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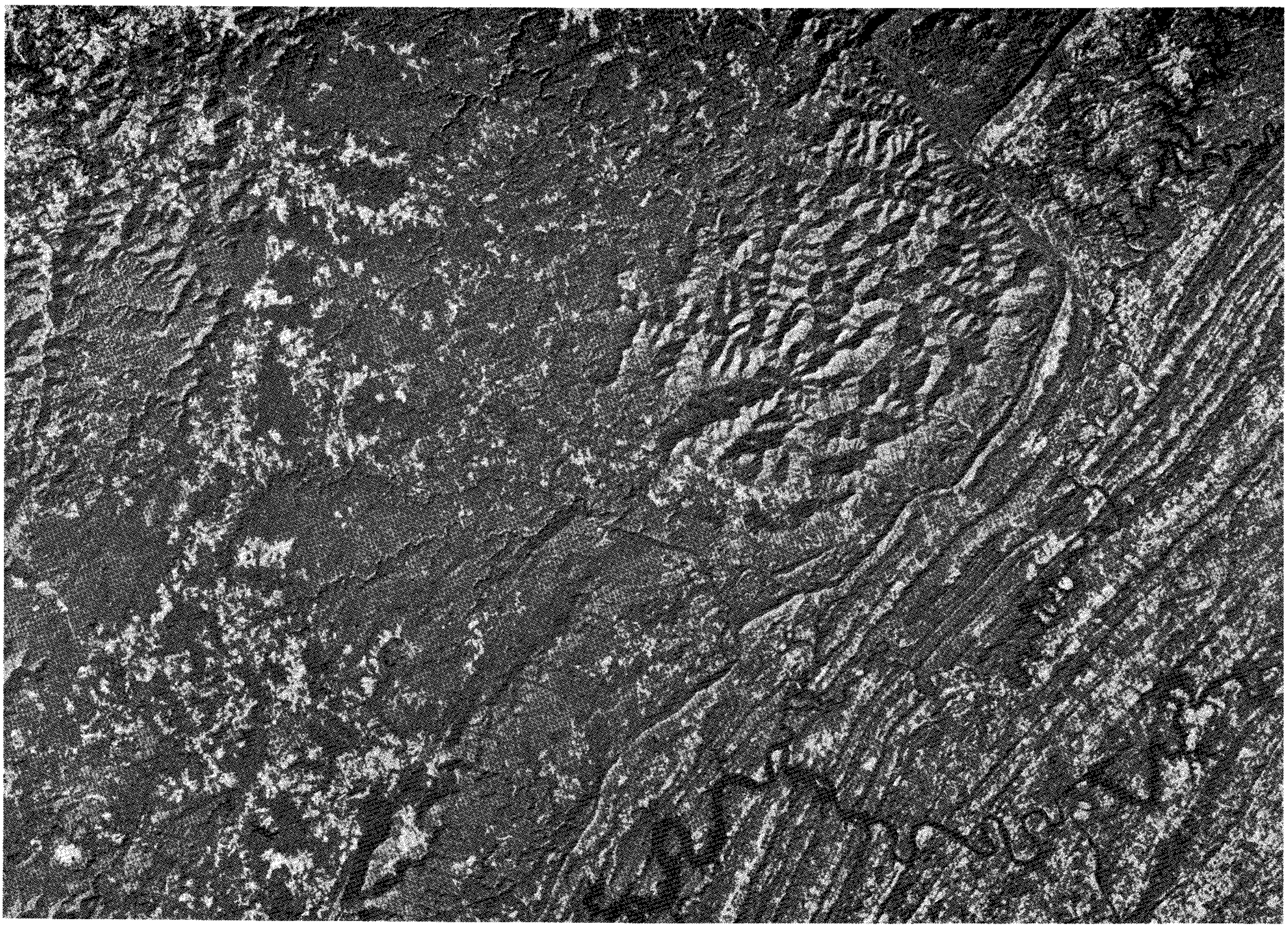
CONTRIBUTIONS TO VIRGINIA GEOLOGY — V



COMMONWEALTH OF VIRGINIA

DEPARTMENT OF MINES, MINERALS AND ENERGY
DIVISION OF MINERAL RESOURCES
Robert C. Milici, Commissioner of Mineral Resources and State Geologist

CHARLOTTESVILLE, VIRGINIA
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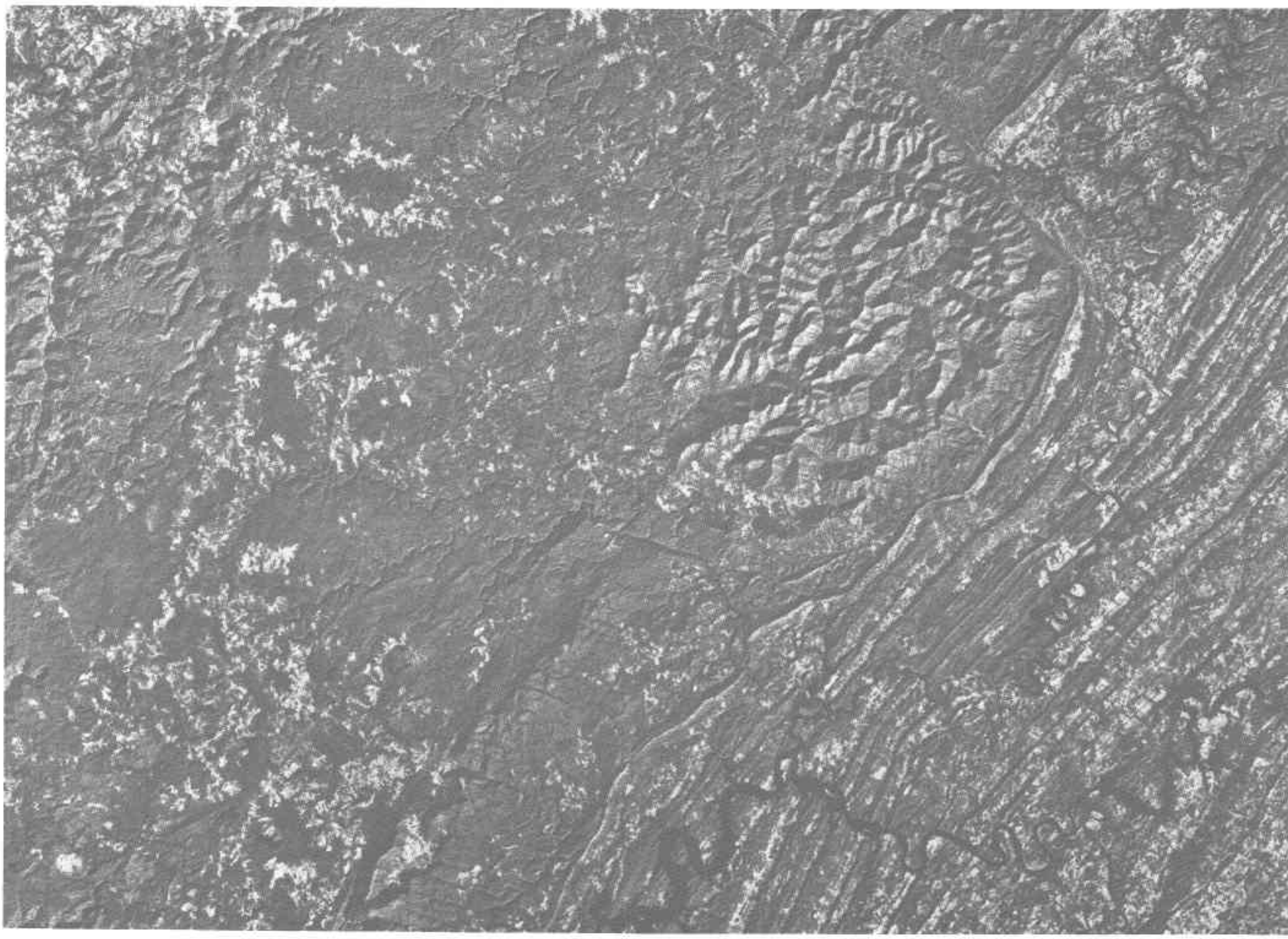

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COVER PHOTO: LANDSAT image (Band 7) showing the western part of the Appalachian Valley and Ridge and Cumberland Plateau of Tennessee. Leonard Harris, to whom this contributions volume is dedicated, did much of his field work on the Pine Mountain block and adjacent parts of the Valley and Ridge province.

