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THE BIG CLIFTY (HARTSELLE) FORMATION (MISSISSIPPIAN) IN SOUTHEAST TENNESSEE, PETROLOGY, LITHOFACIES, AND ORIGIN

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ABSTRACT

Petrologic comparisons and lithofacies distributions in what is currently called the Hartselle Formation in southeast Tennessee suggest that the sandstones are not continuous with the type of Hartselle in Alabama. Such comparisons do suggest that the Tennessee Hartselle is similar to, and was continuous with, the Big Clifty member of the Golconda Formation of the Illinois Basin area hence we recommend reintroduction of the term Big Clifty Formation. In the study area (northeast of Chattanooga) the Big Clifty is dominated by carbonates (11% dolomite, 17% lime micrite and wackestone, 22% packstone, 14% fossiliferous and oolitic grainstones) with 17% arenite sandstones and 19% variously colored shales. The sandstones are very fine sand sized quartz arenites or very quartzose sublithic arenites ($Q_{92}F_0R_8$). This pulse of clastics into the otherwise carbonate sequence of the Mississippian occurred when the Michigan River delta extended southeast out of the Illinois Basin.

INTRODUCTION

The term Hartselle Formation (Smith, 1894) has been in use for the only clastic break within the Mississippian System of Tennessee since the work of Stearns (1963). During a study of the petrology and stratigraphy of this unit it became apparent that the use of the term Hartselle may be inappropriate. The purpose of this report is, therefore, twofold: (1) a description of the petrology and lithofacies distribution of this clastic unit as a basis for determining its origin, and (2) justification for the reintroduction of the term Big Clifty Formation for what is now called Hartselle.

The Big Clifty (Hartselle) Formation is Early Chesterin age (Figure 1). This clastic unit is widespread in the areas of Illinois, Indiana, Kentucky, Tennessee, Mississippi, Alabama, and Georgia and makes an easily recognizable stratigraphic marker in the otherwise continuous carbonate sequence between the Chattanooga Shale below and the terrigenous clastics of Pennsylvanian age above. The Hartselle Formation of Tennessee is related to the Big Clifty Sandstone (Norwood, 1876) of the Golconda Formation of Kentucky, Indiana, and Illinois (Peterson, 1956; McFarlan and Walker, 1956; Vail, 1959; Sable, 1979). This correlation was well understood by Stearns (1963, p. 9); however, he preferred the term Hartselle Formation because of established correlations with the Alabama Hartselle.

METHODS

Seven outcrops were measured and studied in detail with additional observations at six other locations (Figure 2). Detailed directions to the measured section are in Roberts (1980). Field units were determined largely on the basis of clastic content and sedimentary structures. Standard thin section preparation techniques were used and samples were stained with Alizarin Red and Potassium Ferricyanide (Lindholm and Finkleman, 1972). Carbonate percentages were determined by visual estimates of thin sections using Folk's (1962) component terms and by insoluble residue analysis. Clastics were point counted. Carbonates were classified according to Dunham (1962) and clastics according to Dott (1964).

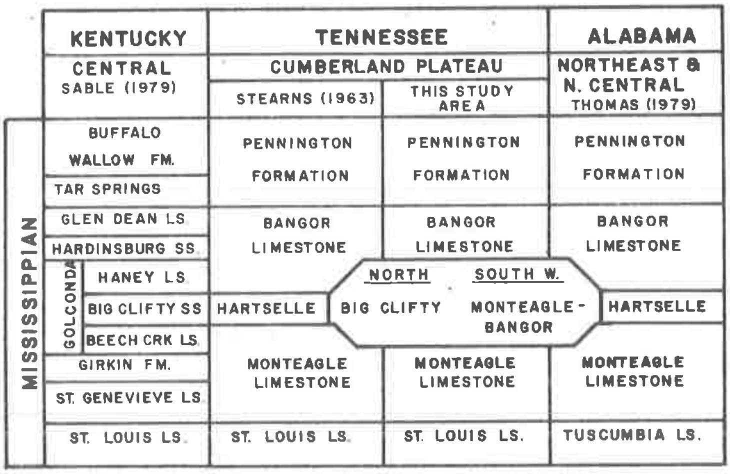


Figure 1. Mississippian stratigraphic terminology recommended in this paper. The Monteagle-Bangor undifferentiated was suggested by Thomas and Cramer (1979).

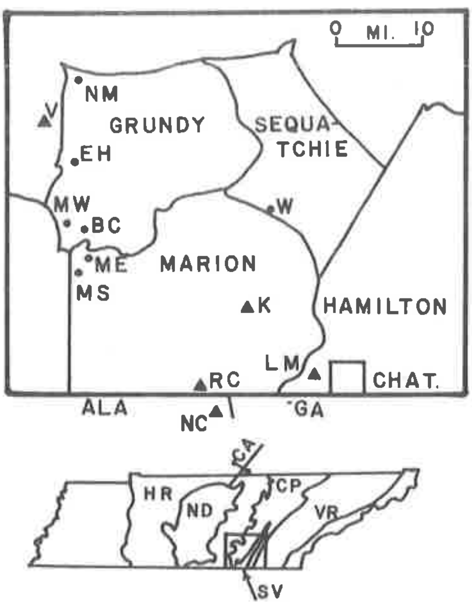


Figure 2. The study area is outlined in the southeast of the small map of Tennessee and is shown in more detail in the larger map above. Topographic and tectonic features mentioned in the text are shown on the small map; HR-Highland Rim; ND-Nashville Dome; CA-Cincinnati Arch; CP-Cumberland Plateau; VR-Valley and Ridge; SV-Sequatchi Valley. In the larger scale map the seven detailed sections are shown as black circles; NM-Nunley Mountain; EH-Elk Head; MW-Monteagle West; BC-Burrow Cove; ME-Monteagle East; MS-Monteagle South; W-Wilmouth. Uncollected locations are marked by triangles; LM-Lookout Mt.; NC-Norwood Cove; RC-Richard City; K-Kimble; V-Viola. Detailed directions to locations are in Roberts (1980).

RESULTS

Petrography

Petrographic examination of thin sections coupled with field data enables nine Big Clifty lithofacies to be distinguished in the study area. The lithofacies were identified

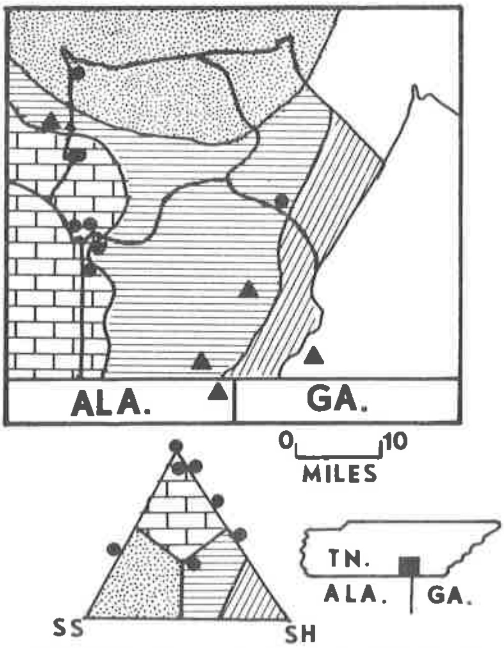


Figure 3. Lithofacies distribution map. Circles plotted on the composition triangle show the approximate lithology mix at the measured sections.

on the basis of general overall rock types, sedimentary structures, grain size, and grain types. The lithofacies are: (1) dolomicrite, (2) lime mudstone, (3) wackestone, (4) skeletal packstone, (5) intraclast packstone and wackestone, (6) skeletal grainstone, (7) arenaceous oolitic grainstone, (8) sandstones, and (9) shale. Detailed analysis can be found in Roberts (1980).

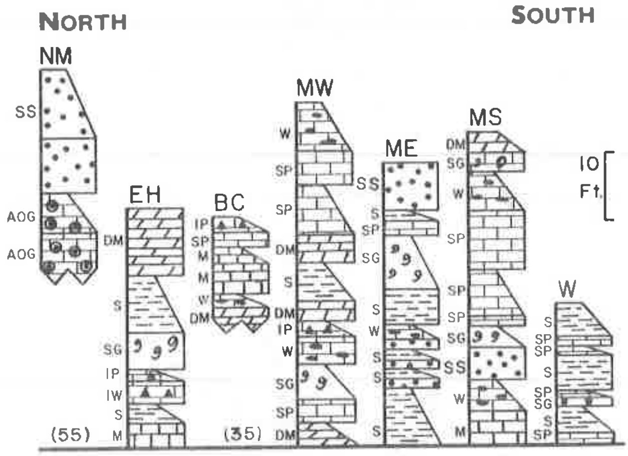
Dolomicrite forms 11% of the cumulative thickness of the measured sections and is most common toward the middle of the study area (locations EH, BC and MW; Figures 3, 4). Mudcrack and birdseye structures are common in addition to a few burrows. Pellets are absent but there are scattered brachiopod and echinoderm fragments and a small amount of quartz. Insoluble residues are high (average 28%, maximum 39%) and consist of a black hygroscopic clay.

Lime mudstone forms 7% of the cumulative thickness and is most common at EH and BC (figures 3, 4). Birdseye and mudcrack structures are not common. Fossils are similar to those in the dolomicrites but include sponge spicules and foraminifera; pellets are common. Phosphatic brachiopods are present at the MS section *(Orbiculodia?)*. Detrital quartz is common (up to 8%) and insoluble residues range up to 29% (average 16%).

Wackestone forms 10% of the cumulative measured section and is most common at locations MS and MW. A few birdseye structures are present. Allochems are normal marine fossils (as in the micrites) or pellets. Dolomite is common and some samples are dolomite wackestones. Up to 11% quartz was observed. Insoluble residues range up to 28% and average 16%.

Intraclast packstones and wackestones together form 5% of the total sequence and are more common toward the north. The poorly sorted intraclasts are buff to maroon dolomicrite chips that vary in size up to an extreme of 18 cm. Up to 87% dolomite and 2% quartz may be present. Cut and fill structures are rare.

Skeletal packstones form 17% of the measured section and are more common to the south and west. Echinoderm fragments are more abundant than brachiopods, bryozoans, and gastropods. Samples have an overcrowded appearance and grain contacts shown pressure solution (Figure 5a). Dolomite forms up to 17%, and quartz up to 4% of the volume. Insoluble residues range from 2% to 18% (average 10%) and are concentrated in the micrite matrix. The rocks are massive bedded and lack sedimentary structures.



	NM	EH	BC	MW	ME	MS	W
THICK. ft	28+	35	14+	49	41	45	20
%SS	61	0	0	0	27	10	0
%SH	0	31	0	11	42	0	55
%LS	39	41	90	69	31	83	45
%DOL	0	28	10	20	0	7	0

Figure 4. Stratigraphic section along north-south line except for W - Wilmouth which is 20 miles to the east. DM - dolomicrite; M - lime mudstone; W - wackestone; I - intraclast; SP - skeletal packstone; SG - skeletal grainstone; AOG - arenaceous oolitic grainstone; S - shale; SS - sandstone. Thicknesses of the covered intervals at NM - Nunley Mt. and BC - Burrow Cove were assigned on the basis of the first appearance of Monteagle lithologies below the covered zone. Estimated thicknesses are in parenthesis.

Skeletal grainstone forms 9% of the Big Clifty. Echinoderm fragments are abundant whereas pellets are rare. Minor detrital quartz, up to 6% dolomite, and common chert (up to 11%) were observed. Insoluble residues average 5% and are dominated by silt quartz grains. As in the skeletal packstones an over compacted fabric is common.

Arenaceous oolitic grainstone forms 5% of the Big Clifty and is present only section NM. These rocks resemble sandstone in weathered outcrops with abundant assymetrical ripples and tangential cross beds. Together with overlying sandstones this lithology forms a resistant bench easily recognized on topographic maps. Ooids are much more abundant than bryozoan and crinoid bioclasts and medium sand size quartz grains. Ooids are also medium sand size (mode 0.20 to 0.30 mm). Detrital quartz ranges from 27% to 46% of sample volumes. Quartz grains are dominantly clear, single crystals, with sharp extinctions. Ooids typically have detrital quartz cores; however, bioclast cores are also common (Figure 5b). It is the abundance of detrital quartz and the quartz cores of the ooids that distinguish Big Clifty ooids from those in the underlying Monteagle. Interestingly, this lithology has an open, uncompacted fabric (Figure 5b).

Sandstones with common tangential, planar, and herringbone cross-beds, clay drapes and laminations, and burrows form 17% of the measured stratigraphic sequence. Grain composition averages $Q_{92}F_0R_8$ with 9% clay matrix, i.e. a sublithic arenite. Some samples are quartz arenites, others are sublithic wackes (Figure 6). Calcite is abundant as spar cement, micrite matrix (average 9%), and fossil and peloid grains (insoluble residues, 0-34%). Feldspar and mica are rare. Rock fragments are dominantly pelitic. The few chert grains were included in the quartz percentages. Grains are dominated by single crystals of clear quartz but polycrystalline and strained quartz grains are common (1-12%). Grain size is very fine to fine sand with the coarse extreme in the 0.10 and 0.20 mm range. The grains are subangular and poorly to moderately well sorted. The dominant cement is spar calcite. The lesser quartz cement is more comon in the high clay matrix samples. Comparison of the interstitial clays with SEM photos in Wilson and Pittman (1977) indicates that the clays are

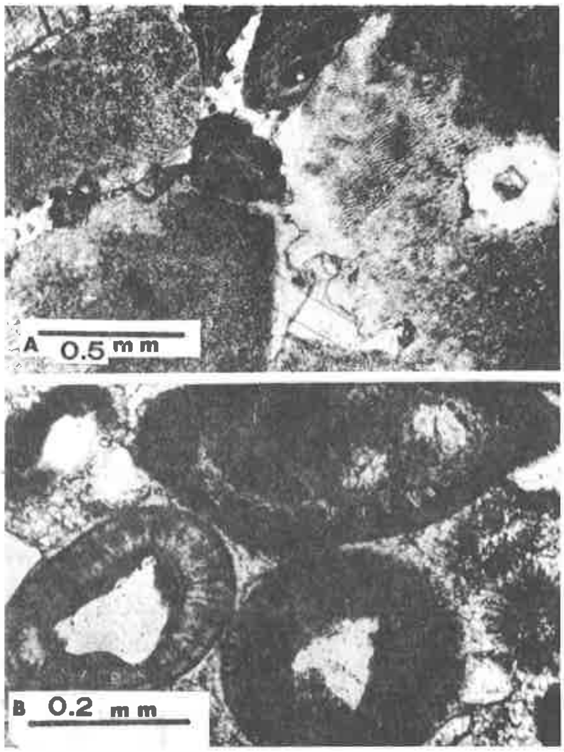


*Figure 5. Photomicrographs. A - illustrates the over-compacted appearance of the skeletal grainstones (and skeletal packstones) of the Big Clifty (Hartselle).
Note that the stylolite line separates spar cement areas as well as grains implying post cementation compaction. Uncrossed nicols, unstained, 40X sample ME - 6a:
B - illustrates the open uncompacted appearance of the oolitic grainstones as well as the clear simple quartz cores and thick cortex commonly observed. Uncrossed nicols, unstained, 10x, sample NM-2a.*

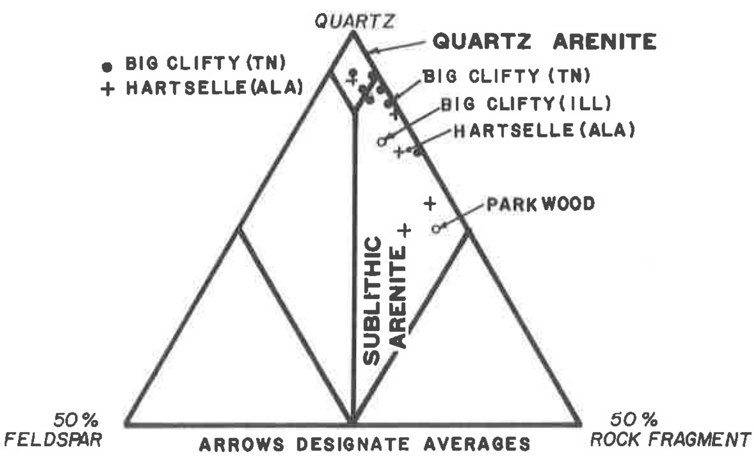


*Figure 6. Upper portion of the arenite composition triangle. Arrows point to average composition values. Parkwood and Illinois Basin data from Mack et al., (1980).
The Hartselle (Alabama) samples are wacke or near wacke in composition.*

dominantly detrital in origin; though some appear to be authigenic (recrystallized?). Sparse fossils are dominated by crinoids with abundant (20-30%) large bryozoan fronds near the base of the Monteagle East (ME) section.

Shale is the most abundant lithology, 19% overall, and is most common to the west (Figure 3, 4). Colors include red, green, maroon, olive, greenish-gray, and dark gray. Shale beds range from 2 in (5 cm) to 10 ft (2.5 m) thick and commonly pinch out in a few meters along outcrop.

Lithofacies

The pattern of geographic change in the Big Clifty is illustrated in the lithofacies map of Figure 3 and the cross section of Figure 4. To the southwest the proportion of carbonates increases. Indeed it would be impossible to recognize the Big Clifty as a mappable unit were it not for the presence of erratically distributed lenses of sandstone and shales. To the east (location W) shales are more abundant than packstones.

DISCUSSION

Petrology

Sandstones: Mack *et al.*, (1980) petrographically compared the Hartselle and associated Parkwood sandstones of northern Alabama with the Hardinsburg and Big Clifty Sandstones in the Eastern Interior Basin. The Parkwood is a lithic sandstone ($Q_{75}F_3R_{22}$) with 20% schiostose rock fragments and 15% unstable quartz. In contrast the Hardinburg and Big Clifty samples averaged $Q_{86}F_4R_{10}$ in which the rock fragments are dominantly sedimentary rock grains. The Alabama Hartselle has a close affinity to the Parkwood with more quartz and pelitic rock fragments.

