Swanson, S.E., 1981, Mineralogy and petrology of the Day Book dunite and associated rocks, western North Carolina: Southeastern Geology, v. 22, p. 53-77.

Trommsdorff, V., and Evans, B.W., 1974, Alpine metamorphism of peridotitic rocks: Schweiz. Mineral. Petrol. Mitt., v. 54, p. 333-352.