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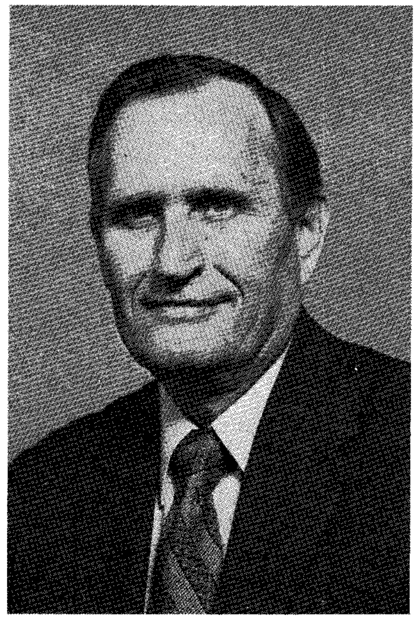
DEPARTMENT OF MINES,
MINERALS AND ENERGY

RICHMOND, VIRGINIA

O. GENE DISHNER, Director

Leonard D. Harris
1925-1982



Leonard D. Harris was born in Kansas City, Missouri, on March 10, 1925, and died in the Washington, D.C. area on July 27, 1982. He served with the U.S. Marine Corps in the South Pacific during World War II, participating in amphibious operations against the Japanese. After the war he obtained a bachelor's degree in geology from the University of Missouri in 1949 and in 1950 joined the U.S. Geological Survey.

Leonard was first and foremost an able field geologist. He began his mapping for the U.S. Geological Survey with a short tour in the Ouachita Mountains. In 1952 he moved to the Appalachians to work in the coal fields, where he mapped and compiled coal resource information in Kentucky and studied deep oil and gas tests in West Virginia. He began his studies of the Valley and Ridge shortly thereafter and, with Ralph Miller, published his first geologic quadrangle report "The Geology of the Duffield Quadrangle, Virginia" in 1958. In 1958 Len started a regional project in the southern Appalachians which was designed to test the thin-skinned hypothesis by mapping a block of quadrangles across the Valley and Ridge. This project, which resulted in detailed field mapping of eight 7½-minute quadrangles and in several regional stratigraphic and structural geology studies in the imbricate thrust belt of eastern Tennessee and southwestern Virginia, provided him with the experience essential for his subsequent regional studies and syntheses of Appalachian fold and thrust belt tectonics.

Leonard had many and diverse career interests and a curiosity that was constantly stimulated by his work in the field. He often said that he approached each study using "imagination and geo-logic." He wrote papers on stratigraphy, facies, and depositional environments; on paleotopography and paleoaquifers; and on Appalachian structure and tectonics. Len was interested in the Knox Group and its mineral resource potential, both as a reservoir for oil and gas and for its Mississippi Vally-type ore deposits. He revised

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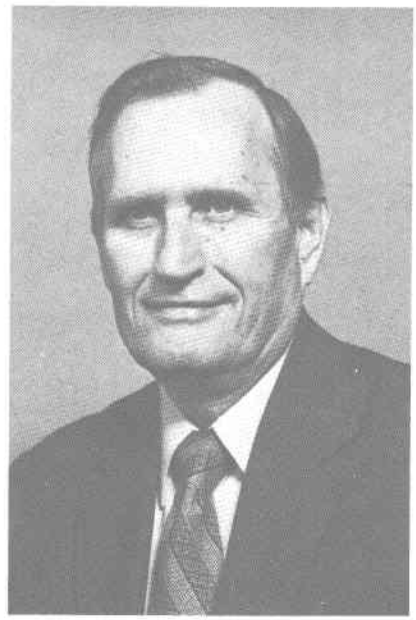

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As part of his work for the Branch of Eastern Environmental Geology, Len made an exhaustive environmental study of Knox County, Tennessee. In one map series he (and several associates) compiled and published data on the geology, mineral resources, land slopes, and urbanization, stratigraphy, structure, sinkholes, soils, engineering geology, and flood potential in Knox County.

During the later part of his career, after the oil embargo of 1973 had focused national attention on the need for domestic petroleum resources, Len returned to his interests in regional studies related to Appalachian hydrocarbon potential. Philosophically, he was a converted thick-skinner, and although little data were available to the public to prove thin-skinned deformation of the Appalachians even in the early 1970s, Len had accepted and endorsed the concept. He obtained a short Vibroseis profile in eastern Tennessee from Geophysical Services, Inc. and with their permission in 1976 published his interpretation, which demonstrated the nature of thin-skinned tectonics in the Tennessee imbricate thrust belt. When the Tennessee Division of Geology, under contract to the U.S. Department of Energy, was conducting regional Vibroseis surveys of the Valley and Ridge in 1977, Len convinced the U.S. Geological Survey to contribute to the project while it was in mid-shoot. This Survey contribution allowed one of the lines to be extended entirely across the Valley and Ridge to the toe of the Blue Ridge, where it became evident that about 23,000 ft of Paleozoic strata were projecting eastward beneath crystalline overthrusts. The success of the Tennessee surveys encouraged Len to pursue this line of research, and the resulting regional seismic profiles by the U.S. Geological Survey in Tennessee and North Carolina, and later in Virginia, are fundamental to our understanding of Appalachian thrust systems. Leonard provided the inspiration, justification, and leadership in the U.S. Geologcal Survey Branch of Oil and Gas Resources that was neccessary to secure appropriate funding for the contract geophysical work and to interpret data once they were obtained.

These studies, and others concerned with regional structural and stratigraphic syntheses and with thermal maturation of Paleozoic strata in the Appalachians, provided basic geologic data essential for exploration in the region. Leonard worked with and instructed many company geologists interested in exploring the Appalachians for oil and gas. He contributed greatly to their understanding of this complex geology. Len was an excellent speaker, conducting numerous seminars, both formal and informal, for company geologists. He was at his best, however, when leading fields trips and discussing thin-skinned tectonics with those most important exposures providing the background for understanding Appalachian structural geology.

Leonard was a hard-working and dedicated individual. He gave much more to his associates than he expected in return. His natural ability, intelligence, hard work, dedication to the profession, and unabated curiosity concerning Appalachian geology constantly stimulated and inspired those of us associated with him to vigorously pursue our own, often related studies. When he began working in the Appalachians, the thick-skin—thin-skin debate was unresolved. When he finished, regional geologic and geophysical surveys funded by industry and government had provided a great deal of detailed information concerning the framework of the Appalachian basin and the geometry of Appalachian thrust systems, and overthrusting of Valley and Ridge Paleozoic rocks by Blue Ridge and Piedmont cystalline rocks had been clearly demonstrated by extensive seismic profiling—he had played a leading role in the process.

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CONTENTS

PRE-TACONIAN DEFORMATION IN THE PIEDMONT OF THE POTOMAC VALLEY—PENOBSCOTIAN, CADOMIAN, OR BOTH?

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CONTENTS

ILLUSTRATIONS

ABSTRACT

The crystalline rocks in the Piedmont of the Potomac Valley occur in three tectonic motifs of allochthon and underlying precursory melange that are overlain unconformably by the Popes Head Formation of probable Cambrian age. The highest tectonic motif consists of the Piney Branch Complex (an ophiolite fragment) and the Yorkshire Formation; the middle motif consists of the Peters Creek Schist (a submarine fan that contains ophiolitic debris) and the Sykesville Formation; and the lowest motif consists of the Annandale Group (a submarine fan that does not contain ophiolitic debris) and the Indian Run Formation. All these rocks and the Popes Head

Formation are older than the Occoquan Granite, which has been dated at 560 Ma by Pb-U techniques on zircon and at 494±14 Ma by whole rock Rb-Sr techniques and is herein thought to be probably of Late Cambrian age.