We compared the Big Clifty of our study area to the Hartselle of northern Alabama. Big Clifty samples average $Q_{92}F_0R_8$ with 9% clay matrix and a major carbonate fraction (up to 34%). The quartz is 9:1 simple versus polycrystalline and strained types. The Alabama Hartselle averages $Q_{85}F_2R_{13}$ with 16% clay matrix, virtually no carbonates, and a simple to polycrystalline quartz ratio of 6:1. This difference between the petrologic character of the Alabama Hartselle and Tennessee Big Clifty is obvious by cursory comparison of fabrics (Figure 7). Although it is not pronouced, it seems improbable that such a contrast is attributable to increasing maturity going from south to north from a southerly source. Rather the similarity of the Big Clifty in the study area with published data on the Illinois Basin area suggests a northerly source for the Big Clifty.

Peterson (1956) inferred two sandstones, one, originating from the north, overlying a second originating from the south. He also shows dominantly carbonate rocks in the interval between his northerly (Big Clifty) and southerly (Hartselle) source areas. This is in agreement with our findings.

Carbonates: The petrology of the Big Clifty carbonates shows strong similarities to both the underlying Monteagle and overlying Bangor. The single characteristic that makes the Big Clifty lithologies readily distinguishable from its neighbors is the abundance of clastics. The insoluble residue content of the Big Clifty is quite high averaging 28% in dolomicrites, 16% in lime micrites, 16% in wackestones, 10% in packstones, and 5% in fossiliferous grainstones. Clearly there is a consistent relationship in which higher energy lithologies have less clastic residue. The oolitic grainstones also have a high clastic residue (22% to 48%) but the resiude is fine sand sized quartz whereas in the other carbonates it is dominantly clay mud.

Lithofacies Interpretation

Swann (1964) suggested that many of the smaller Chesterian sands of the Illinois Basin, such as the Big Clifty, might be offshore bar deposits. This suggestion is consistent with our observations at the northern (NM) study location (very fine sand grain size, ripple marks, cross beds, mineralogical maturity, carbonate cement). Nearby, to the south (EH), large plant fragments were found in an intraclast

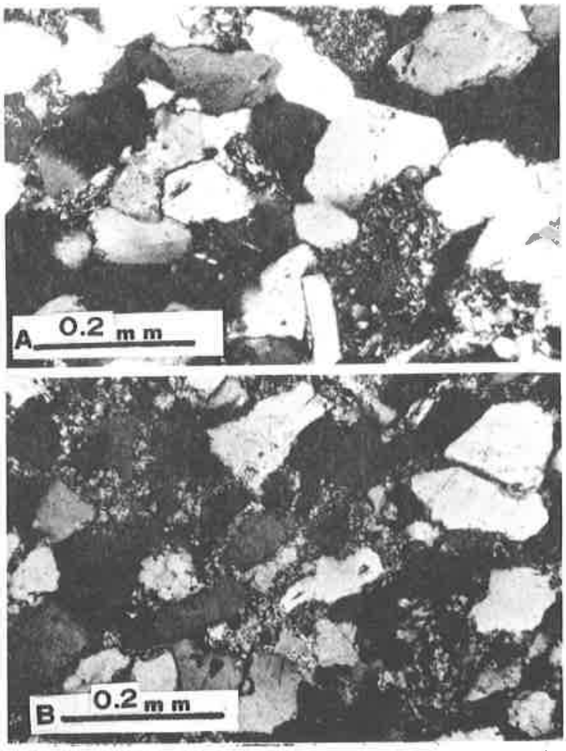


Figure 7. Photomicrographs selected to illustrate the compositional character and contrast of samples from the Alabama Hartselle and samples from nearby locations of equivalent age clastics in Tennessee, the Big Clifty. A 0.2 mm bar and cross nicols are common to both photos. A - Hartselle (Ala). An argillaceous sublithic arenite with abundant chert, polycrystalline quartz, rock fragments (14%), and clay matrix (14%). Very porous and quartz cemented. B - A sample of the Big Clifty (Hartselle) of Tennessee. A quartz arenite with finely crystalline calcite spar cement. Relatively few strained/ploycrystalline quartz grains and rock fragments are present. Sample NM 1a.

wackestone filled channel, further suggesting nearshore deposition. Although the evidence is certainly not conclusive the erratic arenite pods may represent isolated sandbar sets that have a northern source area. The arenaceous oolite at NM probably formed under similar physical conditions to that of the associated sandstones, perhaps with a sharper topographic relief or perhaps it was more distant from the shoreline.

The shale intervals represent varied subenvironments that were closely associated with shallow marine water. This is indicated by their proximity to normal marine limestones and the common occurrence of whole fossils similar to those found in the adjacent limestones. The shales may simply represent more sheltered subenvironments or periods of time when clastic input was either all fine grained or more distant from the depositional area.

Peterson (1956) illustrates the lithofacies distribution of the Big Clifty and, although different outcrops were studied, his general pattern agrees well with that in Figure 3.

The only channel fill deposit noted was to the north (Elk Head). Here channels were cut into shale and later filled with dolomicrite, clay seams, and plant debris. The source of the dolomicrite was probably the supratidal flats represented by the dolomicrite overlying the shale (Figure 4). The influence of tides is indicated by clay drapes in dolomicrites and sandstones, and herringbone cross-bedding in the arenites.

The petrology and sedimentary structures observed in both carbonate and clastic lithologies are common to shallow water, nearshore, wave and tide dominated environments. Probably wave action was the dominant sedimentary process operating during the deposition of both carbonates and clastics. Once waves created shoals and offshore bars and kept fine grained material suspended, the wave generated currents could carry the fines into low energy areas between the bars and shoals. Tides played a subdued role in deposition of the Big Clifty. This is indicated by a notable lack of cut and fill structures and other characteristics of channels (except at Elk Head).

A Depositional Model

An interpretation of lateral changes is here developed from vertical changes in the measured section. This was done by means of an evaluation using transition count arrays (Krumbein, 1967). The number of counts is not sufficient for Markov Chain analysis but the results can be summarized in schematic cross sections (Figure 8). Three situations occur depending on clastic input. If no sand is available skeletal grainstone dominates the high energy deposit. Input of a small amount of sand size clastics results in arenite oolitic grainstones; an abundance of sand results in quartz or sublithic arenites. The pattern of lithofacies distribution is a function of spasmodic clastic input rather than any change in the overall water depths.

Source Area

The western shelf of the Appalachian Basin was the site of carbonate deposition during Mississippian time except for the Big Clifty and its equivalents. There are several possible source areas for these clastics. To get to the study area, clastics may have (1) come across the Appalachian Basin; (2) crossed the Black Warrior basin; or (3) crossed the area of the Cincinnati Arch. The lack of equivalent clastics to the west, across the Nashville Dome, precludes a western source area.

1. The Appalachian Source Area. To the northeast th Big Clifty grades into a shale facies finally grading into the Newman-Greenbrier Limestone of the Appalachian Geosyncline (DeWitt and McGrew, 1979; Milici *et al.*, 1979). To the southeast the Big Clifty-Hartselle interval cannot be distinguished within the Monteagle-Bangor sequence (Thomas and Cramer, 1979). To the east the interval grades into Greenbrier Limestone (DeWitt and McGrew, 1979). Thus, an easterly source for sand size clastics seems unlikely. The eastermost outcrop studied here (location W) is 55% shale (45%

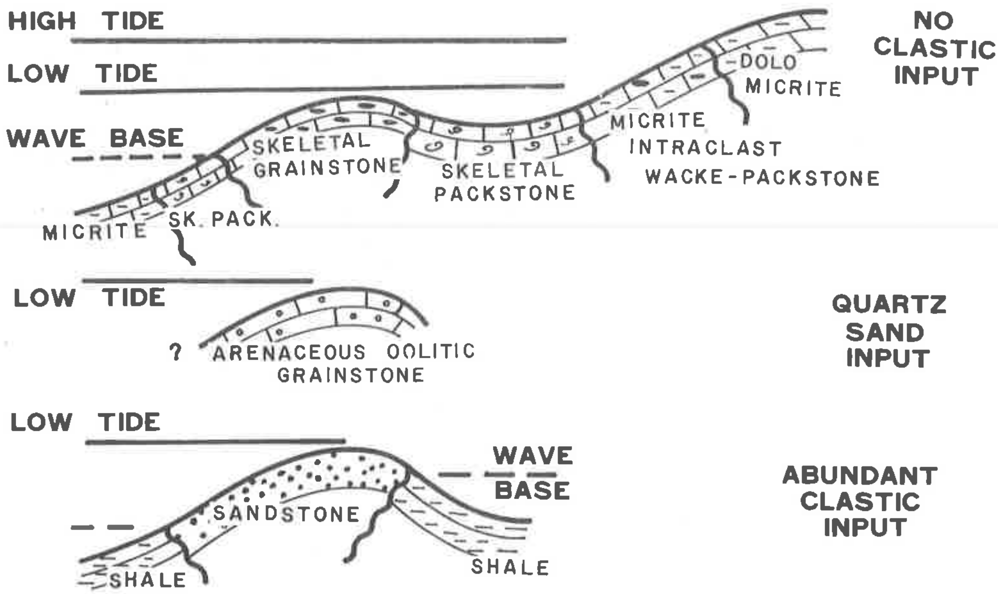


Figure 8. Idealized lateral variation in lithofacies. Based on counts of the vertical transitions observed in a composite columnar section. Relative high tide, low tide, and wave base depths are based on interpretation of sedimentary structures and presence of the fine detritus.

limestone). The source of these fine clastics is probably the Big Clifty or Hartselle depocenter.

2. The Appalachian-Ouachita orogenic belt (southern) source. To the south lies the type Hartselle Sandstone. Thomas (1972, 1979) and Thomas and Mack (1980) concluded that the thick sandstone trend of the Hartselle in Alabama is an association of barrier sand deposits with a source area to the southwest in the Appalachian-Ouachita orogenic belt. However, they also noted that the thick, continuous sandstones cannot be traced to the north or northeast into Tennessee; rather the sands become discontinuous and the Hartselle interval is marked by sandy and argillaceous carbonates (Thomas, 1979, p. 11). This thinning of coarse clastics and a change to a carbonate dominated facies was also noted by Peterson (1956).

3. We are left with the only likely prospect, a northern or northwestern source spilling across the Cincinnati Arch. To the northwest the Chester series of the Illinois Basin was deposited as a dominantly clastic sequence under the influence of cyclic variations in the Michigan River Delta (Potter *et al.* 1958; Swann, 1964). some scenarios envision extension of this delta complex far to the south and it is quite possible that clastics may have spilled over the Cincinnatic Arch into Kentucky and Tennessee (Sable, 1979, p. 93). Abandonment and subsequent reworking of this delta extension by nearshore marine processes may well have produced the Big Clifty lithofacies complex of the study area.

Potter *et al.* (1958, p. 1028) noted that the Big Clifty is the only sandstone in the Eastern Interior Basin with a westerly transport direction, indicating a source area to the east of the Big Clifty outcrops in Kentucky. Sable (1979) illustrates longshore currents that could have transported sands from the abandoned deltiac lobe back to the west thus supplying clastics to form the submarine bars and associated environments now called the Big Clifty. The Big Clifty thickens and becomes more sandy north toward the Kentucky-Tennessee line (Milici *et al.*, 1979, p. 17) and is lithologically similar to equivalent sands in western Kentucky (Sable, 1979, p. 89).

Figure 9 summarizes our interpretation of the regional relationships between the Big Clifty and Hartselle Formations. It is modified from Sable (1979) principally by extending the Michigan River delta complex further to the southeast. Area 1 is the main Big Clifty sand which we suspect was once continuous across the Cincinnati Arch into the type area of western Kentucky. Area 3 contains the type Hartselle and an

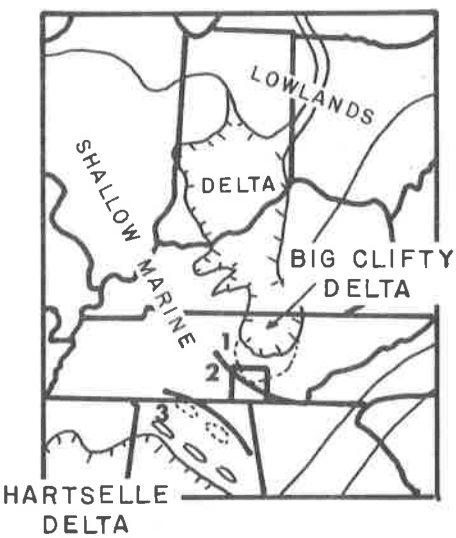


Figure 9. A diagram of the possible regional relationships between Big Clifty and Hartselle depocenters. Modified from Sable (1979) and Thomas and Mack (1980). Area 1 is the area of the Big Clifty deposition. Area 2 is the area of Monteagle-Bangor undifferentiated where clastics are only sporadically developed. Area 3 is the Hartselle depositional area where dashed ellipses are submarine bars and solid ellipses are barrier islands.

area of adjacent sand lenses. These sands thin and become discontinuous northeast into Tennessee giving way to Area 2, an area dominated by shallow marine limestones and dolostones with arenaceous limestone, sandstone, and shale lenses. Area 2 is interpreted as a carbonate bank that separates the Hartselle and Big Clifty delta fringes.

SUMMARY AND CONCLUSIONS

The reintroduction of the term Big Clifty Formation (Peterson, 1956; Vail, 1959) for what is currently called the Hartselle Formation (Stearns, 1963) in Tennessee literature will help clarify the stratigraphic relationships involved. Specifically, the type Hartselle in Alabama appears to be separated from the bulk of the Tennessee Hartselle by an area of dominantly carbonate deposition. In contrast the Tennessee Hartselle was apparently once continuous with the Big Clifty of the Kentucky and Illinois Basin area. Furthermore the petrology of the sandstones involved suggests that the Tennessee Hartselle is more similar to the Big Clifty than to the Alabama Hartselle. Thus it appears that two clastic depocenters were active; the one to the North (Big Clifty) extended down into central and southern Tennessee, the one to the South (Hartselle) brought clastics only to the southern border of Tennessee.

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THE HURRICANE RIDGE, GLEN LYN, AND CALDWELL SYNCLINES OF SOUTHEASTERN WEST VIRGINIA AND SOUTHWESTERN VIRGINIA: A REINTERPRETATION

By

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ABSTRACT

The Hurricane Ridge syncline of Tazewell and Giles Counties, Va., and Mercer and Monroe Counties, W. Va., lies along the boundary between the Appalachian Plateau and Valley and Ridge provinces, which is here marked by the St. Clair thrust fault. Recent reconnaissance mapping, together with a reexamination of earlier work, show that the current perception of this syncline is incorrect and actually includes three separate and distinct synclines: an overturned syncline that parallels the St. Clair thrust throughout its length; a broad, open syncline at the "type area" in Mercer County; and a broad, gentle, northerly trending syncline in Monroe County. The first of these is herein named the Glen Lyn syncline, the second retains the name of Hurricane Ridge, and the third is correlated with the Caldwell syncline of Greenbrier County, W. Va.

INTRODUCTION

The Hurricane Ridge syncline of southwestern Virginia and southeastern West Virginia, first noted by Campbell (1896), was named and described by Reger (1926, p. 146-149). The axial trace shown by Reger (1926, Maps II and IV) in Mercer and Monroe Counties, W. Va., and Giles County, Va., has since been used widely; it appears on the Geologic Map of West Virginia, (Cardwell and others, 1968) and was shown with only minor modifications by workers in the "type area" (Cooper, 1961, p. 48; Cooper, 1971; Thomas, 1966, Figure 2). Recent field reconnaissance by the present author, together with analysis of published and unpublished maps and cross sections, show clearly that the "Hurricane Ridge syncline" as used by Reger (1926) is in fact three separate synclines that differ in form and trend. The distinction between these three folds is useful in assessing the problematic relationship between sedimentation and structure in the area, and in analyzing the relationship between southern and central Appalachian structures across the nearby juncture of these two regions; these are subjects for separate paper. Accordingly, the Hurricane Ridge syncline is redefined herein, whereas the name Glen Lyn syncline is introduced, and the Caldwell syncline of Price and Heck (1939) is extended, for parts of the original structure as mapped by Reger.

The structural setting of the Hurricane Ridge syncline of Reger (1926) is shown in Figure 1. The principal structure in this area is the St. Clair thrust fault, which extends from the core of an anticline in southwestern Alleghany County, Va., southwestward for more than 150 km to Richlands, Va., and, according to Harris (1965), nearly 450 km farther into northwestern Georgia as the Clinchport and related thrust faults.

The St. Clair thrust fault, which generally separates Ordovician rocks from underlying Devonian and younger beds, has long been considered part of a major structural boundary, called the "Appalachian structural front" by Price (1931) and the "Allegheny front" by Rodgers (1964), between folded and faulted rocks of the Valley and Ridge province to the southeast and relatively undeformed rocks of the Appalachian plateau to the northwest. Northwest of the St. Clair thrust, and roughly parallell to it, the Boissevain and Richlands reverse faults extend from the Richlands area northeastward across Tazewell County (Figures 1, 2); this structural element is extended farther northeastward in the form of the associated Abbs Valley anticline to

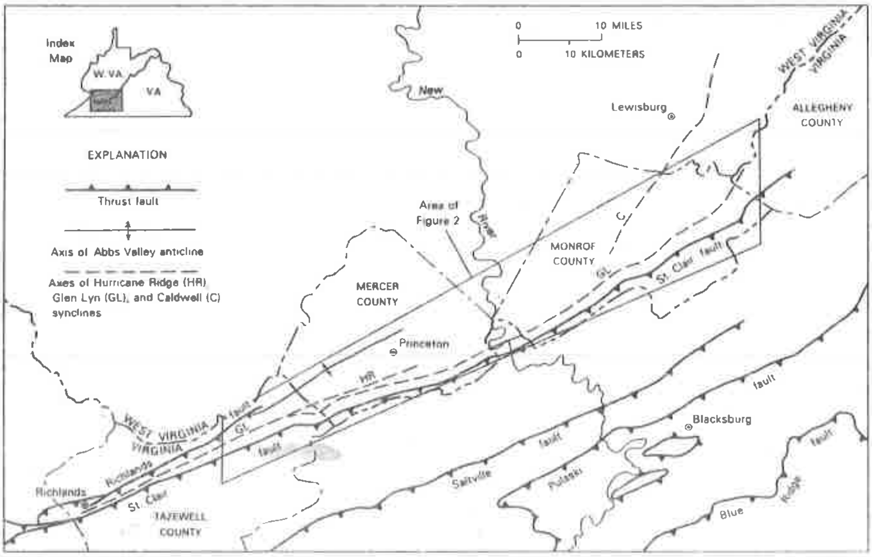

Figure 1. Structural setting.