Structural analysis of the Popes Head Formation, Annandale Group, and Peters Creek Schist shows that the Popes Head contains three fold phases and that the other units have had more complex structural histories, that of the Peters Creek Schist being the most complex. The Occoquan Granite appears to have been emplaced synkinematically during the oldest fold phase in the Popes Head Formation. If the Occoquan is of Late Cambrian age, the deformation in the Popes Head dates from the Penobscot orogeny. Older structures in the Annandale Group and Peters Creek Schist are overprinted by all deformations experienced by the Popes Head and are thought to date from the Cadomian orogeny.

Cadomian and Penobscot deformations have not been recognized in rocks that can be certified as North American, so this part of the Piedmont should be considered a suspect terrane or terranes. The different lithotectonic environments represented by the Piney Branch Complex, Peters Creek Schist, and Annandale Group suggests that they are mashed-together terrane fragments. After their assemblage, Popes Head sedimentation, and Penobscotian deformation, they formed the composite Potomac terrane. On the west, in southern Maryland, the Potomac terrane is in contact with the Ijamsville Phyllite. This contact is marked by a zone of phyllonitized Peters Creek and may be the Taconian suture in this part of the Appalachians.

INTRODUCTION

A current major thrust in Appalachian geology is an attempt to correlate the geologic history of this orogen with that of the Caledonian, Hercynian, and Mauritanide orogens of the British Isles, Europe, and North Africa through the International Geologic Correlation Program's (I.G.C.P.) Project 27. A major topic of this program is time-of-deformation.

Times of deformation are difficult to determine in the central Appalachian Piedmont because, with a few exceptions, the rocks are unfossiliferous. The exceptions are some of the carbonate rocks in southeastern Pennsylvania, which are of Cambrian and Early Ordovician (?) age (Freedman and others, 1964; Higgins, 1973), and the Quantico Formation of Late Ordovician age (Pavlides and others, 1980). This situation is further complicated by the total lack of Silurian and younger cover rocks. There has been a tendency to ascribe various phases of deformation to the orogenic periods recognized in the foreland and successor basins to the west, that is, the Taconian, Acadian, and Alleghanian (see for example, Amenta, 1974; Freedman and others, 1964; Higgins, 1973). Crowley (1976) inferred that deformation in the Baltimore area resulted from the Taconic orogeny, whereas Fisher (1970) cautiously ascribed the deformation along the Potomac River to a single, complex, but essentially unified orogeny, which took place mainly in Paleozoic time. Hopson (1964), taking a longer view, perhaps because he believed that many of the rocks in the Maryland Piedmont were of Late Proterozoic age, recognized that the deformation there was as least as old as syntectonically emplaced quartz diorite plutons dated at about 500 Ma and the quartz monzonite plutons dated at about 370 Ma were either latest syntectonic or post-tectonic. Pavlides (1973) recognized evidence for a deformation in the Chopawamsic Formation of Early Cambrian(?) age that is absent in the overlying Quantico Formation of Late Ordovician age in the central Virginia Piedmont. Pavlides and others (1982) reported recumbent folding and thrusting in the Fredericksburg, Virginia area that postdates the emplacement of plutons dated at 410 Ma, suggesting major Acadian deformation. In the same general area, other plutons dated at 300-325 Ma are strongly to weakly foliated, suggesting Alleghanian deformation (Pavlides and others, 1982). In the Piedmont of the Potomac Valley, all the metamorphic rocks are older than the Occoquan Granite (Drake and Lyttle, 1981), which has been dated at 560 Ma by U-Pb techniques on zircons (Seiders and others, 1975) and at 494 ± 14 Ma by Rb-Sr wholerock techniques (Mose and Nagel, 1982). Much of the deformation in the Potomac Valley was recognized as predating the emplacement of the Occoquan Granite (Drake and Lyttle, 1981). Deformation in the Potomac Valley then was described as being Taconian or older. It is clear now, however, that the Occoquan Granite is of Cambrian age and that the structural features that predate it are of Cambrian or older age, perhaps both. The purpose of this paper is to describe some of the structural features in the metamorphic rocks of the Potomac Valley, to attempt to place them in their proper orogenic context, and to evaluate the tectonic implications of these observations. Some of these ideas were presented at a Penrose Conference on the timing of orogenic activity in the Appalachian-Caledonian orogenic system in 1981, and the thesis of the paper was presented at a symposium, "Tectonics and Stratigraphy of Virginia," held in honor of the late Leonard Harris in 1983 at the annual meeting of the Virginia Academy of Science (Drake, 1983). The data upon which this paper is based were gathered during my mapping of the crystalline rocks of Fairfax County, Virginia, at the scale of 1:24,000, which led to the preparation of a preliminary geologic map of that county at the scale of 1:48,000 (Drake and others, 1979). I am indebted to Philip Osberg for prodding me to come to grips with the reality of the evidence for pre-Taconian deformation in the Potomac Valley.

REGIONAL GEOLOGY

The metasedimentary, metavolcanic, and transported intrusive rocks in northern Virginia occur in a stack of four lithotectonic units (Figure 1). The youngest and highest of these, the Popes Head Formation, unconformably overlies all the other units that constitute three thrust sheet-precursory tectonic melange pairs that have been termed "tectonic motifs" (Drake, in press). The highest of these motifs consists of the Piney Branch Complex and Yorkshire Formation (Drake and Morgan, 1981, 1983), the middle motif consists of the Peters Creek Schist and Sykesville Formation (Drake and Lyttle, 1981; Drake and Morgan, 1981, 1983), and the lowest motif consists of the Annandale Group and Indian Run Formation (Drake, 1983).

Popes Head Formation

The Popes Head Formation (Drake and Lyttle, 1981) unconformably overlies all the other lithotectonic units and consists of a lower Old Mill Branch Metasiltstone Member and an upper Station Hills Phyllite Member. Both members contain interbedded felsic and mafic metatuff. These rocks have been interpreted as belonging to Bouma's turbidite sequence T_{de} and to Mutti and Ricci Lucchi's (1978) turbidite facies D, suggesting deposition from weak turbidity flows (Drake and Lyttle, 1981). Drake and Lyttle (1981) have suggested that the formation results from the bilateral filling of a back-arc basin. The typical mineral assemblage of rocks of the Popes Head Formation is quartz-muscovite-biotite-plagioclase-chlorite, and it is at biotite grade of metamorphism. The unit has a maximum thickness of more than 1,000 m.

Piney Branch-Yorkshire Motif

The Piney Branch Complex, which is believed to be a fragment of a large dismembered ophiolite (Drake and Morgan, 1981), is an intimate mixture of highly metamorphosed ultramafic and mafic rocks, which are now serpentinite, soapstone, actinolite schist, and metagabbro, all intruded by small dikes and sheets of plagiogranite. The complex lacks discernible order and is thought to be a tectonic melange resulting from the autoclastic deformation of a layered complex that contained repetitive cycles of ultramafic and mafic layers (Drake and Morgan, 1981, 1983). The rocks within the complex have been metamorphosed to mineral assemblages stable in the greenschist facies of regional metamorphism, and almost no relict original minerals survive (Drake and Morgan, 1981).

The Yorkshire Formation forms a thin discontinuous lower border to the Piney Branch Complex. This formation is a sedimentary melange that has a quartz-plagioclase-chlorite matrix and contains small to large fragments of the overlying Piney Branch Complex as well as quartz and foreign rock types. The unit has been interpreted as a precursory melange to the Piney Branch (Drake and Morgan, 1981, 1983).

The Yorkshire, like the Piney Branch, contains mineral assemblages stable in the greenschist facies of regional metamorphism. The Yorkshire and the overlying Piney Branch Complex constitute the highest tectonic motif in northernmost Virginia, and the two units together compose the Piney Branch allochthon. This allochthon is thought to have had a multiple movement history which involved the subaqueous emplacement of the Piney Branch upon the Yorkshire and the later thrusting of this motif onto the Peters Creek Schist (Drake and Morgan, 1981). The discontinuous nature of the Yorkshire is thought to reflect partial shearing out during the later thrust event.

Peters Creek-Sykesville Motif

The Peters Creek Schist (Drake and Morgan, 1981) consists of quartz-rich schist and graywacke. Its general characteristics suggest that

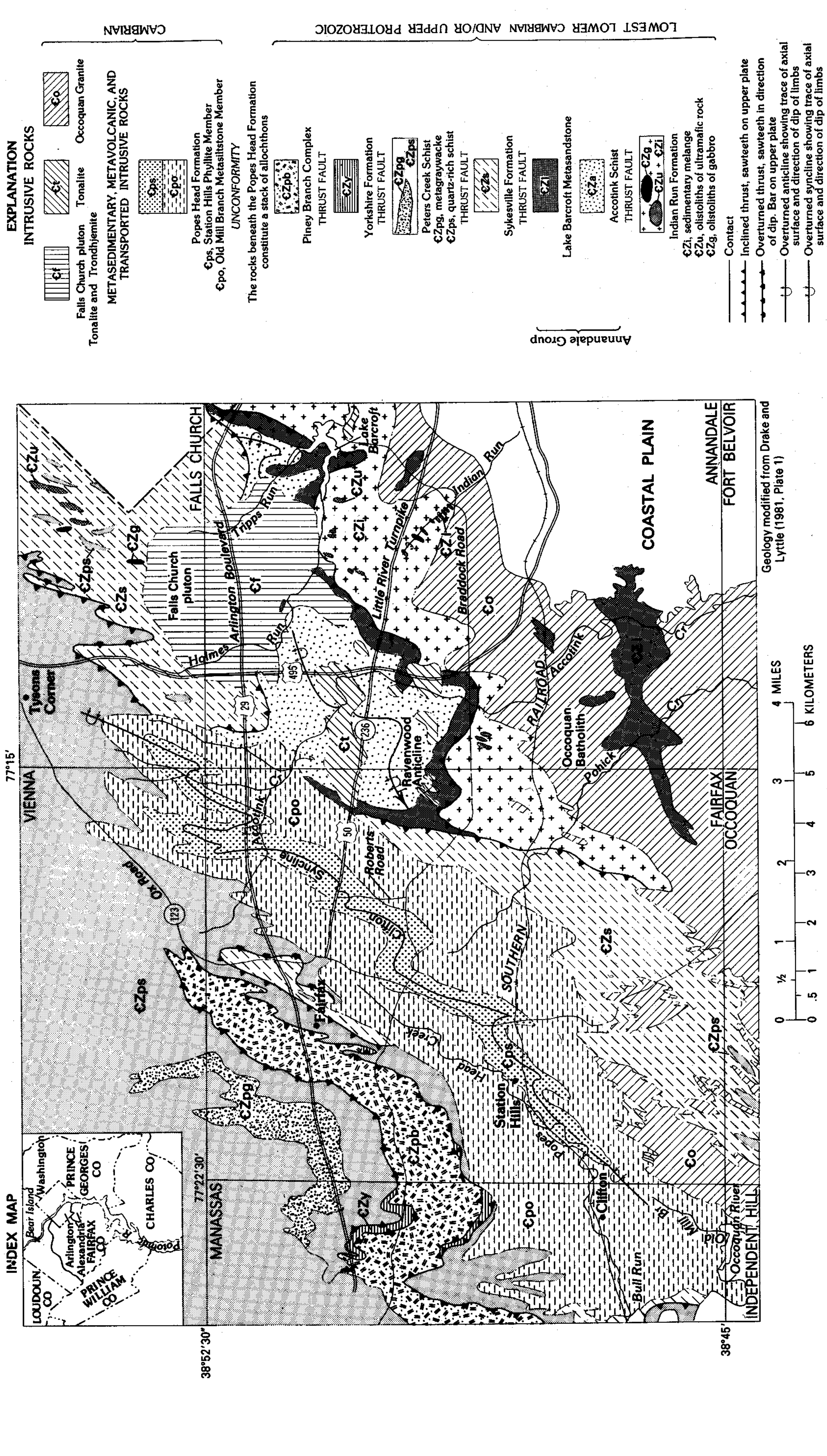


Figure 1. Generalized geologic map of part of Fairfax County, Virginia.

it is a turbidite deposited under fairly high energy conditions in a large submarine fan (Drake and Morgan, 1981). In northern Virginia and Maryland, the Peters Creek characteristically contains large olistoliths of ultramafic rocks (Drake and others, 1979; Froelich, 1975) and is considered to be a type of ophiolitic melange (Drake and Morgan, 1981, 1983). The presence of the Ultramafic olistoliths suggests that the Peters Creek may have been deposited on or adjacent to oceanic crust.

The Peters Creek Schist is polymetamorphic. It has undergone a major prograde event and ranges from chlorite grade in the west to sillimanite grade in the east (Fisher, 1963, 1970; Drake and others, 1979; Drake and Morgan, 1981; Drake, 1985) the highest grade rocks and migmatites become severely sheared and are retrogressively metamorphosed to chlorite-sericite phyllonites. The last metamorphic event recognized in these rocks was the growth of chloritoid within shimmer aggregate pseudomorphs of staurolite, andalusite, and garnet (Drake, 1980), and, in places, new muscovite and biotite have formed.

The Sykesville Formation (Drake and Morgan, 1981) is a complicated sedimentary melange that has a light- to medium-gray, medium-grained, quartz-plagioclase-muscovite-biotite-chlorite matrix that locally contains sparse amounts of garnet. Set in this matrix are characteristic quartz lumps as well as fragments of serpentinite, metagabbro, felsic and mafic metavolcanic rocks, Peters Creek Schist, and other exotic rock types. The olistoliths of ultramafic and mafic rocks were also deformed before being introduced into the Sykesville (Drake and Morgan, 1981, 1983), and the olistoliths of Peters Creek are invariably foliated and polydeformed and in many places are migmatite. Directly beneath the phyllonitized Peters Creek, the Sykesville is choked with fragments of Peters Creek phyllonite (Drake and Morgan, 1981, 1983).

The Peters Creek Schist was lithified, metamorphosed, deformed, migmatized, and phyllonitized prior to Sykesville deposition, and it now overlies its own debris within the Sykesville. The Sykesville then is a classic precursory melange (Elter and Trevisan, 1973), and together with the overlying Peters Creek Schist, it constitutes the structurally middle tectonic motif in northernmost Virginia.

The mineral assemblage of the Sykesville suggests that it is at biotite ± garnet grade and there is no evidence of polymetamorphism.

Annandale-Indian Run Motif

The Annandale Group (Drake, 1985) consists of the lower Accotink Schist and the upper Lake Barcroft Metasandstone. The group was interpreted by Drake and Lyttle (1981) to be a coarsening-upward sequence of an outer submarine-fan association of rocks. So far as is known, it contains no fragments of ophiolitic rocks.

The Accotink Schist (Drake and Lyttle, 1981) is a quartz-muscovite-biotite-chlorite-plagioclase (garnet-magnetite-epidote) schist that is characterized by the Bouma (1962) turbidite sequence T_e and T_{de} and was assigned to turbidite facies D and E of Mutti and Ricci Lucchi (1978) by Drake and Lyttle (1981).

The Accotink is polymetamorphic. The first metamorphism resulted in the growth of muscovite and biotite in the first-generation schistosity, and the second metamorphism resulted in the growth of new mica in a second generation schistosity as well as porphyroblasts of garnet and chlorite (Drake and Lyttle, 1981). The unit is at biotite ± garnet grade. Its thickness is not known because its base is not exposed.

The Lake Barcroft Metasandstone (Drake and Lyttle, 1981) consists of thick-bedded quartzofeldspathic granofels without interbedded schist, and thin- to medium-bedded micaceous metagraywacke containing schist layers. The quartzofeldspathic granofels typically is a meta-arenite, interpreted by Drake and Lyttle (1981) as being a sequence of T_a turbidites and as belonging to turbidite facies B_2 of Walker and Mutti (1973). Micaceous metagraywacke has been interpreted by Drake and Lyttle (1981) as belonging to turbidite facies C of Mutti and Ricci Luchi (1978) and Walker and Mutti (1973).

The Lake Barcroft has had the same metamorphic history as the Accotink Schist and is at biotite ± garnet grade. As much as 400 m of the unit is present in the Annandale 7.5-minute quadrangle, where it is best exposed.