Princeton, W. Va., where the anticline loses relief and disappears. The interval between these structures and the St. Clair fault, a distance of about 1 km near Richlands to about 12 km at Princeton, is occupied by the Hurricane Ridge syncline as originally mapped by Campbell (1896) and later extended by Reger (1926).

Details of the structure along the northeastern two-thirds of the St. Clair thrust fault are shown in Figure 2. From the Richlands area (Figure 1) to near Bluefield, Va., there is a single syncline, overturned to the northwest, with both limbs dipping southeastward, as shown by Cooper (1944) (cross section A-A' on Figure 2). At Bluefield, detailed mapping by Englund (1968) clearly shows a structure that can be resolved into two separate synclines: the overturned syncline on the southeast, near the St. Clair thrust fault, and a broad, open syncline on the northwest (Figure 2, cross

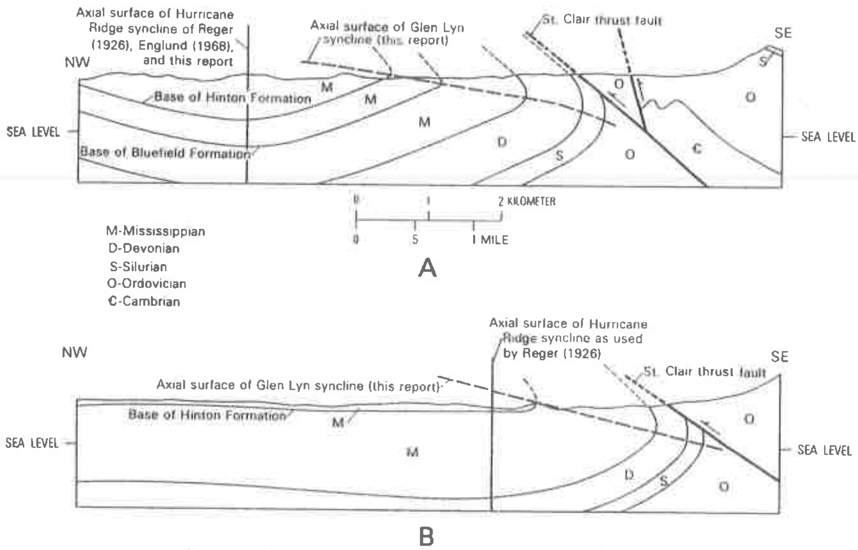

Figure 3. Cross sections showing location of axial surfaces of synclines. A, near Bluefield, W. Va.; from Englund (1968) and Cooper (1944): B-B' and C-C' (in part) of Figure 2, this report. B, near Glen Lyn, Va.; from Reger (1926); E-E' (in part) of Figure 2, this report.

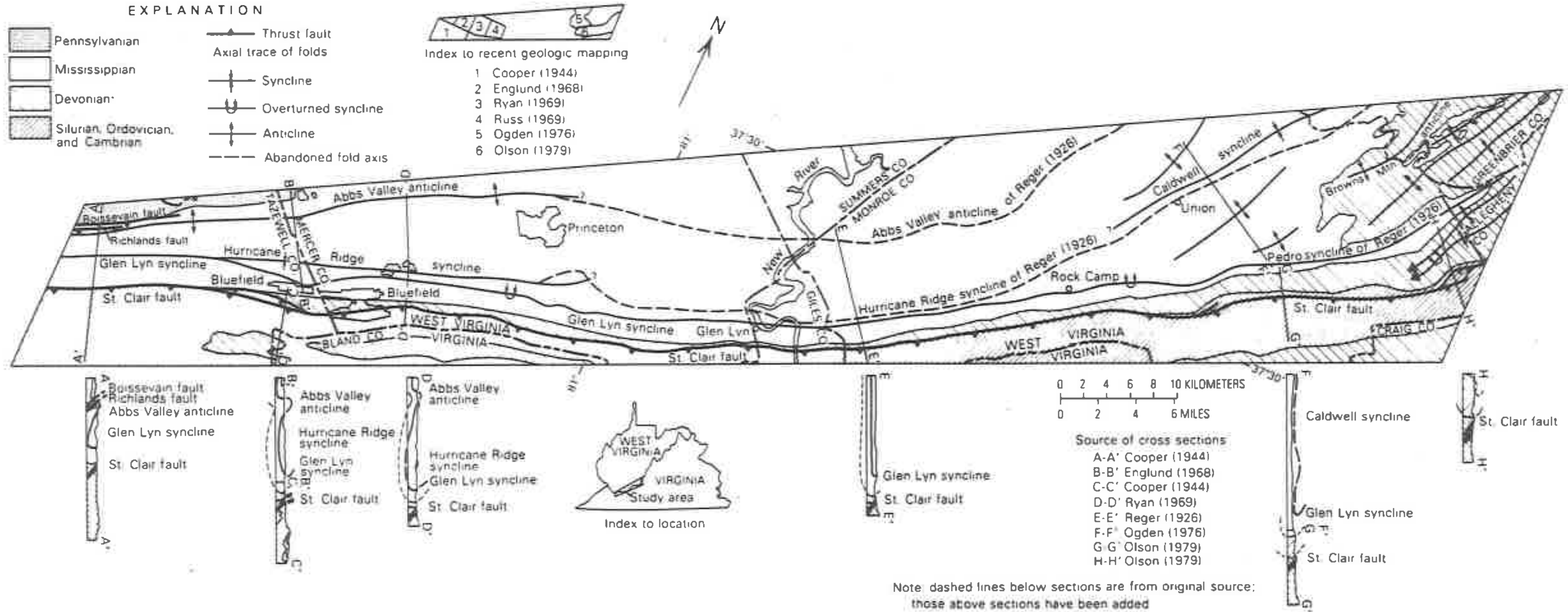

Figure 2. Generalized geologic map and cross sections of the Hurricane Ridge, Glen Lyn, and Caldwell synclines.

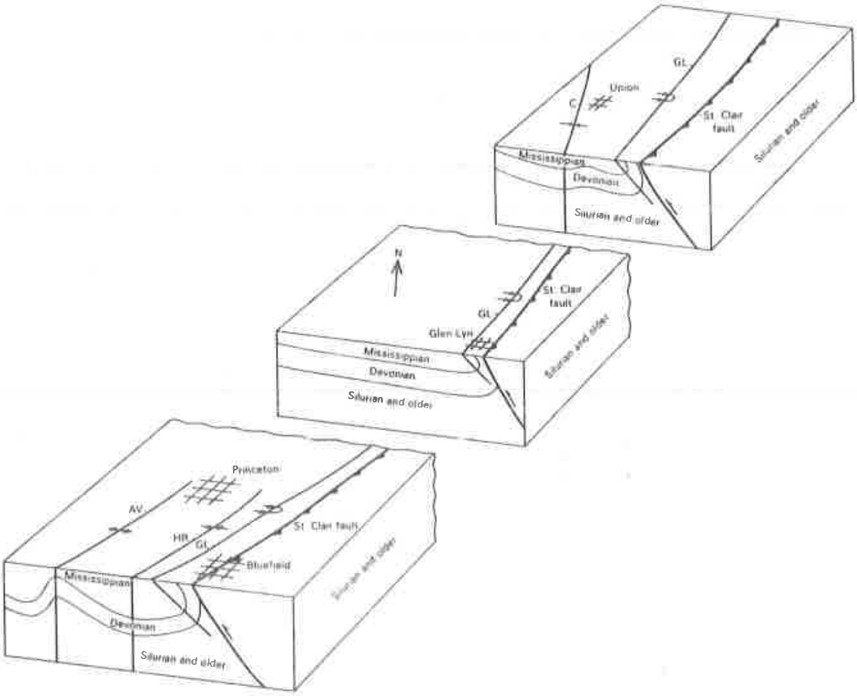

Figure 4. Simplified block diagram showing relationship of St. Clair Fault, Abbs Valley anticline (AV), Hurricane Ridge syncline (HR), Glen Lyn syncline (GL), and Caldwell Syncline (C). Not to scale.

section B-B'). These two synclines merge laterally into each other without an intervening anticline, and with a common limb, as shown in Figure 3A, and have consequently been considered by previous workers, including Reger (1926), to be a single syncline. As such, Reger's axis northeast of Bluefied could be considered a trough line. It is useful, however, to consider the two folds separately, defined by two separate and distinct axial surfaces, each of which has been considered by different workers at different times to be the axis of the "single" syncline. Tracing of these axes toward the northeast through Mercer County, W. Va. (including the "type area," Hurricane Ridge, between Bluefield and Princeton) shows that they remain separate and roughly parallel to a point southeast of Princeton where the northwestern, open fold loses structural relief and disappears in a manner similar to the Abbs Valley anticline northwest of Princeton. There Reger (1927, Map II) incorrectly showed the axial trace of the open fold curving back to the southeast near the position of the overturned syncline (see Figure 2).

In Giles County, Va., and Monroe County, W. Va., the axial trace shown by Reger (1926, Map IV) roughly coincides with that of the overturned syncline (Figure 2, cross section E-E'; Figure 3B) northeastward to Rock Camp, where the trace mapped by Reger swings northward toward Union and away from the St. Clair fault. However, Reger's map shows, and field reconnaissance bears out, that the overturned beds of the southeast limb of the syncline continue northeastward parallel to the fault for some 35 km, to the northern boundary of Monroe County. Recent mapping by Ogden (1976) and Olson (1979) has shown the nature of the overturned syncline in northeastern Monroe County (Figure 2, cross sections F'-F', G-G'; Ogden called it the "Canterbury Cave Syncline"). The northeastern end of the overturned syncline is represented by Reger's (1926, Map IV) overturned Pedro syncline, which apparently disappears to the northeast in Allegheny County, Va. The axial trace drawn by Reger through Union (Figure 2), which represents a broad, gentle syncline, has been replotted by Ogden (1976) somewat westward of Reger's position, so that it connects with the trace of the Caldwell syncline in Greenbrier County, W. Va. (Price and Heck, 1939).

Figure 5. Axis of the Glen Lyn syncline at Glen Lyn, Virginia, along County Road 648 about 200 m northwest of U.S. Route 460. Looking southwest; hammer (near base of outcrop) marks approximate position and dip of axial surface. Beds are sandstones and shales of the lower part of the Hinton Formation (Upper Mississippian).

Thus, three separate synclines have been confused under the name "Hurricane Ridge syncline": an overturned syncline, parallel to the St. Clair thrust fault for virtually its entire length; a broad, open syncline in Mercer and northeastern Tazewell Counties, which appears to be related to the Abbs Valley anticline; and a gentle more northerly trending syncline in Monroe County (Figure 4). The name "Hurricane Ridge syncline" should be restricted to the open Mercer County fold, the axial trace of which passes along Hurricane Ridge. The name "Caldwell syncline" is extended from Greenbrier County into Monroe County for the open syncline near Union. For the overturned syncline, the name "Glen Lyn" is proposed for excellent exposures of the axial portion of this fold at and near Glen Lyn, Giles County, Va. (Figure 5; see also Cooper, 1961, p. 95, 160). These revisions in the fold nomenclature are intended to draw attention to the erroneous perception of the fold trends in this area.

The Glen Lyn syncline is probably related to movement along the St. Clair thrust fault, and may be the result of ramping of a deep decollement combined with drag effects on the footwall. The Hurricane Ridge syncline is genetically separate and may be related to the formation of the Abbs Valley anticline, with which it is roughly parallel and coextensive, except where overridden to the southwest by the St. Clair thrust and Glen Lyn syncline. The Caldwell syncline is one of several of more northerly-trending plateau folds characteristic of the central Appalachians in this area, such as the Browns Mountain anticline (Figure 2).

Cooper (1961, p. 95-96; 1964, p. 101-103) and Thomas (1966) have argued that the Hurricane Ridge Syncline is a "depositional syncline" which was actively subsiding during deposition of post-Maccrady Mississippian sediments. Their data are derived mostly from the Hurricane Ridge syncline as redefined herein. The present distinction between the Hurricane Ridge and Glen Lyn synclines complicates the anlayses of these authors, who made no such distinction, and suggests that a reexamination of their conclusions is necessary.

The axial traces of gentle plateau folds are difficult to determine without detailed geologic mapping. Thus, the exact extent of the Abbs Valley anticline and Hurricane Ridge syncline northeast of Princeton is not known, nor is the number, extent, and trend of comparable folds northwest of the Glen Lyn syncline between Princeton and northern Monroe County, where folding becomes more pronounced. The axial traces shown by Reger (1926, Maps II and IV) are open to question because of the lack of detailed mapping and stratigraphic analysis; for example, structure contours (on which axial traces were apparently based) were drawn on the base of Reger's (1926) Avis Limestone Member of the Hinton Formation, which has been totally removed by erosion throughout Monroe County; moreover, the contours do not correspond closely with mapped contacts. Distinction between the Glen Lyn syncline and the Caldwell syncline may be useful in analyzing structural contrasts across the juncture between the southern and central Appalachians; the Glen Lyn syncline is aligned with the trend of the southern Appalachians, and the Caldwell with the trend of the central. Detailed information on gentle, minor folding in this area may also prove to be of value.

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RADIOMETRIC AND MAGNETIC SURVEYS OF THE

DARK RIDGE DUNITE, JACKSON COUNTY, NORTH CAROLINA[1]

By

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ABSTRACT

Both radiometric and ground magnetic surveys were employed in an attempt to outline the partially exposed Dark Ridge dunite body which occurs in a terrain of mica-bearing, quartzo-feldspathic gneiss and schist. The gamma radiation data define a series of closed contours concentrated over the dunite consistent with the mapped distribution of all outcrops. Values for the dunite fall generally below 85 counts per second; those for the country rock are typically within the 100-150 counts per second range.

A total field ground magnetic survey fails to reflect the presence of the dunite as outlined by field mapping combined with the radiometric survey. The distinctive, elongate, closed magnetic contours, which pass through the area in which the dunite is outlined, are interpreted as reflecting lithologic variations within the country rock below the dunite mass. Depth estimates, based on profiles across three magnetic highs distributed along the major linear trend in the eastern half of the dunite area, suggest that the Dark Ridge body is on the order of 100 meters thick, comparable to the thickness of the Balsam gap dunite to the north.

The absence of a magnetic anomaly associated with the Dark Ridge body is directly related to the lack of a significant magnetic susceptibility difference between the dunite and the gneiss country rock. Mean susceptibilities are 3.14×10^{-5} (dunite) and 2.44×10^{-5} (gneiss). The relatively small susceptibility values for the dunite are the result of its low degree of serpentinization (and thus weak development of secondary magnetite).

INTRODUCTION

Earlier reconnaissance mapping of the ubiquitous ultramafic rocks in the southern Appalachian Mountains focused on their value as a source of olivine (Hunter, 1941) and their potenial for mineable chromite (Hunter *et al.*, 1942). Subsequent studies emphasized their asbestos content (Conrad *et al.*, 1963). Because of increased demand for refractory olivine, there is currently considerable renewed interest in these rocks.

Since this is a region of intense chemical weathering, thick soil development, and extensive vegetative cover, outcrops are sparse and most of the ultramafic masses are known at best only approximately on the basis of conventional geologic mapping. Aside from a lack of precise information regarding the areal extent of individual bodies, there is little known about their subsurface shapes and thicknesses, with the exception of two studies (Greenberg, 1976; Honeycutt *et al.*, 1981).

Geochemical and geophysical exploration techniques are essential for defining the areal extent of these bodies and for obtaining information on their subsurface extent. With regard to the former approach, geochemical surveys for nickel in soils and stream sediments have been successful in outlining Southern Appalachian ultramafic bodies (Worthington, 1964; Carpenter and Hale, 1967; Callahan *et al.*, 1978).

The earliest geophysical studies of ultramafic rocks in this region were conducted by Hunter *et al.*, (1942) who found low magnetic values characteristic of several North

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Carolina dunites in contrast with higher values typical of the enclosing gneisses and schists. Similar results of a ground magnetic survey were obtained by Callahan *et al.*, (1978) who found lower magnetic values associated with two ultramafic masses outlined by means of a geochemical survey. Other than an abstract by Greenberg (1976) and a paper by Honeycutt *et al.*, (1981), no published studies have reported geophysical interpretations for any of the ultramafic rocks in the Southern Appalachian Mountains. Honeycutt *et al.*, (1981) showed the presence of a well-defined positive ground magnetic anomaly over the Balsam Gap dunite in contrast with lower magnetic values over the enclosing gneiss. Modeling of the data indicated that this body extends to a shallow depth (50-100 meters).

The only published radiometric survey of an ultramafic body in the Southern Appalachian area is that conducted by Callahan *et al.*, (1978). This study showed the presence of generally lower scintillometer values associated with the dunite, outlined by their geochemical survey, as compared with greater values over the enclosing amphibolite.

The purposes of the study reported here were to characterize the Dark Ridge ultramafic body mineralogically and to test the utility of ground magnetic and gamma ray surveys for outlining the body. It was hoped that, in addition to determining the areal extent of the mass, the magnetic data could be used to infer its subsurface shape and depth.

The study was supported by the Society of the Sigma Xi and the North Carolina Department of Natural Resources and Community Development which provided financial assistance. We are grateful to these organizations, to Emmanuel Perez (who assisted the senior author in the field), and to James Kirkendall (who gave permission to work on his property).

MINERALOGY

Located in northern Jackson County, North Carolina (Figs. 1 and 2), the Dark Ridge body consists of a slightly altered dunite core rimmed by a thin sheath (1.5 meters) of highly altered dunite. The mass occurs in a terrain of mica-bearing, quartzo-feldspathic gneiss and schist. Foliation in these rocks strikes generally northeast-southwest (Figure 3) and dips 22° to 90°. Dip direction is commonly southeast. Layering is lacking within the dunite which is cut by joints of varying attitude (Figure 3).

Samples of relatively unaltered dunite are dominated by olivine (Table 1, Figure 4) which occurs as unstrained polygonal grains, ranging from 0.2 to 4 mm in diameter, and possessing triple-joint junctions which meet at approximately 120°. Locally the olivine occurs as large irregular porphyroclasts approximately 3-7 mm in diameter, characterized by strain bands and undulatory extinction. Atomic absorption spectrophotometric analysis of olivine separates (and several whole-rock samples) indicates that Fe and Mg contents vary little (Table 2). Computed Fo values for the olivine fall within the narrow range of 85.1-90.9 (Table 1) and there is no systematic variation in

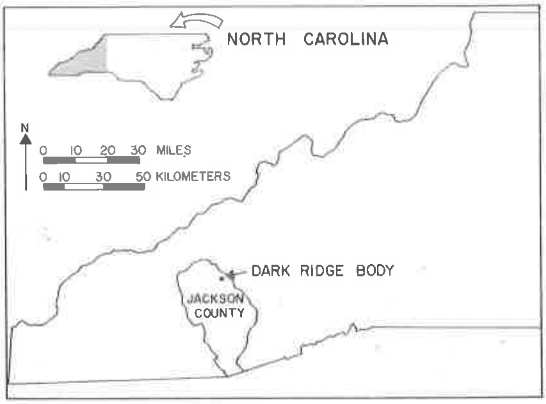

Figure 1. Location Map for the Dark Ridge dunite.