The Indian Run Formation (Drake, 1985), like the Sykesville Formation, is a complicated sedimentary melange that has a quartz-plagioclase-muscovite-biotite-chlorite matrix; it commonly contains abundant garnet. The Indian Run appears to contain more plagioclase and phyllosilicate and less quartz than does the Sykesville. Like the Sykesville it contains quartz lumps, but they are generally fewer and smaller. In addition, the matrix contains fragments of foliated felsic and mafic metavolcanic rocks, metagabbro, and ultramafic rocks. The ultramafic fragments, either serpentinite or soapstone, range in size from microscopic to mesoscopic to mappable olistoliths as long as 350 m. In addition to blocks of quartz and mafic and ultramafic rocks, the Indian Run contains fragments of Accotink Schist and Lake Barcroft Metasandstone. It differs from the Sykesville in that it does not contain Peters Creek Schist olistoliths.

The Indian Run is overlain by the allochthonous Annandale Group and together with those rocks, it constitutes the lowest exposed tectonic motif in northernmost Virginia. The Indian Run apparently has undergone only one phase of prograde metamorphism and is at biotite ± garnet grade.

Intrusive Rocks

The metamorphic rocks of the area have been intruded by the Occoquan Granite, the Falls Church pluton (Steidtmann, 1945) of tonalite and trondhjemite, and smaller unnamed plutons of tonalite (Drake and others, 1979; Drake and Lyttle, 1981; Drake, 1983). The Occoquan Granite forms the small Occoquan batholith and several satellitic plutons and consists mostly of monzogranite but has lesser granodiorite and tonalite phases (Seiders and others, 1975). Only the northern part of the Occoquan batholith crops out in the area of this study, and the border rocks here appear to be largely tonalite. Seiders and others (1975) obtained both concordant and discordant U-Pb ages of about 560 Ma from zircon from this granite that suggest an Early Cambrian age. Higgins and others (1977) questioned the zircon ages and suggested that the zircons may have inherited a component of older radiogenic lead in seed crystals derived from Proterozoic basement rocks. Seiders (1978) pointed out that neither analytical nor textural evidence exists for xenocrystic zircons in the Occoquan Granite and that there is no direct evidence that Occoquan magma ever passed through zircon-bearing Proterozoic basement rocks. More recently, Mose and Nagel (1982) determined a Rb-Sr whole-rock isochron age of the Occoquan of 494 ± 14 Ma. In my experience, Rb-Sr ages tend to be on the young side. In any case, the Occoquan almost certainly is of Cambrian age.

The small tonalite plutons seem similar to the tonalite of the Occoquan batholith and may be cupolas of an as-yet-unroofed composite intrusion, as suggested by Drake and others (1979). The Falls Church pluton is very poorly exposed, but most of the rocks therein are identical with the tonalite plutons and the tonalite of the Occoquan batholith. The Falls Church Pluton does, however, contain some trondhjemite. Unfortunately, the relation of the trondhjemite to the tonalite cannot be determined.

Two other types of intrusive rock, occupying areas too small to show at the scale of Figure 1, are also present. One rock type is light-gray to pink, medium-grained muscovite monzogranite containing lesser amounts of pegmatite that intrudes the Accotink Schist and Indian Run Formation and appears to predate the Occoquan Granite (Drake and others, 1979). The monzogranite has not been isotopically dated.

The other intrusive rock, the Bear Island Granite (Cloos and Cook, 1953), forms dikes and small crosscutting intrusive bodies. A very leucocratic rock, it ranges in composition from monzogranite to albite granite and has been called quartz monzonite, adamellite, granodiorite, leucotonalite, and quartz diorite by various geologists (see Hopson, 1964, for a summary). I choose to follow Hopson (1964) and call the unit Bear Island Granite because of its silicic nature, because the plagioclase is albite, and because some specimens contain as much as 18 percent microcline. Many bodies of Bear Island Granite locally coarsen into pegmatite, and muscovite from two such pegmatite bodies has been dated by Rb-Sr techniques at 469 ± 20 Ma (Muth and others, 1979). The Bear Island Granite appears to intrude only the Peters Creek Schist.

AGES OF METAMORPHIC ROCKS

The ages of metasedimentary and metavolcanic rocks cannot be directly determined and depend on the age of the Occoquan Granite, which has been shown to be of Cambrian age. The Occoquan can be directly observed to intrude the Popes Head Formation and rocks of the Annandale Group (Drake and Lyttle, 1981). Map relations clearly show that it intrudes the Sykesville and Indian Run formations. The relation of the Occoquan to the Peters Creek Schist, Piney Branch Complex, and Yorkshire Formation cannot be directly determined, but all these rocks are older than the Sykesville and Popes Head formations; therefore, they must be older than the Occoquan and thus must be of Cambrian or older age. The Popes Head Formation was originally defined as being of Cambrian or Late Proterozoic age (Drake and Lyttle, 1981). I now believe it to be of Cambrian age because of its unconformable relation to the other lithotectonic units and because the Occoquan may be younger than Early Cambrian age.

The Annandale Group and, particularly, the Peters Creek Schist have had much more complicated structural and metamorphic histories than have their respective precursory melanges—the Sykesville and the Indian Run formations—and the younger Popes Head Formation. This suggests that the Annandale and Peters Creek may be of Late Proterozoic age. The Sykesville and Indian Run are probably of Cambrian rather than of Late Proterozoic age, but that is uncertain at this time.

STRUCTURAL GEOLOGY

Stratigraphic and tectonic relationships indicate that the pre-Popes Head rocks were assembled into a stack of three tectonic motifs prior to Popes Head deposition and also show that the Popes Head was deposited before the emplacement of the Occoquan Granite of Cambrian age. To arrive at the time (times) of deformation for this area, the structural geology of the Popes Head must be analyzed and compared with that of the older rocks.

Folds in the Popes Head Formation

The Popes Head Formation lies within the Clifton syncline (Figure 1). The outcrop pattern of the syncline suggests that it has been refolded by two later fold phases. Direct observation in the field and structural geometric analyses show that three fold phases are present in rocks of the Popes Head Formation. These fold phases, following the usage of Tobisch and Fleuty (1969), are named for geographic localities where they are well exposed, or for the major fold of a particular phase.

Clifton Folds

The oldest fold phase in rocks of the Popes Head Formation, the Clifton, is named for the Clifton syncline, a major doubly-plunging structure that contains the Popes Head Formation. As defined by the extent of the Popes Head Formation, the syncline has a strike length of at least 24 km. Clifton folds have long planar limbs and most are close to tight folds, commonly having interlimb angles of less that 50°; some are isoclinal. Most limbs dip steeply northwest. Hinges are difficult to find in the field because of the tightness of the folds, the strong axial planar phyllitic schistosity, and later transposition. Because of this problem, axial surfaces at many places are determined by facing directions based on graded beds. Poles to bedding in the Popes Head Formation (Figure 2A) plot on crossed great-circle girdles defining two fold axes, B_1 and B_2, reflecting both the Clifton and a later fold event. The axis B_1 reflects Clifton folding and plunges 10°N. 14°E. as a result of sample bias from the south end of a doubly plunging structure. That most beds dip northwest reflects the overturned isoclinal folds of the Clifton phase, and the lack of moderate to gently dipping beds reflects the narrow hinges of these upright folds. The axis B_2 results from the later Fairfax fold phase that plunges 60°N. 30°W.; this fold phase will be discussed below.

Axes of small folds of bedding lineations formed by intersections of schistosity and bedding are plotted on Figure 2B for the northern part of the area and on Figure 2C for the southern part of the area. These plots both define great circle girdles, a geometry that has been known for some time to result from the refolding of earlier formed lineations (Ramsay, 1960, 1967; Turner and Weiss, 1963). Great-circle distribution of linear elements is commonly thought to result from the slip folding of earlier structures, but Bayley (1971) has demonstrated that in some cases, this distribution can result from buckle folding. In any case, lineations produced by Clifton folding have been rotated from their original, relatively gentle, north-northeast or south-southwest plunges (Figure 2B) to a variety of moderate to steep westerly plunges.

Refolding of Clifton folds is also shown by the plot of poles to Clifton schistosity, which fall on a partial great circle (Figure 2D). The axis to this great circle and to folds of Clifton schistosity is 60°, N. 15°W., similar to B_2 in the poles to bedding plot. To summarize, Clifton folds are tight, mostly overturned and isoclinal, long-limbed folds that have a strong axial-surface phyllitic schistosity. Where not rotated by later deformation, they plunge relatively gently to the north-northeast or south-southwest.