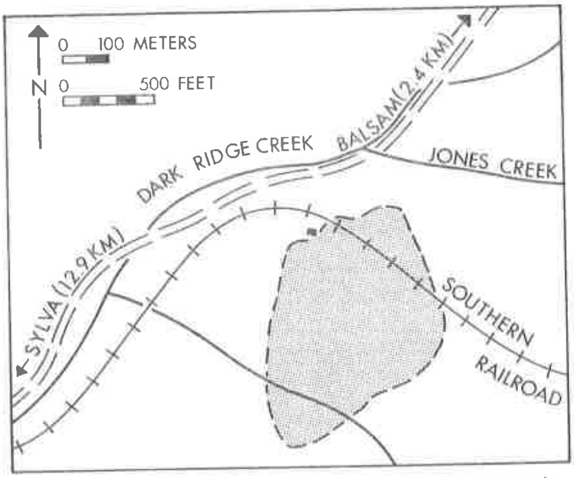

Figure 2. Cultural features in the Dark Ridge dunite (stippled) area.

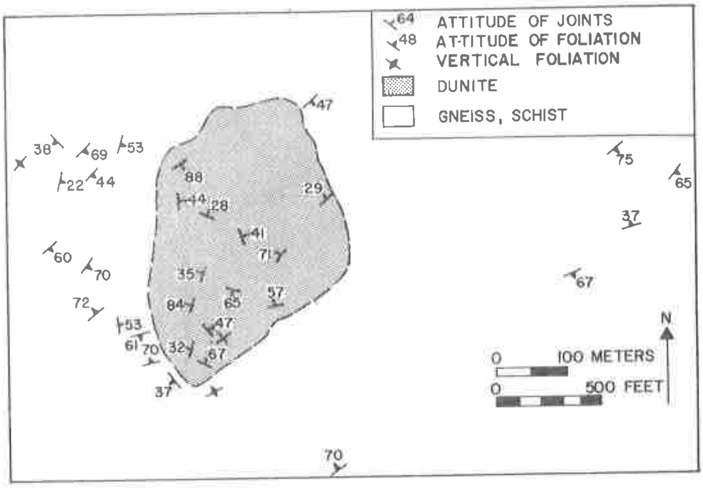

Figure 3. Geologic map of the Dark Ridge dunite area.

the Mg/Fe ratio across the dunite (Table 2, Figure 4). As the only other primary mineral, chromite is found typically as disseminated euhedral to subhedral grains which have a range in grain size from .05 mm to 2 mm (in diameter).

Secondary minerals consist of serpentine, talc, chlorite, tremolite, vermiculite, anthophyllite, and calcite, which make up from 1.6 to 41 percent of the rock by volume (Table 1). Talc, the most abundant alteration mineral, formed exclusively at the expense of olivine which it replaces entirely in some samples. Although some olivine has altered to chlorite, the majority of the chlorite surrounds and embays the chromite grains. As shown in Table 1, serpentine comprises typically less than one percent of the rock, in contrast with its common abundance in many Southern Appalachian ultramafc rocks (Misra and Keller, 1978). Petrographic examination indicates that the serpentine mineral is antigorite with which a very small amount of magnetite dust is associated.

Table 1. Modal analyses of Dark Ridge dunite and olivine Fo content.

Sample Number	3	4	5	11	13	25	26	27
Olivine	71.3	77.8	75.6	73.6	70.1	63.5	72.6	61.1
Chromite	0.8	1.5	1.1	0.9	1.4	1.5	1.1	0.3
Talc	17.7	11.4	14.8	17.9	19.2	6.2	16.5	3.6
Chlorite	3.6	3.8	3.9	2.9	3.1	22.5	5.4	30.6
Tremolite	5.5	5.3	4.2	4.3	5.4	4.7	3.9	1.7
Serpentine	0.8	0.2	0.2	0.4	0.8	0.9	0.4	1.4
Vermiculite		tr	0.2	tr		0.3		0.7
Anthophyllite	0.3						0.1	0.2
Calcite	tr					0.4	tr	0.4
Olivine Fo %	87.3	88.0	87.6	87.8	87.1	85.1	87.0	86.8

Sample Number	28	29	32	33	35	36	31
Olivine	58.3	72.6	97.7	91.6	31.7	84.3	95.9
Chromite	0.7	0.7	0.7	3.3	46.8	0.2	0.5
Talc	9.3	15.4	0.5	0.8	0.8	8.7	1.3
Chlorite	29.6	4.4	0.8	4.0	19.3	3.5	0.7
Tremolite	1.6	5.5				tr	
Serpentine	tr	1.4	0.2	tr	0.2	1.1	0.3
Vermiculite	0.2		0.1				
Anthopyllite	0.1						
Calcite	0.2	tr		0.3	1.2	2.2	1.3
Olivine Fo %	85.9	87.4	90.1	90.9			89.8

Table 2. Atomic absorption analyses of Dark Ridge olivine and dunite.

Sample Number	Fe_2O_3*	MgO	Mg/Fe ratio
25	9.00	37.00	5.7
26	8.57	41.37	6.7
27	8.75	41.33	7.1
28	8.65	38.00	6.1
28 whole rock	9.05	39.79	6.1
29	8.91	41.28	6.9
11	8.35	43.72	7.2
3	8.31	41.20	6.8
3 whole rock	8.51	43.11	7.0
4	7.15	38.05	7.3
5	8.57	43.94	7.1
33	6.89	48.00	10.0
33 whole rock	7.11	48.89	9.8
32	7.11	47.20	9.1
31	7.22	46.30	8.8
31 whole rock	7.39	47.92	9.0
13	8.25	40.40	6.8

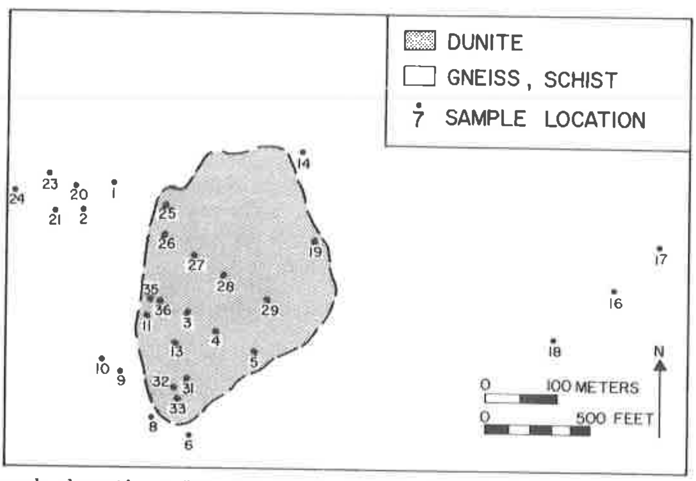

Figure 4. Sample locations for the Dark Ridge dunite area.

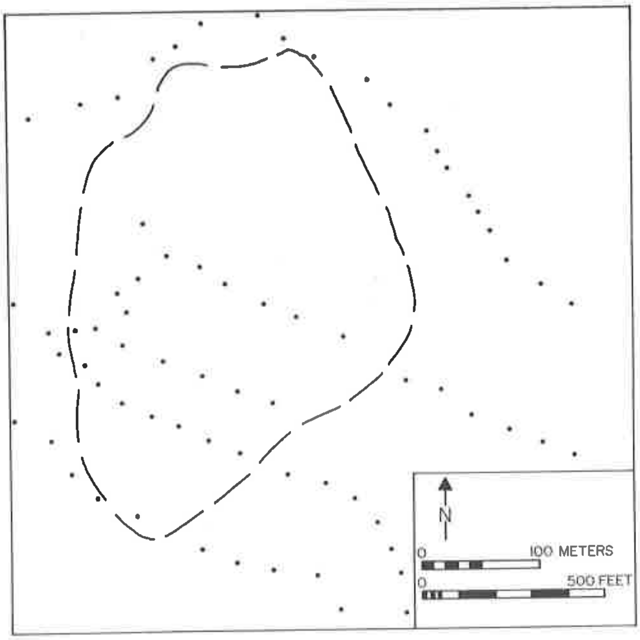

Figure 5. Station locations for radiometric (solid circles) and magnetic (open circles) surveys in the Dark Ridge dunite area.

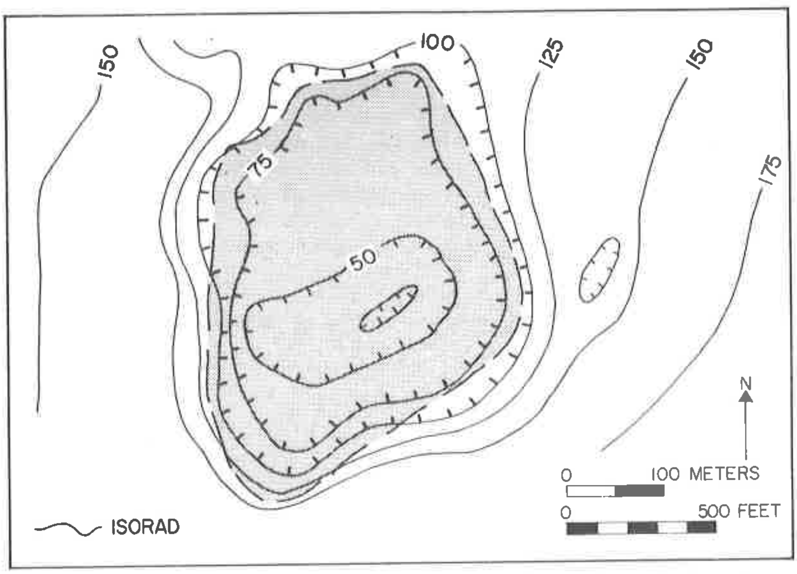

Figure 6. Radiometric (gamma ray) contour map of the Dark Ridge dunite (stippled) area. Contour interval 25 counts per second.

RADIOMETRIC SURVEY

Subsequent to geologic mapping and locating of the 15 or 20 dunite exposures on an enlargement of the U.S. Geological Survey Hazelwood Quadrangle topographic map, a Soiltest DISA-300 Portable Gamma Ray Spectrometer was employed to measure gamma radiation across the dunite and adjacent country rock. The survey was conducted along a series of approximately parallel, northeast-southwest traverses spaced roughly 50-100 meters apart (Figure 5). Instrument readings, using a six-second

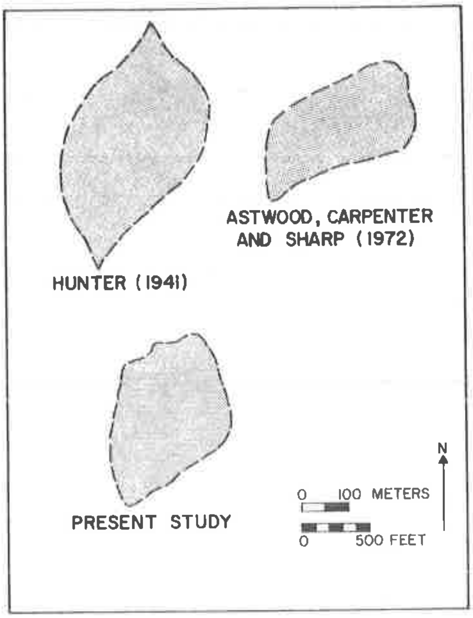

Figure 7. Earlier mapped outlines of the Dark Ridge body compared with that resulting from the present study.

time constant, were made at 30-meter intervals along the traverses. Gamma radiation values were then reduced to counts per second, plotted and contoured (Figure 6).

The data define a series of closed contours concentrated over the weakly radioactive dunite. The dunite-country rock contact, as drawn in Figure 6, is consistent with the configuration of the contours and the mapped distribution of all dunite outcrops. As indicated, gamma radiation values for the dunite fall generally below 85 counts per second (C.P.S.), whereas those for the country rock are typically within the 100-150 C.P.S. range. Combined with the results of surface mapping, the radiometric survey reveals a size and shape for the Dark Ridge body different from earlier indications based on the reconnaissance studies of Hunter (1941) and Astwood *et al.*, (1972), as shown in Figure 7.

Table 3. Magnetic susceptibility measurements on samples of the Dark Ridge dunite and country rock.

DUNITE

Sample Number	Degree of Serpentinization*		Magnetic Susceptibility**
3	0.8		3.09×10^{-5}
5	0.2		3.81×10^{-5}
26	0.4		3.54×10^{-5}
29	1.4		5.09×10^{-5}
31	0.3		1.46×10^{-5}
13	0.8		1.83×10^{-5}
		Mean	3.14×10^{-5}
		St. Dev.	1.34×10^{-5}

COUNTRY ROCK

Sample Number		Magnetic Susceptibility**
1		1.64×10^{-5}
8		1.84×10^{-5}
16		3.10×10^{-5}
10		3.16×10^{-5}
	Mean	2.44×10^{-5}
	St. Dev.	0.81×10^{-5}

*Modal percent serpentine
**emu/cm^3

MAGNETIC SURVEY

A total field ground magnetic survey of the Dark Ridge area was conducted with a Geometrics G-816 Portable Precession Magnetometer. As with the radiometric survey, measurements were made primarily along a series of roughly parallel, northwest-southeast traverses spaced 50-100 meters apart (Figure 5). Readings were taken along the survey lines at stations spaced every 15 meters, and drift was corrected by reading the instrument hourly at a base station. The data were then plotted on drift curves and a linear regression computation was used to standardize the values relative to a constant base station value. Following reduction of all values by 54,000 gammas, the data were plotted and contoured (Figure 8).

The ground magnetic map fails completely to reflect the presence of the dunite as located by field mapping and by means of the radiometric survey. The distinctive elongte closed contours are aligned with the strike of foliation in the metamorphic country rock and relate, undoubtedly, to variation in the magnetism of different lithologic layers within the gneiss.

To determine why a magnetic anomaly is not associated with the Dark Ridge dunite, a Soiltest Magnetic Susceptibility Bridge was used to measure susceptibility in samples of the dunite and the gneiss country rock (Table 3). The results show that there is no significant difference (0.70×10^{-5}) between the means of the magnetic susceptibiliy of the dunite and the country rock sampled.

DISCUSSION AND CONCLUSIONS

Of the two approaches used in this study, clearly the radiometric suvey is most effective in outlining the position of the dunite in relation to actual outcrops mapped (Figures 3, 4, 6). Thus, we conclude that in this region gamma ray spectrometer surveys may be particularly useful in outlining ultramafic bodies which are concealed by thin soil or forest litter. The pattern of isorads is much sharper with respect to defining the location of the Dark Ridge ultramafite as compared with that obtained in a comparable survey of the Rich Mountain body (Callahan *et al.*, 1978). Obviously,

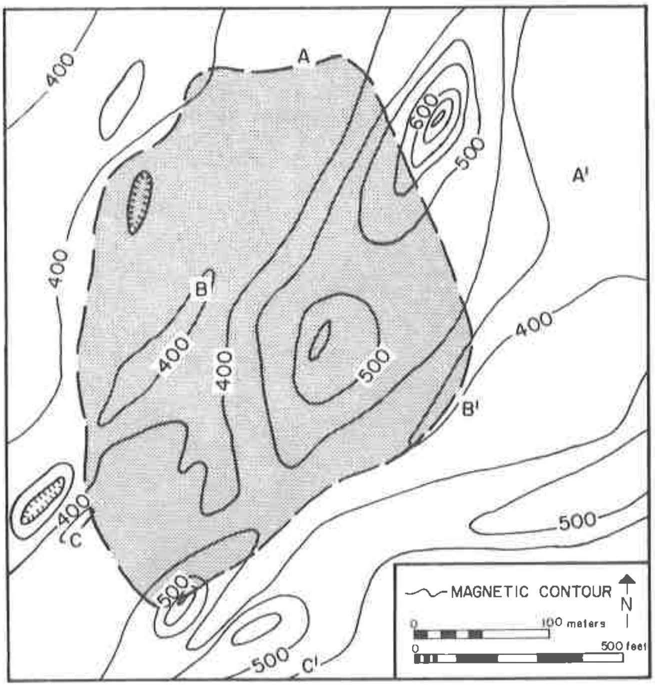

Figure 8. Ground magnetic map of the Dark Ridge dunite (stippled) area. Contour interval 50 gammas; base value 50,000 gammas.

isolated blocks of ultramafic rock, which may have migrated downslope from their source, will also show low gamma radiation. Thus, a radiometric survey may indicate greater aerial extent of ultramafic rock than is actually present. Therefore, radiometric surveys utilized in exploration for ultramafic deposits must be used with considerable caution.

Although a well-defined positive magnetic anomaly is associated with the Balsam Gap dunite (Honeycut *et al.*, 1981), located just 700 meters north of the Dark Ridge body, no such anomaly is associated with the latter. Of critical interest relative to interpretation of magnetic surveys over ultramafic bodies is comparison of the magnetic susceptibility of the ultramafic rock with that of the country rock. The mean susceptibility difference between the Dark Ridge dunite and its gneiss country rock is slight (0.70×10^{-5}, Table 3), thus this dunite does not show a distinctive magnetic anomaly. On the other hand, the magnetic susceptibility of the Balsam Gap dunite is roughly five times that of its country rock (Honeycutt *et al.*, 1981), and thus, it has a strong positive magnetic anomaly associated with it. Other ultramafic bodies have negative magnetic anomalies, presumably because the ultramafic rock in those cases has a magnetic susceptibility much lower than that of the enclosing country rock.

The difference in magnetic susceptibility within and among individual ultramafic bodies is clearly a function of the degree of alteration and, more specifically, the type of alteration. As pointed out by Saad (1969), the magnetic character of these rocks is largely the result of chemical remanent magnetization acquired during their alteration to serpentine. During the serpentinization of the olivine in these dunites, the relatively small amount of contained iron is released to form a fine secondary magnetite dust because the atomic structures of other typical alteration minerals such talc, chlorite, and tremolite, as well as the serpentine, admit little iron (Whittaker and Wicks, 1970; Moody, 1976, Wicks and Whittaker, 1977). A close correlation was found between magnetic susceptibility and degree of serpentinization for the Red Mountain ultramafite, California (Saad, 1979), the Twin Sisters dunite, Washington (Thompson and Robinson, 1975) and the Balsam Gap dunite, North Carolina (Honeycutt *et al.*, 1981). Samples of the Dark Ridge dunite (Table 1) from across the body show very minimal development of serpentine (maximum of 1.4 percent in two samples), thus little development of secondary magnetite, and there is no magnetic anomaly present. The main alteration minerals are talc, chlorite and tremolite, with trace amounts of vermiculite and anthophyllite. By contrast, the Balsam Gap dunite exhibits greater alteration to serpentine (greater than 2.0 percent in all but one sample, up to 39.6

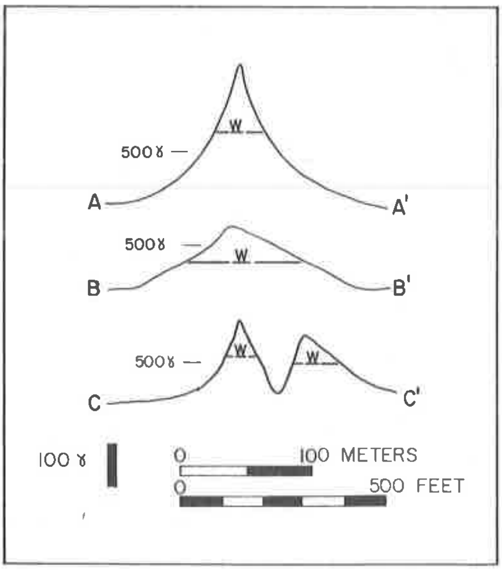

Figure 9. Magnetic profiles across the Dark Ridge dunite body. See Figure 8 for locations.

percent), regular development of secondary magnetite, and a positive magnetic anomaly coincident with the body (Honeycutt *et al.*, 1981; Honeycutt and Heimlich, 1980). The higher magnetic values over the dunite agree well with portions of it that are more heavily serpentinized.

Although magnetic data for the Dark Ridge body could not be modeled to reveal information on its subsurface shape and thickness, close examination of the ground magnetic map (Figure 8) provides a possible clue in this regard. We refer to the generally aligned contours which are parallel to the regional strike of the country rock foliation in the region. These pass through the area in which the dunite is outlined. Particularly notable is the alignment of three magnetic highs through the eastern half of the dunite area. This linear trend is parallel to the structural grain in the gneiss, passes through the dunite body, and may reflect the presence of an amphibolite, or other mafic rock, layer within the gneiss below the dunite. The dunite, then, acts as a screen which tends to subdue the amplitude of the central anomaly in this trend. Cross sections taken through the three magnetic highs show this change in the central magnetic high relative to the other two (Figure 9). Most notable is the increase in the width at half amplitude of the central anomaly (Section B-B^{1}) which shows that it originates at a greater depth (below the dunite) than the anomalies to the north and south. Depth estimates based on both the width at half amplitude and on Peters's slope method (Dobrin, 1976) show that the source in the country rock causing the central anomaly is on the order of 100 meters below the surface as compared to the very shallow sources (20 meters) for the northern and southern anomalies. The 100-meter depth, then, is related to the thickness of the dunite body in this area, and is consistent with reports of the thin, rootless character of dunite bodies elsewhere in the area (Honeycutt *et al.*, 1981; Perez, 1979).

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SEDIMENTOLOGY OF A CARBONATE-RED BED ASSOCIATION, MISSISSIPPIAN GREENBRIER GROUP, EASTERN WEST VIRGINIA

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ABSTRACT

Sandy, oolitic, and dolomitic limestones of the Greenbrier Group constitute one of the major gas and oil reservoirs in western West Virginia. Similar facies exposed in eastern outcrops provide accessible rocks for petrologic and sedimentological studies that supplement depositional and diagenetic information about these reservoirs. The Greenbrier sediments of Spruce Knob Mountain, Pendleton County, were deposited in the eastern portion of a large, slowly subsiding basin in which a shallow-water carbonate platform developed. The deeper slope of the platform was characterized by three limestone microfacies, all of which accumulated under well oxygenated, low- to moderate-energy conditions. Abundantly fossiliferous biomicrite and biopelmicrite formed below wave base, whereas well-washed biopelsparite occurred above wave base. The platform-margin oosparite and sandy pelsparite of the open platform lagoon represent very shallow, high- to moderately high-energy conditions. Pelmicrite developed in a shallow, restricted portion of the lagoon. Periodic clastic influxes from eastern sources produced red, calcareous shale which interfingered with laterally adjacent pelmicrite in the marine lagoon.

INTRODUCTION

Interbedded red shales and limestones are not uncommon throughout Paleozoic rocks of the Appalachians. The Cambrian Rome Formation, the Ordovician Sequatchie and Moccasin Formations, the Silurian Wills Creek Formation, and the Greenbrier Group of Late Mississippian (Meramecian and Chesterian) age (Arkle and others, 1979) are a few notable examples of this close association. These and other occurrences reflect similar depositional environments of shallow-marine carbonates and transitional-marine clastics. However, petrologic studies generally focus solely upon the carbonate units, virtually ignoring the closely associated red beds.

The Greenbrier Group, exposed in eastern West Virginia, consists primarily of limestones but contains shale, siltstone, and sandstone interbeds which have been used to subdivide the unit. Previously, petrologic and paleontologic investigations have concentrated on the carbonate formations (McCue, Lucke, and Woodward, 1939; Martens and Hoskins, 1948; Rittenhouse, 1949; Wells, 1950; Wray, 1952; Youse, 1964; and Leonard, 1968). In Pendleton County, however, field and petrologic studies reveal important depositional similarities between shallow-lagoonal limestone (pelmicrite) and the interbedded red shale (Wray, 1980).

Furthermore, these microfacies belong to a carbonate-platform environment in which lateral facies equivalents have been found to contain economic reserves of gas. In western portions of the state, the Greenbrier limestones constitute a major gas and oil reservoir (Haught, 1959; Ruley, 1970). The eastern outcrop belt provides accessible

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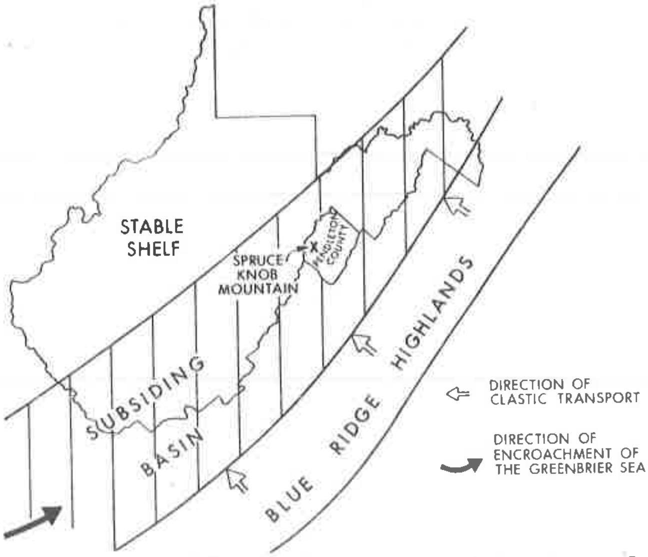

Figure 1. Map of West Virginia showing location of Pendleton County as well a. tectonic structures present during the Mississippian Period. (Modified from Arkle 1974; Donaldson, 1974; and Leonard 1968).

exposures and large-scale field relations that supply supplemental data for interpretation of the gas-producing subsurface facies.

STRATIGRAPHY AND GEOLOGIC SETTING

In the Spruce Knob region of Pendleton County, West Virginia (Figure 1), Greenbrier stratigraphy clearly illustrates alternating carbonate and clastic deposition (Figure 2). Sedimentation patterns during Mississippian time were primarily controlled by existing tectonic features of the Appalachian basin (Figure 1). Following early to middle Paleozoic structural developments, primarily in response to tectonic pulses from the east (notably orogenic deformation generated by plate divergences and convergences; Thomas, 1977), the paleogeographical stage was set for Late Mississippian sedimentary accumulations within a shallow carbonate sea. Encroachment of this epeiric sea was from the southwest, and at first it was confined to a large, slowly subsiding basin (Colton, 1970). With time, the sea transgressed northward to cover the adjacent stable shelf (Leonard, 1968). The Greenbrier sediments were deposited on a broad carbonate platform in this basin and across the stable shelf. Periodic invasions of the shallow sea by terrestrial sediments were from the east and/or southeast (Thomas, 1977). The resultant interbedding of shallow-marine carbonates and transitional-marine clastics is clearly distinguishable in outcrop (Figure 3). Factors such as eustatic sea-level changes, basin subsidence, rate of carbonate deposition, and amount of terrigenous influx probably all contributed to the lateral and vertical arrangement of the carbonate-platform sediments (Figure 4).

METHODS

Ninety-five meters of exposed Greenbrier Group, the total thickness on Spruce Knob Mountain, were measured, described, and sampled at one complete and two partial exposures. Thin sections were prepared from sixty-nine samples collected at a 1.5-meter interval and where important lithologic changes were observed. After three hundred points were counted for each thin section, Folk's (1959) classification system was used to identify and group the eight recognized microfacies. X-ray diffraction and whole-rock elemental analyses were most useful in substantiating the presence and varieties of clay minerals.

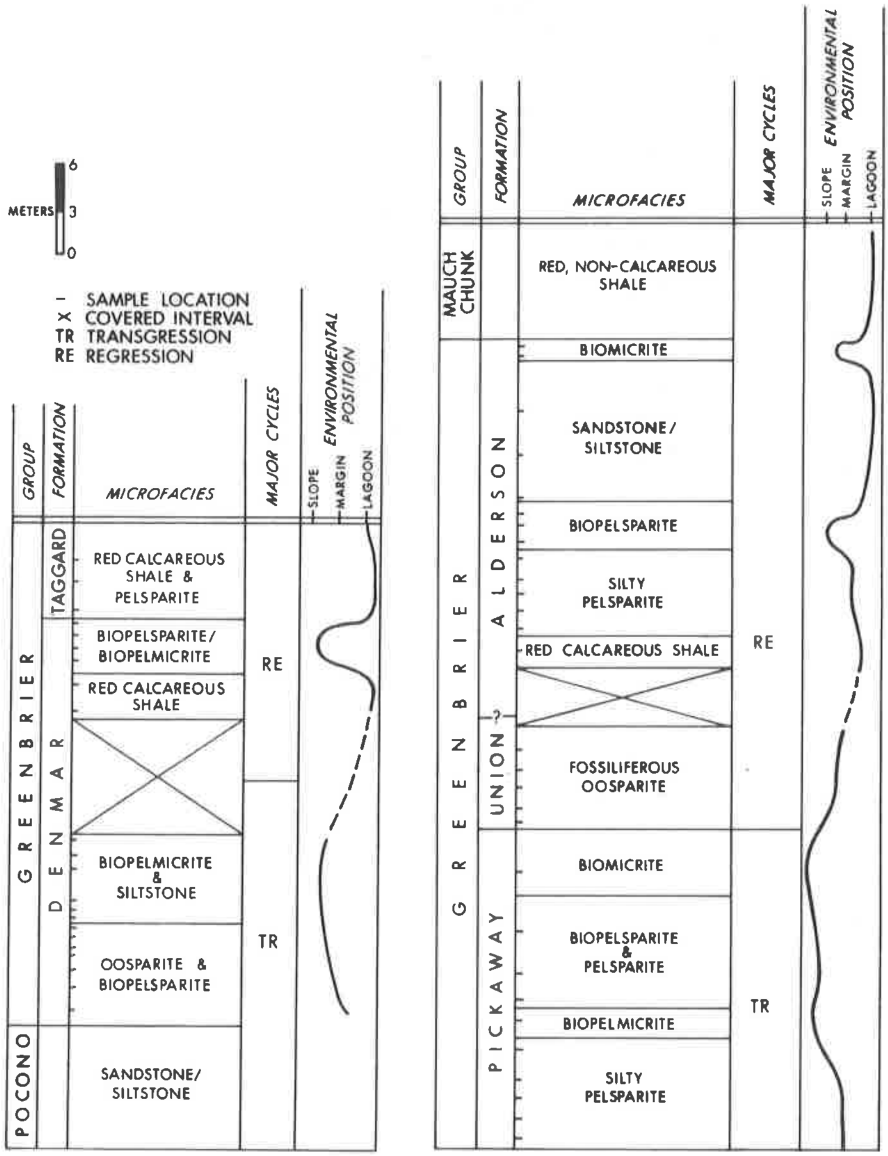

Figure 2. Generalized lithostratigraphic units, sample locations, and inferred transgressive-regressive sequences of the Greenbrier Group, Spruce Knob Mountain, Pendleton County, West Virginia.

MICROFACIES AND DEPOSITIONAL MODEL

In the paleoenvironmental model suggested for Greenbrier sedimentation in Pendleton County (Figure 4), eight microfacies represent distinct environmental conditions across a broad, shallow carbonate platform. The model is based upon Wilson's concepts of carbonate facies belts (1974, 1975). Variables considered in the designation of each environment include relative values of hydrodynamic energy, water depth, circulation, faunal communities, fossil diversity, and amount of terrigenous influx.

Three platform-slope microfacies accumulated under normal-marine conditions: biomicrite, biopelmicrite, and biopelsparite. These lithologies have petrologic affinities

Figure 3. Differential weathering in the Greenbrier sediments produced as a result of interbedded limestone (projecting layers) and red calcareous shale (recessed layers).

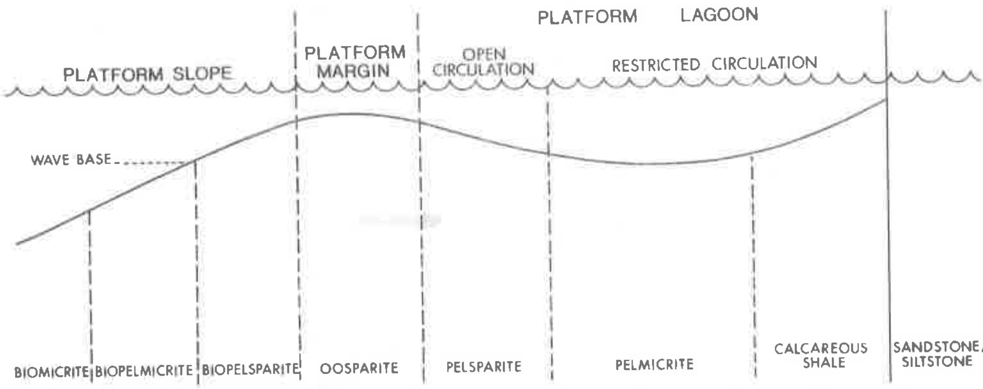

Figure 4. Depositional model of the Greenbrier Group, Spruce Knob Mountain, Pendleton County, West Virginia.

with one another but can be distinguished by textural differences related to hydrodynamic energy and water depth. Of all the microfacies, these three contain the most abundant remains of diverse infaunal and epifaunal organisms, including pelmatozoans *(Pentremites)* and echinoids, fenestrate bryozoans, brachiopods *(Spirifer, Productid, Chonetes,* and *Composita),* as identified by hand specimens, endothyrid foraminifera, high-spired gastropods, and pelecypods (Table 1). Selected fossils, primarily brachiopods and mollusks, were commonly micritized by endolithic algae (Bathurst, 1975, p. 381). The biomicrite (Figure 5A) contains no ooids, pellets, or intraclasts, though these non-skeletal grains are very significant to the composition of the biopelsparite (Figure 5B) and somewhat less so to the biopelmicrite (Figure 5C). Terrigenous material is rare although intercalated shales are seen in the field. Sorting is generally poor, suggesting little if any transport prior to deposition.

Fossiliferous oosparite (Figures 5D and 6A), averaging 48% ooids and containing nearly 7% echinoderm and bryozoan remains, characterized the shallow, highly turbulent margin of the carbonate platform. The amount of terrigenous material is the

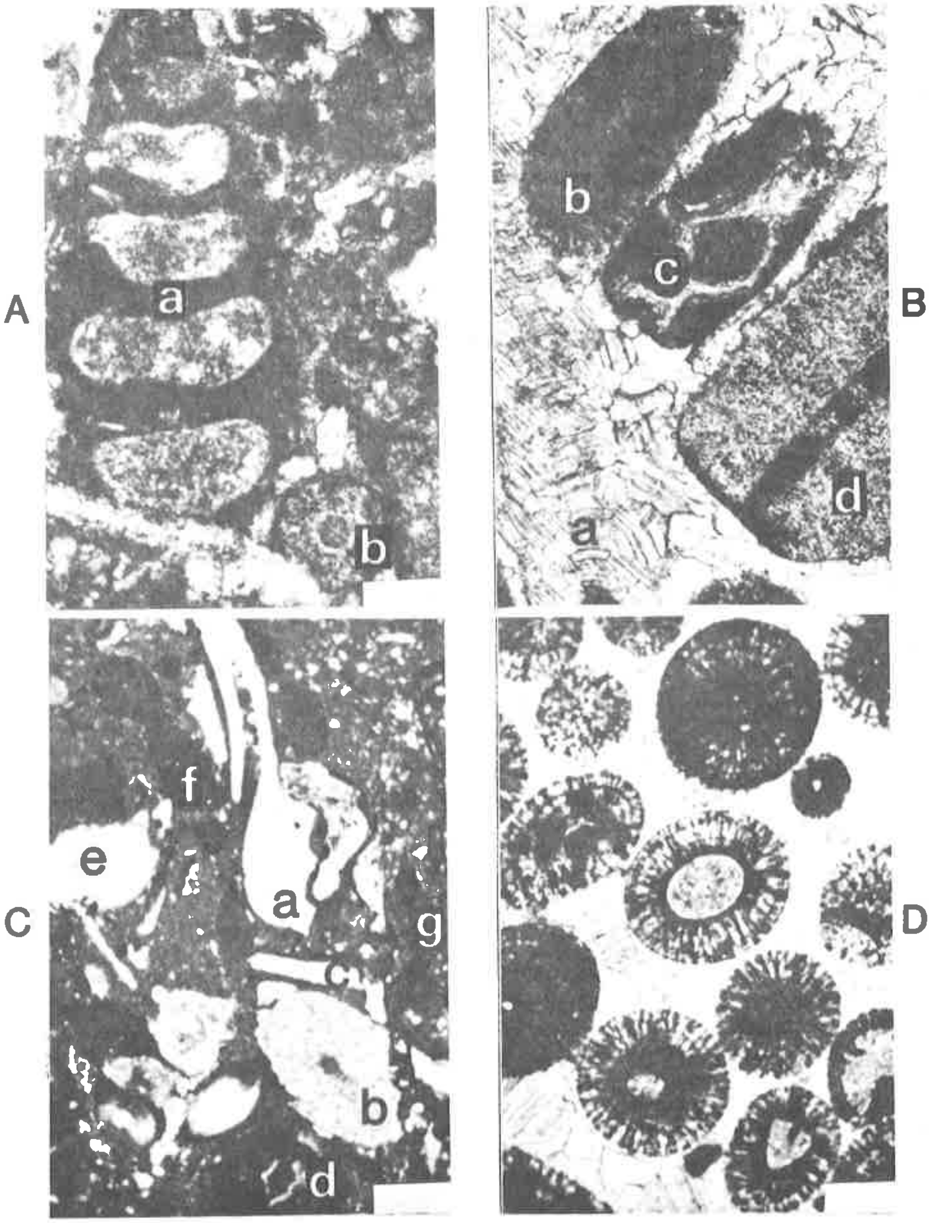

Figure 5A. High-spired gastropod (a) and pelmatozoan columnal, (b) preserved within micrite matrix, biomicrite facies. The skeletal walls of the gastropod are micritized and the chambers are filled with both micrite and sparry cement. The pelmatozoan columnal has a thin micritic envelope. Bar scale 0.1mm.