Accotink Creek Folds

The outcrop pattern of the Popes Head Formation, particularly that of the Station Hills Phyllite Member, strongly suggests that the Clifton syncline and associated structures are folded about somewhat more northerly trending axes

(Figure 1). Although difficult to find, 11 such folds were recognized in the Accotink Creek and Popes Head areas. These folds are named for Accotink Creek in the northeastern part of the Fairfax quadrangle (Figure 1). Accotink Creek folds deform both bedding and Clifton schistosity in the Popes Head Formation, and at many places these planar elements are transposed into a strain-slip cleavage that is axial surface to Accotink Creek folds. Accotink Creek folds are dextral folds overturned to the southeast. They are nearly coaxial with Clifton folds, and their axes plot within the Clifton axial girdle (Figure 2B). The critical feature for field identification of Accotink Creek folds is the recognition of folded Clifton schistosity in fold hinges. These can best be seen in exposures along Popes Head and Accotink creeks. They also are nearly coplanar with Clifton folds, and their axial surfaces plot within the Clifton schistosity girdle (Figure 2D). For these reasons, Accotink Creek folds are difficult to recognize in the field. This recognition is further complicated by the tendency of earlier formed fold axes to rotate into parallelism with later-formed axes if the intermediate strain axis of the later deformational field is inclined less than 30° from that of the early intermediate strain axis (Ghosh and Ramberg, 1968). The axial trends of the Clifton and Accotink Creek folds suggests that the intermediate strain axes of both fold phases were inclined at less than 30° to each other.

The rotation of Clifton linear elements onto a great-circle girdle by Accotink Creek folding allows a kinematic analysis of the Accotink Creek deformation by the method of Turner and Weiss (1963). This analysis (Figure 2E) suggests that the *a* kinematic axis of Accotink Creek folding was about 70°, N. 70°W. and that the *b* axis was about 5°, S. 10°W. It is interesting to note that one of the steep west-northwest-plunging maxima on the Clifton lineation girdle (Figure 2B) is exactly coextensive in space with the *a* kinematic axis (Figure 2E). This suggests that the *a* kinematic axis was the X axis of finite strain, perhaps the direction along which translation took place, and that the steep-plunging lineations were rotated into the direction of maximum strain, as in Dalradian rocks in Ireland described by Sanderson (1973). On the basis of the close clustering of structural elements for both fold-phase data presented above, both the Clifton and Accotink Creek fold phases appear to have resulted from progressive deformation within a stress continuum rather than from separate periods of deformation. Regional dynamic metamorphism progressed during deformation and produced schistosity during Clifton folding, which later was folded during Accotink Creek folding.

Fairfax Folds

A third phase of north-northwest-plunging folds deforms the bedding and schistosity in the Popes Head Formation as well as the axis of the Clifton syncline. Folds of this phase are named for two broad folds at Fairfax City shown by the map pattern of the Station Hills Phyllite in the area between Braddock Road, Ox Road, Route 50, and Roberts Road (Figure 1). The few Fairfax folds that can be recognized in outcrop are open and plunge steeply north-northwest because of the original steep northwest dips of the bedding and schistosity in the Popes Head Formation that constitute the form surfaces for the Fairfax folds. Some, but not all, of the Fairfax folds have a north-northwest-striking spaced cleavage parallel to their axial surfaces. Fairfax phase folds are reflected by the B_2 girdle in the plot of poles to bedding in the Popes Head Formation (Figure 2A), which has an axis of 60°, N. 30°W., as well as by the girdle in the plot of poles to schistosity (Figure 2B), which has an axis of 60° , N. 15°W. Fairfax folds are poorly understood at this time, but their orientation suggests that they may result from right-lateral tangential stress.

Folds in the Annandale Group

The rocks of the Annandale Group were folded at least twice before they were emplaced above the Indian Run Formation, on the basis of olistoliths of Lake Barcroft Metasandstone within the Indian Run that contain refolded isoclinal folds, the axes being oriented at about 40° to one another (Drake, 1985). Structural data are not abundant for rocks of the group, but enough were gathered to enable a partial geometric analysis. Poles to bedding (Figure 3A) plot on crossed girdles and define two axes: B_1, 48°, S. 80°W., and B_2, 40°, S. 28°W. Axes of small folds in bedding (Figure 3B) plot on a great circle, reflecting refolding. The axis B_1 plots on the girdle of small folds, but many fold axes plot in the northwest quadrant and define a maximum at 32°, N. 68°W. Small folds observed in the field to have the first schistosity parallel to their axial surfaces plunge moderately west-northwest. These folds are isoclinal, and most are recumbent to nearly recumbent

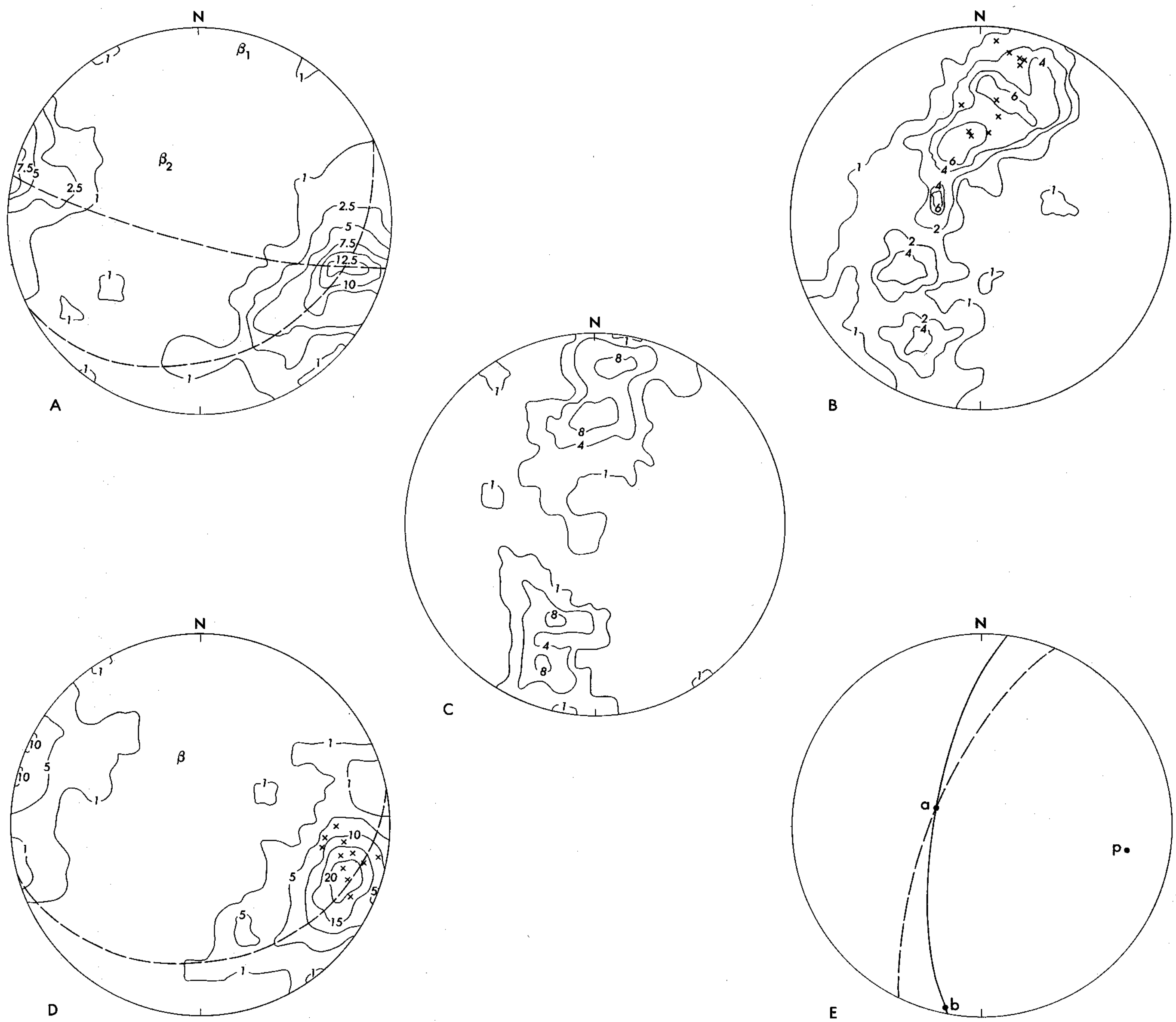


Figure 2. Equal area projections (lower hemisphere) of structural elements in rocks of the Popes Head Formation.