Figure 5B. Recrystallized mollusk shell (a), micritized ooid (b), bryozoan intraclast (c), and echinoderm plate (d) of biopelsparite facies. Bar scale 0.1mm.

Figure 5C. Trilobite fragment (a), echinoderm plate (b), scattered mollusk fragments (c), an ooid (d), spar-filled fossil molds (e), and pellets (f) are incorporated within a pelletal micrite matrix, biopelmicrite facies. A stylolitic seam (g) is delineated by the accumulation of dark, insoluble minerals. Bar scale 0.25mm.

Figure 5D. Well-developed ooids from the oosparite facies, Union formation, bound by rim and equant cements. Both concentric and radial textures within the ooid cortices are clearly developed. Bar scale 0.25mm.

Table 1. Average percentages of carbonate and non-carbonate constituents and relative occurrences of textural features in the Greenbrier Group microfacies, Spruce Knob Mountain, Pendleton County, West Virginia. A=Abundant, C=Common, R=Rare, Tr-Trace.

Constituents	Bio-micrite	Biopel-micrite	Biopel-sparite	Oosparite	Pel-sparite	Pel-micrite	Shale	Sandstone/Siltstone
Ooids	--	0.1	8.9	48.7	1.3	--	0.7	--
Pellets	6.7	14.4	18.3	4.6	25.7	51.7	12.8	1.4
Intraclasts	2.3	0.5	0.7	1.8	0.1	1.0	--	--
Echinoderms	8.6	6.7	11.3	4.3	2.6	1.8	3.4	--
Bryozoans	1.6	3.1	3.4	1.9	0.1	0.7	0.7	--
Brachiopods	0.9	0.8	0.8	0.1	--	0.7	0.1	--
Foraminifera	3.1	2.3	1.3	0.1	0.1	0.9	0.1	--
Arthropods	0.2	0.4	0.3	0.5	0.1	0.5	--	--
Mollusks	2.0	0.8	1.9	1.1	--	0.2	0.3	--
Calcite cement	6.9	5.1	29.9	29.6	23.9	4.4	7.2	4.9
Micrite matrix	54.1	48.5	13.0	7.1	6.9	27.9	3.0	2.5
Quartz	2.2	4.2	4.7	0.2	29.5	2.7	22.3	73.7
Clays	5.5	1.1	0.3	--	3.7	--	40.0	8.5
Other carbonates	0.2	1.3	--	--	--	7.1	--	--
Other non-carbonates	5.7	10.7	5.2	--	6.0	0.4	9.4	9.0
Micritization	A	A	A	A	C	A	A	--
Cross-bedding	--	--	--	C	R	--	--	--
Bioturbation	--	R	R	R	R	C	A	--
Sorting	Poor	Poor	Poor	Good	Good	Poor	Poor	Moderate

least of any microfacies (Table 1). Moderate to good sorting, cross-bedding, and grain abrasion reflect the very high hydrodynamic energy of this environment. "Rip-up" intraclasts of adjacent sediments formed when storms passed over the platform margin, but the presence of some grapestone grains attests to occasional periods of quiescence (Bathurst, 1975, p. 316). Likewise, alternating laminae of coarse and fine ooids point to fluctuating current strength. During initial burial, grains were attacked by endolithic algae (micritized ooids). With deeper burial, compaction often resulted in spalled oolitic cortices and pressure welding.

The pelsparite microfacies (Figure 6B) formed behind the oolitic platform margin in a lagoon subject to circulating currents. Hydrodynamic energy in this environment was markedly less than in the oosparite regime as suggested by fewer ooids (washed in by currents) and intraclasts as well as by a greater number of pellets. Good sorting, cross-bedding, grain abrasion,and clay galls, however, do indicate substantial current activity. Echinoderms are the only significant fossil, though they are not abundant. Quartz is most prevalent in this carbonate microfacies and terrigenous clays, primarily illite and kaolinite, are common.

Pelmicrite (Figure 6C) and its terrigenous counterpart, calcareous red shale (Figure 6D), accumulated within restricted portions of the platform lagoon. Pellets are abundant in the pelmicrite microfacies (51.7%) and less so in the shale (12.8%). Quartz silt is prevalent in the shale, and terrigenous clays form the matrix, whereas in the pelmicrite, carbonate mud is the matrix. The shale is laminated, consisting of alternating layers of large, tightly packed carbonate grains (mostly pellets and fossils) and layers of clays, quartz, and micrite. Echinoderms, fenestrate bryozoans, punctate and pseudopunctate brachiopods, and endothyrid foraminifera are present but rare in both lithologies. Both microfacies exhibit poor to moderate sorting and some grain abrasion. Total micritization of many grains, micritic envelopes around the periphery of others, and bioturbation of the sediment by infauna are all very abundant and collectively suggest that the sedimentation rate was slow (Brathwaite, 1966).

The pelmicrite and red shale appear to have been facies equivalents in the lagoon.

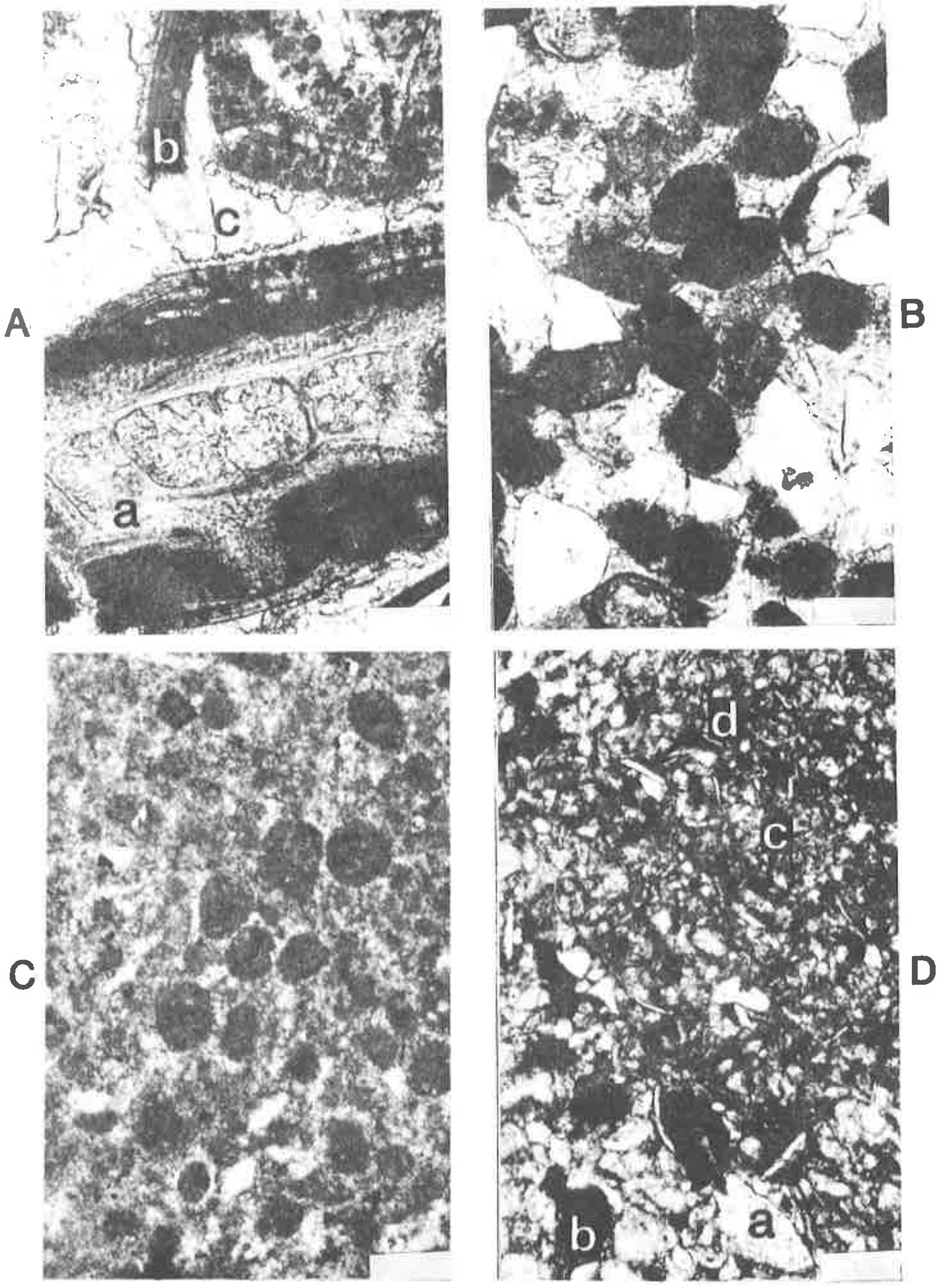

Figure 6A. Oolitic coating on bryozoan fragment (a), spalled ooid lamellae with rim cement (b), and equant calcite cement (c), oosparite facies. Bar scale 0.1mm.
Figure 6B. Pellets and subangular to subround quartz grains within sparry calcite cement, pelsparite facies. Bar scale 0.1mm.
Figure 6C. Pellets with both distinct and fuzzy outlines and spar-filled mold (upper left) surrounded by a micrite matrix, pelmicrite facies. Bar scale 0.25mm.
Figure 6D. Subangular quartz grains with overgrowths (a) and carbonate pellets (b) with matrix of clays (c) and mica grains (d), shale facies. Bar scale 0.1mm.

The close proximity of carbonate and clastic environments allowed some sediment-mixing to occur which resulted in the observed petrologic similarities. In some cases, both the pelmicrite and the red shale contain a mixed argillaceous and micritic matrix in addition to both carbonate and non-carbonate grains.

The sandstone/siltsone microfacies represents significant clastic pulses during Late Mississippian time. These well-sorted, cross-bedded quartz sediments, devoid of any fossils, lack definitive sedimentological evidence to suggest their exact depositional environment. In terms of stratigraphic relationships, these sediments appear to be an integral part of the predominantly shallow-marine environment. What remains to be established are the precise paleogeographical conditions in which deposition occurred.

STRATIGRAPHIC CHANGES

Lithologic changes within the Greenbrier Group reflect shifts in lateral facies brought on by one or more situations: eustatic changes, basin subsidence, rate of sedimentation, or orogenic uplift. Whatever the cause of the apparent transgressions and regressions, resultant depositional patterns can be examined. Each of the five Greenbrier formations shown in Figure 2 has been subdivided into its characteristic microfacies. Based upon the sequence of microfacies in the stratigraphic column, two major cycles of environmental fluctuations have been recognized.

The Denmar formation was deposited during a major transgression in which marine carbonates onlapped terrestrial/transitional-marine Pocono sands below (Dally, 1956). Denmar sediments consist of thirty meters of biopelsparite and biopelmicrite with oosparite and siltstone towards the base and red shale near the top. The oosparite contains superficial ooids that are not as well developed as normal platform-margin ooids; therefore, the lower turbulence involved during Denmar deposition may indicate that a distinct platform margin was not yet fully developed. The Upper Denmar sediments reflect the beginning of a slight regression.

The Taggard formation represents progradation of lagoonal sediments. Red calcareous shale dominates this six-meter-thick unit, but pelsparite is present at the base.

Once again, though, a predominantly transgressive pattern emerged during deposition of the twenty-two meters of the Pickaway formation. Platform-lagoon pelsparite is overlain by the various platform-slope sediments, documenting the return of the shallow epeiric sea. Water depth was probably deepest during this period.

Development of a definite platform margin, marked by an eight-meter-thick oosparite deposit, characterized sedimentation during Union time. True ooids of the oosparite reflect shallow, high-energy conditions which continued for some time, indicating a balance between carbonate accumulation and basin subsidence.

Finally, carbonate deposition drew to a close with subsequent progradation of Mauch Chunk shales. The twenty-nine-meter-thick Alderson formation consists of lagoonal pelsparite and red shale, sandstone, and only a few platform-slope carbonates. These clastics represent the first influx of terrigenous material from the Late Mississippian Mauch Chunk delta complex (Hoque, 1968). With the one exception of deposition of the Reynolds Limestone within the lower Mauch Chunk Group, epeiric seas never again dominated this area.

CONCLUSIONS

Eight Greenbrier microfacies are characterized in terms of their distinct depositional environments. These environments have been compared to those described by Wilson (1974, 1975) as belonging to a broad carbonate platform. The microfacies which developed during Greenbrier deposition include biomicrite, biopelmicrite, biopelsparite, oosparite, pelsparite, pelmicite, shale, and sandstone/siltstone.

The envisioned Greenbrier depositional model depicts a shallow, relatively near-shore marine environment into which varying amounts of detrital material were shed. Components of this environment included: 1) a broad, shallow, carbonate platform with a seaward slope on which the biomicrite and biopelmicrite accumulated below wave base and the biopelsparite above wave base; 2) a platform margin on which the oosparite formed; and 3) a shallow platform lagoon with both regions of good circulation in which the sandy pelsparite developed and regions of restricted circulation in which the pelmicrite formed. The shale represented low-energy clastic invasions of the lagoonal environment. These pulses of terrigenous material, well represented in the Greenbrier Group by the calcareous shale and the sandy and argillaceous carbonates, were part of the depositional pattern of alternating carbonate and clastic influence that was pervasive throughout the Mississippian Period.

Greenbrier limestones mark a major transgression over Lower Mississippian terrestrial sandstones and siltstones of the Pocono Group. No distinct carbonate-platform margin was yet developed during deposition of the Denmar Formation. Interbedded fossiliferous limestone of the platform slope, lagoonal pelsparite, and fine terrigenous clastic rocks in the Denmar, Taggard, and Pickaway Formations denote two major cycles of transgression and regression. The last regression is indicated by the platform-margin oosparite of the Union Formation, lagoonal sediments of the Alderson Formation, and deltaic sediments of the overlying Mauch Chunk Group.

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STRUCTURAL POSITIONS OF SEVERAL ULTRAMAFIC BODIES IN WESTERN NORTH CAROLINA

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ABSTRACT

An investigation of six ultramafic bodies in western North Carolina suggests that their present mineral assemblages are structurally controlled. Structural analysis shows that the enclosing country rocks are isoclinally folded synforms and antiforms with steeply dipping, overturned axial planes. Two deformational episodes are recognized: (1) an early phase which produced isoclinal folds; and, (2) a later phase which generated more open folding. One completely serpentinized ultramafic body is structurally located directly upon a synformal axis. Five partially serpentinized ultramafic bodies are located between fold axes on the fold flanks.

It is postulated that the present mineral assemblages of the ultramafic rocks are syntectonic with the second folding episode, whereby pressure gradients established during the second deformational phase caused diffusion of solutions into low-pressure fold trough zones bringing about varying degrees of hydration of the ultramafic mineral assemblages, including the complete serpentinization of one ultramafic body.

INTRODUCTION

The area investigated is within the Blue Ridge province-ultramafic belt of the Southern Appalachians (Figure 1). It encompasses 390 sq. km. of Yancey County, North Carolina. This ultramafic belt is generally characterized by the presence of ultramafic bodies exhibiting varying degrees of alteration. The structural positions of six ultramafic bodies were investigated to evaluate possible relationships between the ascertained structural positions and the present ultramafic mineral assemblages.

PETROGRAPHY

Country Rock

Two major lithologic units were observed, the older Cranberry Gneiss and the overlying Ashe Formation. The ultramafic rocks of this investigation are within the Ashe Formation (Figure 2).

Thin section analyses of eight Cranberry Gneiss samples show prominent inequigranular muscovite, biotite, quartz, plagioclase, alkali-feldspar, and garnet. Rankin (1970) describes the Cranberry Gneiss as mostly a sheared, recrystallized quartz monzonite grading into flaser-gneisses and augen-gneisses. Small gabbroic dikes cut the Cranberry Gneiss near the overlying Ashe Formation. The Cranberry Gneiss underlies upper Precambrian rocks and has been radiometrically dated at 1,050 m.y. (Davis and others, 1962; Odom and Fullagar, 1971).

Seventy thin sections reveal that the Ashe formation consists of fine to medium grained biotite-muscovite gneisses, amphibole gneisses (plagioclase = An_{38}), and quartz-muscovite schists containing biotite and plagioclase (An_{20}). The An-percent was determined by extinction angle techniques described by Moorehouse (1959). Minor amounts of garnet, kyanite, epidote, sphene, and magnetite are present also. Modal analyses show that these rock types vary both regionally and locally, and tend to grade into one another laterally and vertically. Rankin (1970) interprets the amphibole gneisses as metamorphosed upper Precambrian mafic volcanics and penecontemporaneous shallow intrusions, and the mica gneisses and schists as metamorphosed sulfidic

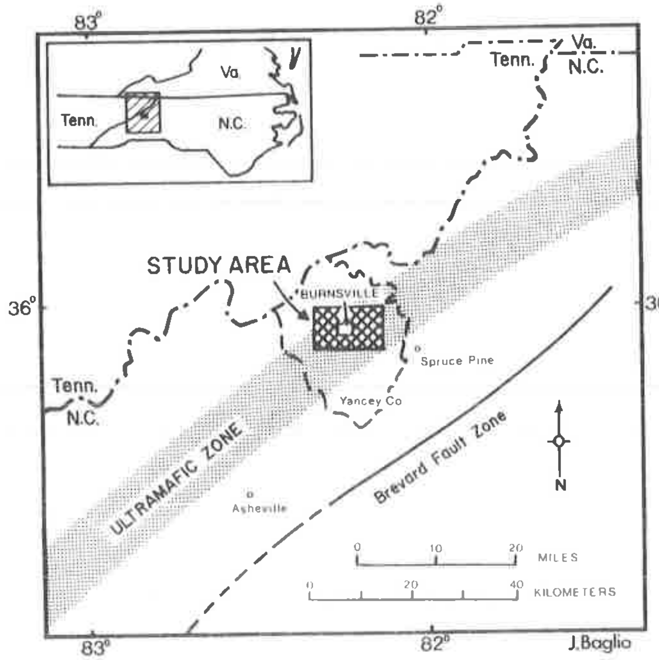

Figure 1. Index map of study area.

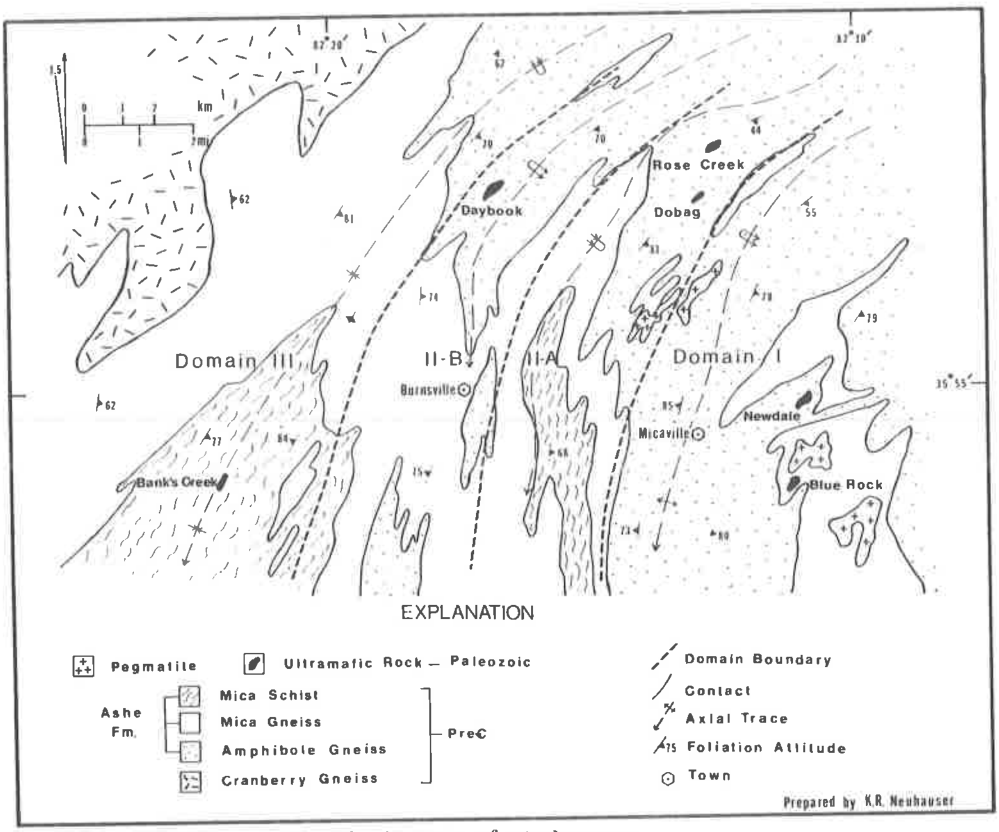

Figure 2. Structural and geologic map of study area.

sandstones and quartz-feldspar rich shales. As indicated in Figure 2, predominant mica schists surround the Bank's Creek serpentinite (Neuhauser and Carpenter, 1971), whereas amphibole gneisses primarily surround four of the other five partially serpentinized ultramafic bodies. The Blue Rock ultramfic body is primarily surrounded by mica gneisses, and is slightly more serpentinized than the amphibole-gneiss enclosed bodies.

Numerous pegmatites cut the Ashe Formation in the eastern one-third of the study area. They vary greatly in outcrop width, from a few centimeters to several hundred meters. They consist of alkali-feldspar, quartz, and muscovite with minor garnet and magnetite (Brobst, 1962; and Butler, 1973). Radiometric analyses of the pegmatites yield K/Ar and Rb/Sr ages ranging from 300 to 430 m.y. (Olson, 1944; Lesure, 1968; Odom and Fullagar, 1973).

Ultramafic Rock

The size and shape of the six ultramafic bodies described herein are based on available outcrops and float. All six bodies are symmetric to slightly asymmetric ellipsoids in plan view (Figure 2). The Bank's Creek serpentinite measures approximately 90 by 225 meters and its long axis is parallel to the regional strike of the Ashe Formation. Four samples collected from the central part of the body to its margin exhibit in thin section a mesh-like texture with reticulate fibers of serpentine occurring as pseudomorphs after equant olivine grains with 120° triple junctions. Irregular curved fractures containing fine-grained magnetite are found within and along the serpentine pseudomorphs. X-ray diffraction analyses of three separate Bank's Creek samples show the pseudomorphs to be antigorite and the opaques to be magnetite. Cutting the serpentinite body are numerous thick (5 to 10 mm.) veinlets of randomly oriented chrysotile fibers. The Bank's Creek body is completely serpentinized (Neuhauser and Carpenter, 1971).

The long axes of the partially serpentinized ultramafic bodies are generally parallel to the regional strike of foliation, including the major fold axes of the Ashe Formation (Figure 2). The Day Book body is the largest exposure measuring 180 by 600 m. The Dobag body is the smallest at 53 by 150 m., with the Newdale, Blue Rock and Rose Creek ultramafic exposures being intermediate in size. Magmatic differentiation zones and contact metamorphic aureoles are not associated wth any of the six ultramafic bodies.

Each ultramafic body contains relatively unaltered olivine as equant, polygonal, recrystallized grains exhibiting 120° triple junctions. Grain size diameters range from 0.1 to 1.0 mm., and average 0.3 mm. Carpenter and Phyfer (1975) report that a magnesian olivine (Fo_{91}-Fo_{94}) is found in approximately ninety percent of the ultramafics in this belt. Swanson (1980) reports that the Day Book olivine grains are homogeneous and lack compositional zoning. Similarly, no significant variation in olivine composition exists within the Day Book body (Carpenter and Phyfer, 1975). In the Day Book body and Woody Asbestos Mine body (located roughly one-quarter mile to the NE of the Dobag and Rose Creek bodies), porphyroclasts of olivine show undulatory extinction, elongation, and granulation (Kingsbury and Heimlich, 1978; Swanson, 1980). Kingsbury and Hemlich (1978) believe these olivine porphyroclasts are primary minerals with a mantle source.

Other minerals present in the ultramafic bodies are chromite, orthopyroxene, talc, and serpentine. The chromite occurs as disseminated euhedral to subhedral grains commonly surrounded by kammererite (a chromium chlorite), and as more massive deformed lenses composed of large anhedral, highly fractured grains that are embayed by olivine and kammererite (Carpenter and Fletcher, 1979). The Day Book body exhibits minor secondary huntite, aragonite, magnesite, chlorite, anthophyllite, and vermiculite (Tien, 1977; Swanson, 1980).

The ultramafic bodies exhibit a higher concentration of serpentine and talc at their peripheries, and each contains some talc as isolated grains and small veins within the interior. Assuming serpentinization proceeds from the exterior of the ultramafic body to the core, the process does not appear to be controlled by the size of the body, as exemplified by the small Dobag body which is peripherally serpentinized and the larger, completely serpentinized Bank's Creek body.

STRUCTURE OF THE ASHE FORMATION

Method

Orientations of foliation, lineation, axial plane, and drag fold elements were measured with a Brunton compass. The planar and linear elements were plotted on standard 20 cm. Schmidt stereographic nets. Structural interpretations of the projection data were based on domainal methods by Turner and Weiss (1973).

Observation

Field observation shows that the schistose and gneissose Ashe Formation units are intensely deformed. At several exposures the foliation is an axial plane foliation rather than a bedding foliation. Competent, thin arenaceous units locally show evidence of early, tight isoclinal folding. Associated well-foliated micaceous and poorly-foliated amphibolitic units do not show evidence of the tight folding. The compositional layering of the folded competent units is interpreted to be a "bedding" foliation. As a result of an intense deformational episode, the bedding foliation in the less competent micaceous and amphibolitic units was destroyed. Subsequently, foliation developed in

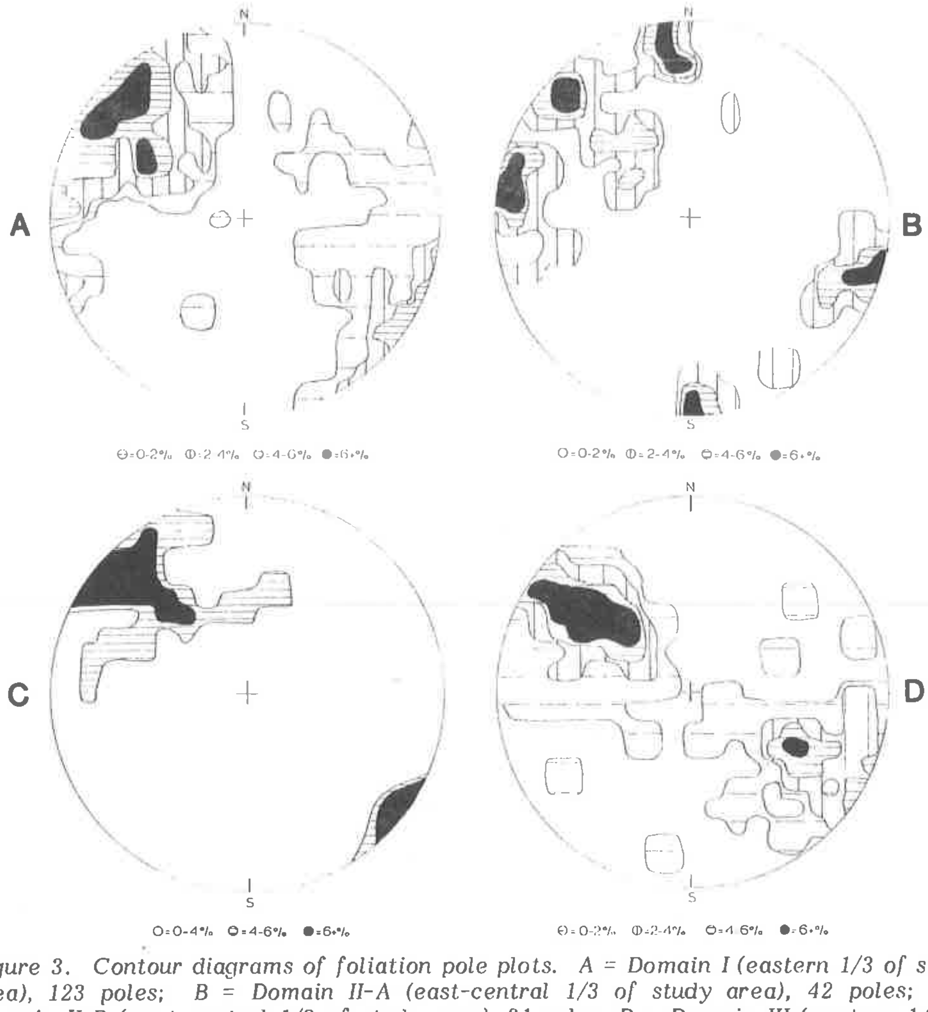

Figure 3. Contour diagrams of foliation pole plots. A = Domain I (eastern 1/3 of study area), 123 poles; B = Domain II-A (east-central 1/3 of study area), 42 poles; C = Domain II-B (west-central 1/3 of study area), 21 poles; D = Domain III (western 1/3 of study area), 140 poles.

the micaceous and amphibolitic units parallel to the axial planes of the tight isoclinal folds in the competent units. Therefore, two planar fabric elements are present. They are defined as: S_1 = compositional meta-arenites and mica-schists, and S_2 = axial plane foliation in mica-schists and amphibole gneisses. Nearly all foliation surfaces measured are S_2-foliation. Few S_1-foliations were found.

Two linear elements are present. Lineations exhibited by the axes of tight isoclinal folds within compositional layers in (S_1) are referred as to as L_2. Superimposed on the S_1 and S_2 surfaces of both the arenaceous and micaceous units,and occasionally on the amphibolitic units, is the second linear element in the form of more open fold axes (crenulations) with an amplitude of 0.2 to 12 cm. This linear element is defined as L_3 and is taken to imply a later episode of deformation. There appears to be no S_3 surfaces related to this implied later deformational event.

Interpretation

Within the eastern one-third (Domain I) of the study area, 123 measurements of S_2 define a slightly asymmetric, tightly folded, overturned antiform striking N36°E, plunging 31°SW, and an axial plane dipping 76°SE (Figure 3-A). Sixty-three S_2 plots from the central one-third of the area indicate two separate fold axes. The easternmost (Domain II-A) fold is an asymmetric, tightly folded, overturned synform with a strike of N44°E, plunging 19°SW, and an axial plane dipping 78°SE (Figure 3-B). The westernmost (Domain II-B) fold is an asymmetric, tightly overturned, folded antiform with a strike of N37°E, a plunge of 14°SW, and an axial plane dipping 74°SE (Figure 3-C). In the western one-third (Domain III), 140 measurements of S_2 define a nearly vertical, tightly folded synform striking N33°E, plunging 12°SW, and an axial plane dipping 86°SE (Figure 3-D). Further field inspection of 104 drag fold axial planes in the four domains of the study area indicate that the major fold axial planes are more symmetrical and vertical in the southern one-half, whereas in the northern one-half they are more asymmetrical and slightly overturned, dipping steeply to the southwest.

Forty-eight measurements of the linear element L_2 within the Ashe Formation and Cranberry Gneiss exhibit random distribution with steep plunges (Figure 4-A). Plots of 125 L_3 elements exhibit two maxima near the horizontal, trending N30°E-S30°W with a plunge variation of 15° to 20° (Figure 4-B).

The ultramafic bodies occupy different structural positions with regard to the fold axes and the enclosing rock type within the Ashe Formation. The Bank's Creek serpentinite lies directly upon the fold axis of a synformal structure and is surrounded primarily by quartz-muscovite schists. The Blue Rock body is located on a synform fold flank and is primarily surrounded by mica gneisses. The remaining ultramafic

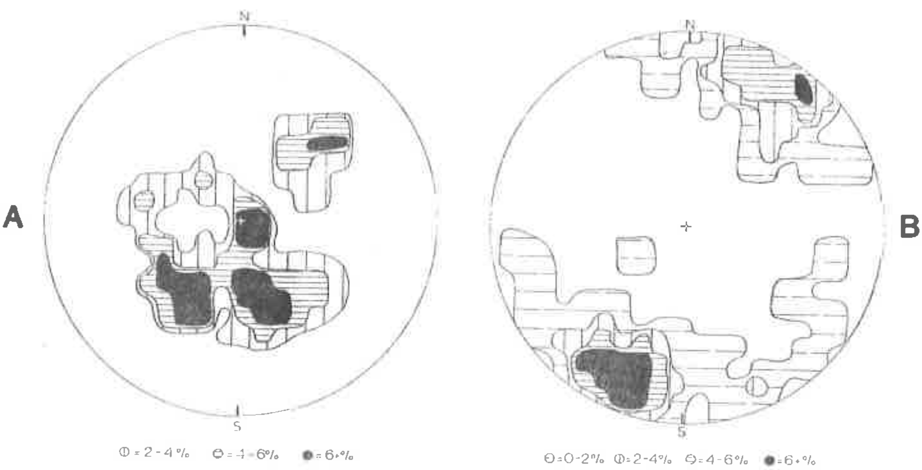

Figure 4. Contours of lineation elements (all domains): A = L_2, 48 points. B = L_3, 125 points. (Note: Lineation element contours for individual domains indicate same contour pattern.)

bodies are located on the fold flanks of synforms and antiforms, and are primarily surrounded by amphibole gneisses.

DISCUSSION

Hypotheses explaining the presence of essentially juxtaposed hydrous and anhydrous ultramafic rocks in this study area should be consistent with the following observations:

(1) The Bank's Creek serpentinite is apparently the result of the *in situ* hydration of a pre-existing olivine-rich rock (Neuhauser and Carpenter, 1971).

(2) The Bank's Creek serpentinite lies structurally on a fold axis. The other ultramafic bodies are located on synform and antiform fold flanks.

(3) The Ashe Formation has been subjected to at least two folding episodes, and a metamorphic thermal maximum of 500^{o}-$600^{o}C$ (Phyfer, 1968).

At least two episodes of folding affected the ultramafic rocks exposed in Yancey County, North Carolina. It is suggested that the following sequence led to the present field conditions. First, an intense compressional episode, probably conincident with the metamorphic thermal maximum, produced the planar and linear fabric elements S_1, S_2, and L_2. This early deformational event probably peaked during the Odovician or middle Devonian (Hatcher, 1972; Cook and others, 1980). Within the Blue Ridge province this major episode of regional metamorphism occurred approximately 430 m.y. ago (Butler, 1973). The Day Book ultramafic body dates older than middle Paleozoic for it is cut by a 375 m.y. old pegmatite (Kulp and Brobst, 1954). If ultramafic bodies such as the Day Book represent subduction zone emplacements of oceanic crust or upper mantle fragments, as suggested by Kingsbury ahd Heimlich (1978), the approximate emplacement time of the Yancey County bodies would have been just prior to, or early in, this first major metamorphic event called the Taconian orogeny. The deformational texture of the relict olivine and orthopyroxene grains seems to suggest a cold, solid-state mode of emplacement of deformed, mantled-derived ultramafic material. The emplacement could have occurred during the close of the basin between the Inner Piedmont-Blue Ridge fragment and the proto-North American plate between 450-550 m.y. ago (Cook, and others, 1980). The formation of the olivine polygonal-recrystallization texture of the ultramafic bodies is believed to have accompanied this first thermal maximum event.