A. 206 poles to bedding. Dashed lines are best-fit great circles to the crossed girdles. B_1 and B_2 are the axes to the girdles. Contours are 12.5, 10, 7.5, 5, 2.5, and 1 percent per 1-percent area.

B. 120 data points that include axes of small folds in bedding and intersections of bedding and schistosity in rocks in the Fairfax, Annandale, and Vienna quadrangels. X's, axes of small folds of Clifton schistosity. Contours are 6, 4, 2, and 1 percent per 1-percent area.

C. 77 axes data points that include small folds of bedding and intersections of bedding and a schistosity in rocks in the Manassas and Independent Hill quadrangles. Contours are 8, 4, and 1 percent per 1-percent area.

D. 255 poles to schistosity. Dots, axes of small folds in schistosity, and X's, poles to axial surfaces of small folds of schistosity. B, axis to the great circle girdle. Contours are 20, 15, 10, 5, and 1 percent per 1-percent area.

E. Location of the *a* and *b* kinematic axes of Accotink Creek folds. Dashed line, lineation circle for Clifton lineation from B. P, pole to mean axial surface of Accotink Creek folds from D. Solid line, mean axial surface of Accotink Creek folds. The kinematic a axis is located at the intersection of the lineation circle and the mean axial surface and the kinematic *b* axis is located on the axial surface normal to the *a* axis.

where not refolded. They can best be seen in outcrops of the Annandale Group along and near Tripps Run (Figure 1), where exposures are somewhat better than the generally very poor average for northeastern Fairfax County. These folds are herein named Tripps Run folds. Tripps Run folds characteristically have long straight limbs and narrow peaked hinges. The axial schistosity of these folds contains the first-generation metamorphic minerals (Drake and Lyttle, 1981) and at many places transposes or nearly transposes bedding; therefore, transposition foliation is often the only planar element that can be measured.

Bedding and schistosity (or transposition foliation) have been folded together at least twice, as shown by comparison of the crossed girdle pattern on the plot of poles to schistosity (Figure 3C) to that of the plot of poles to bedding (Figure 3A). The girdles on the schistosity plot are not well defined, particularly the girdle reflected by statistical fold axis B_1, but a comparison of the statistical fold axes with measured axes of folds in schistosity (Figure 3D) lends confidence to the interpretation, and a slight shifting of B_1 to the north would fulfil the geometric requirements.

The second phase of folds in rocks of the Annandale Group is in both bedding and schistosity and is represented by the axis B_1 on Figures 3A and 3C and the west-southwest-plunging maximum shown on Figure 3D. On the scale of the Annandale quadrangle (1:24,000), these folds plunge moderately west-southwest. At the map scale (Figure 1), the Annandale Group defines the large Ravenwood anticline, which is overturned to the southeast and, though refolded, plunges west-southwest. The Ravenwood anticline is thought to be a large phase-two structure, so this phase of folds is herein called Ravenwood. Small Ravenwood folds in transposition foliation are exposed at the head of the Tripps Run arm of Lake Barcroft in the refolded hinge zone of the Ravenwood anticline. Neither here, nor elsewhere, are these folds seen to have an axial-surface schistosity.They may well have formed in a stress continuum with the Tripps Run folds, as did the Clifton and Accotink Creek folds in the Popes Head Formation. The Tripps Run and Ravenwood folds are thought to be the fold phases recognized in Annandale olistoliths in the Indian Run Formation, because they do not overprint the boundaries of the group. They, therefore, were imposed on these rocks before the Annandale Group was emplaced in its present position.

The third phase of folds recognized in the Annandale Group is represented by the statistical axis B_2 on the plot of poles to bedding (Figure 3A), the statistical axis B_2 on the plot of poles to schistosity (Figure 3C), and the south-southwest-plunging maximum on the plot of axes to small folds in schistosity (Figure 3D). The folds deform bedding as well as schistosity, although such folds are represented only poorly on the plot axes of folds in bedding (Figure 3B). This poor representation perhaps results from an inadequate representation in the data base. These folds have long planar limbs, narrow hinges, and a good axial-surface, strain-slip schistosity, represented by the northeast-striking maximum shown on Figure 3E. This schistosity contains the second-generation phyllosilicates and rotates the garnet and chlorite porphyroblasts (Drake and Lyttle, 1981). These third-phase folds overprint the boundaries of the Annandale Group, folding these rocks with those of the Indian Run and Popes Head Formations. These folds are thought to be Clifton folds because of their orientation and style, and are believed to have been the first phase of folds to form after the emplacement of the Annandale Group above the Indian Run and the sedimentation of the Popes Head Formation.

A fourth fold phase is represented by the strong maximum on the plot of axes to folds in bedding (Figure 3B). Because of its orientation, this phase is thought to represent Accotink Creek folds. This interpretation is supported by the maximum of poles to late cleavage that strikes slightly east of north, and earlier formed Clifton intersection lineations have been rotated to steep westerly plunges.

A fifth phase of folds in rocks of the Annandale Group causes the northwest-plunging maximum on the plot of axes of small folds in schistosity (Figure 3D). The strong maximum is 35°, N. 50°W., although there is a wide range of trends, and some folds plunge to the southeast. These folds are all small and are not apparently reflected in the outcrop pattern (Figure 1). They have a strong axial surface strain-slip cleavage, which forms the strong tight maximum on Figure 3E that has the attitude of N. 43°W., 75°SW. The spread of poles toward the north from this maximum represents slightly different orientations of this generation of strain-slip cleavage. Because of their orientation and style, these folds are thought to be Fairfax folds.

To summarize, the Annandale Group has been affected by five phases of folding. Two phases predate its emplacement on the Indian Run Formation; the others postdate this emplacement and