A second less intense deformational episode, dated at 380 m.y. by Butler (1973), caused the development of more open folding which affected all foliations (S_1 and S_2), and the creation of the linear element L_3. This episode would accompany Cook's (1980 theory of the closing between the Inner Piedmont-Blue Ridge fragment and the Carolina Slate Belt fragment. It is felt that serpentinization of the ultramafic bodies occurred during this second folding episode when pressure gradients established in the country rocks led to the migration and concentration of water in low-pressure fold hinges. According to this view, olivine-rich rocks occurring in low-pressure structural sites and within quartz-muscovite rich schists, would be subjected to a higher water concentration with an activity of water adequate to bring about complete hydration to serpentinite as exemplified by the Bank's Creek body, located directly on a fold hinge. This serpentine process would be possible if the temperature was at, or below, the thermal hydration temperature for the reaction: olivine + talc + water = serpentine. Olivine-rich rocks located on fold flanks and within the amphibole gneisses would experience a lower water concentration, and thus a lower water activity, and therefore would undergo peripheral serpertinization, as evidenced by the five partially serpentinized ultramafic bodies analyzed in this study area.

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GEOLOGY AND GEOCHRONOLOGY OF THE PARKS CROSSROADS GRANODIORITE,

NEAR SILER CITY, CENTRAL NORTH CAROLINA PIEDMONT

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ABSTRACT

A calc-alkaline hornblende granodiorite crops out in the Parks Crossroads area between Siler City and Asheboro, North Carolina, where it intrudes a thick sequence of metamorphosed felsic tuffs, rhyodacite and dacite flows, volcanic comglomerates, sandstones, siltstones, and argillites situated on the southeastern flank of the Troy anticlinorium. Part of that sequence is correlated here with the early to middle Cambrian Uwharrie Formation. Five granodiorite samples give a Rb-Sr whole rock isochron age of 566 $\pm$ 46 m.y. and an initial $^{87}Sr/^{86}Sr$ ratio of 0.7028 $\pm$ 0.0006, which correlates well with the age and initial Sr ratios determined for other similar plutons in the Carolina slate belt.

INTRODUCTION

A stock of medium-grained equigranular hornblende granodiorite crops out along US 64 between Asheboro and Siler City, North Carolina. This stock intrudes predominantly felsic volcanic, volcaniclastic, and epiclastic rocks of probable late Precambrian to middle Cambrian age (Hills and Butler, 1969; St. Jean, 1973; Wright and Seiders, 1980). The pluton's boundary was mapped and samples were collected to determine its age and whole rock chemistry. Previous work in the immediate Parks Crossroads area is restricted to reconnaissance mapping by Conley and Bain (1965) and Bain (1966). More detailed data have been collected in the Asheboro, Denton, and Albemarle areas to the west (Conley, 1962; Butler and Ragland, 1969; Stromquist and Sundelius, 1969; Stromquist, Choquette, and Sundelius, 1971; Seiders, 1978; Wright and Seiders, 1980) as well as the Moore County area to the southeast (Conley, 1962; Cavaroc and others, 1979).

GEOLOGY

Country Rocks

The country rocks are felsic tuff and lava flows, volcanic conglomerates, sandstones, siltstones, and argillites. The rocks have been regionally metamorphosed to the chlorite-zone greenschist facies and they bear two foliations. The major foliation strikes northeast and dips steeply to the northwest; the minor foliation strikes east-west and is nearly vertical. Bedding in the clastic units strikes northeast and dips moderately to steeply southeast (Figure 1). Sparce sedimentary structures indicate that the units are stratigraphically upright. The stratigraphy is grossly oversimplified on the accompanying geologic map (Figure 1) as most of the beds appear to be lenses or tongues that are complexly interlayered and few marker beds exist; the lack of outcrop is also a factor.

The lowest unit mapped is a porphyry lava flow (plf, Figure 1) which is characterized by phenocrysts of euhedral to subhedral saussuritized plagioclase laths and rounded quartz grains, about 1-2 mm across, set in a quartzofeldspathic groundmass with abundant disseminated pyrite. The next unit (vvc, Figure 1) is a grouping of various silicic lava and tuff flows with intercalated fine-grained epiclastic rocks. Just

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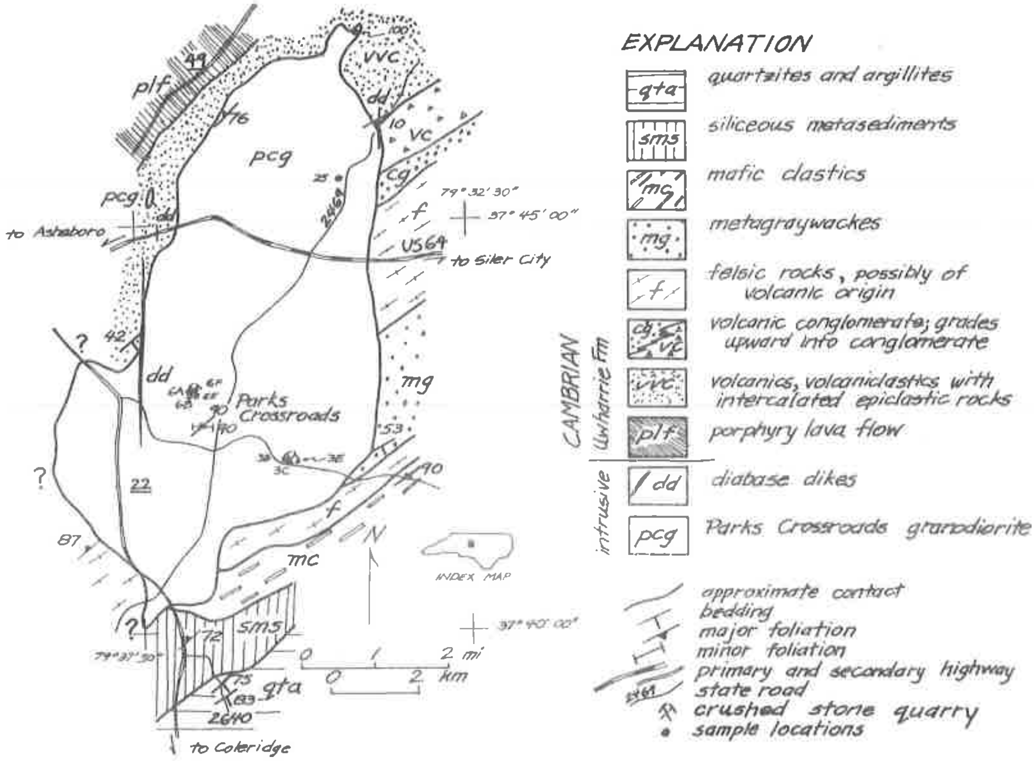

Figure 1. Geologic map of the Parks Crossroads area with sample locations.

above this sequence lies a conglomerate (vc, Figure 1) which may be the Denny Conglomerate Member of the Efland Formation reported by Conley and Bain (1965). It contains sub-rounded to sub-angular pebbles of potassium feldspar, plagioclase, quartz, and mafic volcanics in a feldspathic matrix. The presence of iron oxide gives the weathered rock an orange-red to yellow color. This rock grades upward into a conglomeratic sandstone. The felsic clastic unit (f, Figure 1) above the conglomerate weathers to a distinctive orange-red clay-rich soil. The rock is well foliated and bedding is rarely seen. Siltstones and graywackes crop out locally in this unit. A thick sequence of generally greenish to grayish texturally immature clastic rocks, herein termed metagraywackes (mg, Figure 1) lies above the felsic clastic unit. Clasts ranging from small pebbles to sand are sub-angular and are of probable volcanic origin (note: the term "graywacke" has no genetic significance here). The mafic clastic unit (mc, Figure 1) above the graywacke is also texturally immature though generally better sorted and its soil is more brown than others in the area. The overlying siliceous metasediments (sms, Figure 1) are poorly exposed due to a thick cover of vegetation. Where seen in streamcuts, the rocks have contorted and convoluted bedding not unlike sedimentary features in turbidites. The uppermost unit in the area is a sequence of bedded argillites (qta, Figure 1), some of which are quartz-rich. They are well exposed along Highway 22 south of Coleridge, North Carolina.

Unmetamorphosed diabase (dd, Figure 1), probably Triassic-Jurassic in age, crops out in north-south and northwest-trending dikes. The mineralogy varies from olivine 35 to 0% with andesine, pigeonite or sub-calcic augite, and magnetite.

Pluton

The Parks Crossroads pluton (pcg, Figure 1) is a very homogeneous medium-grained equigranular hornblende granodiorite. Its contact with the country rocks is sharp and discordant and it too has been metamorphosed. The plagioclase is commonly saussuritized and chlorite has grown in turbid aggregates with epidote and actinolite to

Table 1. Major element analysis for six samples of the Parks Crossroads granodiorite. SiO_2 was determined by colorimetry. CaO, TiO_2, and MnO were determined by wavelength x-ray fluorescence spectrometry Al_2O_3, Fe_2O_3, MgO, K_2O, and Na_2O were determined by atomic absorption spectrophotometry. Tracy N. Tingle performed the analyses at the University of North Carolina, Chapel Hill.

	SC-3B	SC-3C	SC-3E	SC-6A	SC-6B	SC-6E
SiO_2	69.22	67.70	65.69	70.59	71.71	71.31
TiO_2	0.42	0.41	0.32	0.32	0.34	0.33
Al_2O_3	15.25	15.55	17.25	15.11	14.18	15.13
Fe_2O_3 T	4.26	4.23	5.10	2.68	2.90	2.82
MnO	0.06	0.08	0.06	0.08	0.08	0.08
MgO	1.29	1.24	1.85	0.85	0.91	0.89
CaO	3.14	3.46	2.93	2.52	2.53	2.52
Na_2O	4.08	3.80	3.80	3.72	3.61	3.69
K_2O	2.38	2.62	2.18	3.10	2.92	2.96
Total	100.10	99.09	99.18	98.97	99.18	99.73

replace hornblende. Two crushed stone quarries in the pluton afforded fresh samples for whole-rock chemistry and isotopic age determination. The whole rock chemistry is given in Table 1.

GEOCHRONOLOGY

Nine samples were collected for the purpose of Rb-Sr isotopic age determination and their locations are shown on the geologic map (Figure 1). Samples were analyzed by isotope dilution methods. The isotopic ratios were measured on a solid source 60^{0} Nier-type mass spectrometer at the University of North Carolina, Chapel Hill. The isochron was fitted by the least squares method of York (1969) assuming a decay constant for ^{87}Rb of 1.42×10^{-11} yrs^{-1} and errors are given at the one-sigma confidence level.

Four samples (3C, 3E, 6F, 25) were not used in the age calculation due to the advanced stage of plagioclase saussuritization (Figure 2a). The remaining five samples (3B, 6A, 6B, 6E, 100) were chosen on the basis of the preservation of igneous plagioclase twinning (Figure 2b) and yield a whole rock isochron age of 566 ± 46 m.y. (York Model II) with an initial $^{87}Sr/^{86}Sr$ ratio of 0.7028 ± 0.0006 (Figure 3). The altered samples fall on a line roughly parallel to the 566 m.y. isochron suggesting systematic loss of ^{87}Rb, perhaps during metamorphism.

Avalonian ages have been reported for other plutons in the slate belt that are mineralogically very similar to the Parks Crossroads pluton (Fullagar, pers. comm., 1981; Fullagar, 1971; Glover and Sinha, 1973; Black and Fullagar, 1976). Glover and Sinha (1973) report a 575 ± 20 m.y. U-Pb concordia age for the Roxboro pluton. Fullagar (1971) reports Rb-Sr whole rock isochron ages for the Farrington complex of 519 ± 25 m.y. with an initial $^{87}Sr/^{86}Sr$ ratio of 0.7034 ± 0.0014 and a 554 ± 63 m.y. age for the Great Falls pluton with an initial ratio of 0.7041 ± 0.0029. The 566 m.y. age for the Parks Crossroads pluton is also in good agreement with model ages for the Edgmoor, Sparta, and Hatcher plutons of 545, 535, and 595 m.y. respectively, as reported by Fullagar (1971).

DISCUSSION

The Rb-Sr whole rock isochron age determined in this study for the Parks Crossroads pluton constrains the age of the country rocks to be older than 566 ± 46 m.y. Age relationships are consistent with the correlation of lithologies in the Parks Crossroads area with those in the Albemarle and Asheboro areas. The Uwharrie

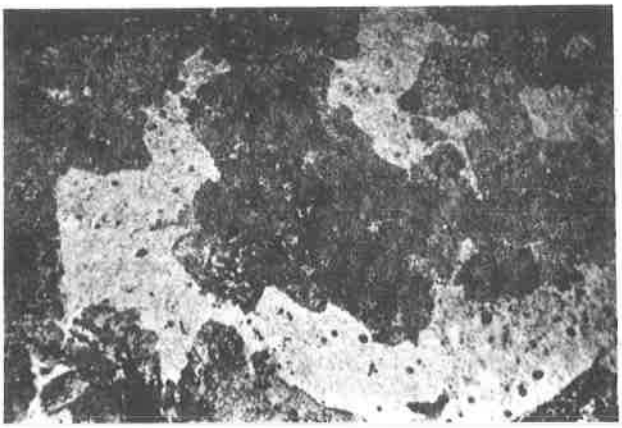

Figure 2a: SC-3E. Completely saussuritized plagioclase with cloudy aggregate of epidote and albite surrounded by quartz. (Crossed-nichols, 46X). The spots are air bubbles beneath the cover slip.

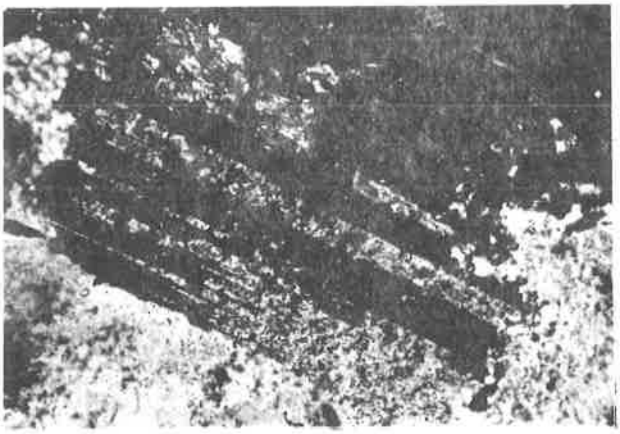

Figure 2b. SC-6E: Tabular plagioclase with primary albite twinning. Note the saussuritization on the lower corner of the grain. (Crossed-nicols, 132X)

Formation has yielded a Rb-Sr whole rock isochron age of 554 ± 50 m.y. (Hills and Butler, 1969) and a U-Pb concordia age of 586 ± 10 m.y. (Wright and Seiders, 1980). Furthermore, the porphyry lava flow (plf, Figure 1) mapped in this study is along strike with porphyry lava flows in the Asheboro quadrangle mapped by Wright and Seiders (1980) in the Uwharrie Formation. Also, epiclastic rocks along Highway 22 near Coleridge, North Carolina are similar in lithology to rocks described by Conley and Bain (1965) in the McManus Formation. The interpretation made here is that the country rocks in the Parks Crossroads area are correlative with Uwharrie Formation and Albemarle Group rocks. Field data to fully substantiate these correlations are lacking, however.

Limited bedding attitudes in the Parks Crossroads area increase from 10^{o} southeast along state road 2469 to near vertical along state road 2640 (Figure 1). These data imply that this sequence of rocks lies on the flank of large structure to the northwest of the area. If lithologies in the Parks Crossroads area are correlative with Uwharrie Formation and Albemarle Group rocks, then it seems likely that the probable structure to the northwest of the study area is an extension of the Troy anticlinorium.

CONCLUSIONS

A thick and apparently conformable sequence of volcanic, volcaniclastic, and largely volcanic-derived epiclastic rocks in the Parks Crossroads area has been

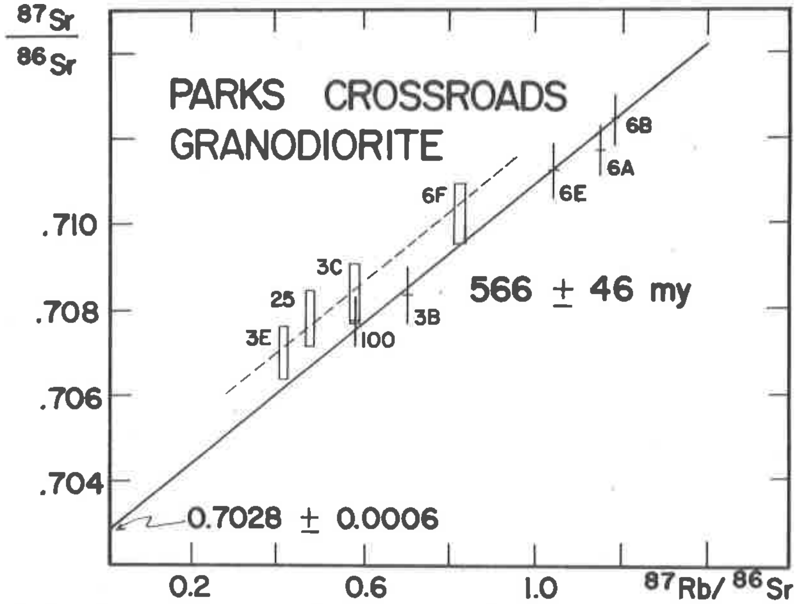

Figure 3. Rb-Sr whole rock isochron for the Parks Crossroads granodiorite. The dashed line is drawn through the four samples not used in the age calculation due to the advanced saussuritization of the plagioclases. Sample SC-3E shown in Figure 2a is typical of these four samples.

correlated with similar lithologies reported in the Asheboro and Albemarle areas that range in age from late Precambrian to middle Cambrian (Uwharrie Formation and possibly Albemarle Group rocks). Structural data suggest that the sequence rests on the southeastern flank of the Troy anticlinorium. The 566 ± 46 m.y. Rb-Sr whole rock isochron age for the Parks Crossroads pluton, which intrudes that sequence, is consistent with the 586 ± 10 m.y. U-Pb concordia age for the upper Uwharrie Formation obtained by Wright and Seiders (1980), and is also consistent with Avalonian ages reported for other plutons in the Carolina slate belt (Fullagar, 1971).

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