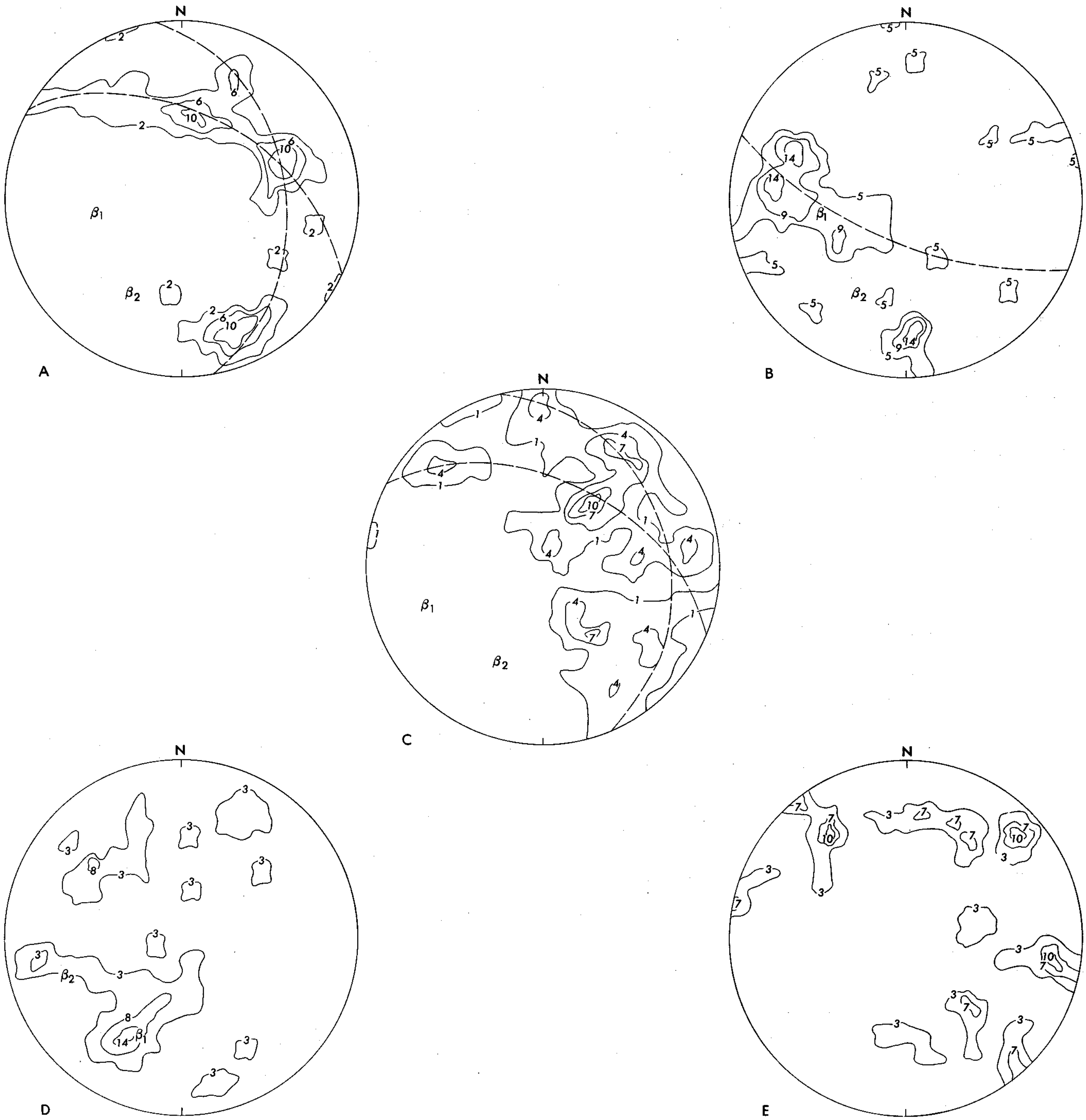


Figure 3. Equal-area projections (lower hemisphere) of structural elements in rocks of the Annandale Group.
A. 36 poles to bedding. Dashed lines, best-fit great circles to the crossed girdles. B_1 and B_2, axes to the great circles. Contours at 10, 6, and 2 percent per 1-percent area.
B. 32 axes of small folds of bedding. Dashed line, best-fit great circle to girdle. B_1 and B_2, axes shown on A. Contours at 14, 9, and 5 percent per 1-percent area.
C. 76 poles to schistosity. Dashed lines, best-fit great circles to the cross girdles. B_1 and B_2, axes to the great circles. Contours at 10, 7, and 4 percent per 1-percent area.
D. 37 axes of small folds of schistosity. B_1 and B_2, axes shown on C. Contours at 14, 8, and 3 percent per 1-percent area.
E. 30 poles to late schistosity and cleavage. Contours at 10, 7, and 3 percent per 1-percent area.

the sedimentation of the Popes Head Formation and are correlated with Clifton, Accotink Creek, and Fairfax folds. The spread of these structural data probably results from the superposition of the late folds upon already complexly folded rocks, resulting in variations in style, plunge direction, and amount of plunge, a common characteristic of polydeformed terranes (see, for example, Tobisch, 1967).

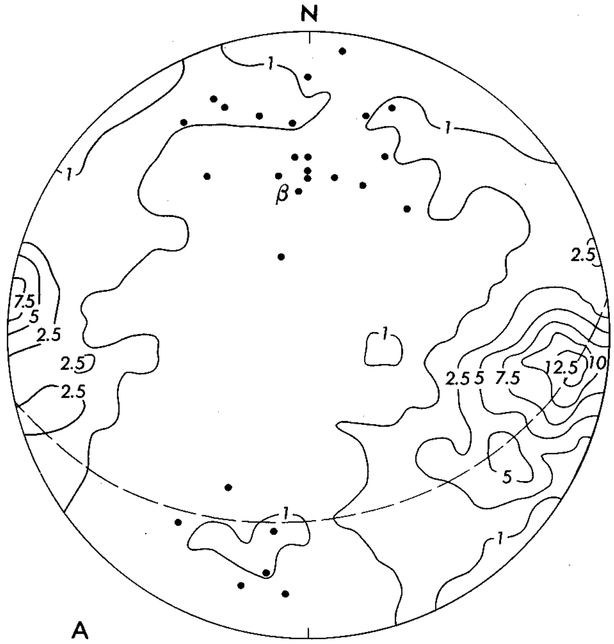
Late folds in the Peters Creek Schist

As pointed out above, the rocks of the Peters Creek Schist have had a much more complex geologic history than have the other rocks in the area. Regionally, the entire formation is inverted as recognized by earlier investigators (Fisher, 1963; Hopson, 1964; and Philip Osberg, oral commun., 1976). This interpretation is based on younging directions read from the abundant sedimentary structures preserved in even the highest grade rocks. The rocks are overturned to the west, and the structure flattens to the west and approaches horizontal at the contact with the Mesozoic rocks of the Culpeper basin (Drake and others, 1979). Transposition is common throughout most of the unit's outcrop belt, and first-generation schistosity is nearly everywhere almost parallel to bedding and forms the tectonic surface called "bedding schistosity" by Hopson (1964). Small fold hinges can, however, be found tucked between the schistosity at many places. Transposition becomes stronger as metamorphic grade increases, and the lenticularity of the graywacke intervals within the Peters Creek is thought to have resulted from transposition much like that which produced the lenticular rock bodies in other orogens, such as the Scandanavian Caledonides (see, for example, Williams and Zwart, 1977).

Cloos (1964) describes an abrupt change in lineation style in the Peters Creek Schist (his Wissahickon) from fold axes plunging north-northeast to northeast, which he believed to be parallel to the *b* kinematic axis of deformation, to lineation that plunges down the dip of the foliation that be believed to be parallel to the *a* kinematic axis of deformation. Superficially, this change in lineation attitude is real, although there are many steep fold axes plunging down the dip of the schistosity of his *b* axis belt as well as many gently plunging lineations within his *a* axis belt (Drake and others, 1979). The solution to this problem is polydeformation that was not recognized in the Maryland outcrop belt. The prominent steeply plunging downdip lineation becomes apparent just downstream from Bear Island, where the effects of strain increase dramatically in the rocks, which eventually pass into phyllonites farther downriver (Drake and Morgan, 1981). The steeply plunging lineations are dominant in this outcrop belt downstream to the overlap of Coastal Plain deposits (Cloos and Cooke, 1953). The transition zone of gently northeast plunging lineations to steeply northwest-plunging lineations can be traced at least to the Montgomery-Howard County line in the zone of high strain in the Peters Creek Schist parallel to its contact with the Sykesville Formation (Cloos and Cooke, 1953) The steep lineations consist of aligned minerals, bedding-schistosity intersections, schistosity-schistosity intersections, and relict fold hinges. None of these steep lineations are kinematic *a* lineations in the strict sense, but are fold-axes and fold axis-related lineations that have been rotated into parallelism with the axis of maximum strain within the zone of high strain in the Peters Creek Schist, as were those of the Clifton fold-phase described above, within the Popes Head Formation.

Although the structural geometry of the Peters Creek Schist has not been analyzed, it obviously has had a complex structural and metamorphic history prior to the formation of the phyllonite along its eastern (lower) boundary. The phyllonite has undergone all the structural events recognized in the Popes Head Formation, as Clifton, Accotink Creek, and Fairfax folds can be traced to the contact of the Peters Creek Schist and Popes Head Formation and are found to fold the contact. These folds are particularly well shown at the north end of the Clifton syncline, in the northern part of the Fairfax quadrangle, and the east-central part of the Manassas quadrangle (Figure 1).

A plot of poles to phyllonitic foliation (Figure 4A) forms a great-circle girdle that has a statistical fold axis of 50°N. 10°W. This axis, and the strong maximum of poles just east of north, suggests that the regional distribution of the phyllonitic foliation results from Accotink Creek folding. Because of their orientation, most of the measured fold axes appear to be Accotink Creek phase, although some representation of Clifton phase and probably a few Fairfax axes are present. Lineations on phyllonitic foliation (Figure 4B) plot on a great-circle girdle that is essentially the same as the girdles in the plots of fold axes and intersection lineations in the Popes Head



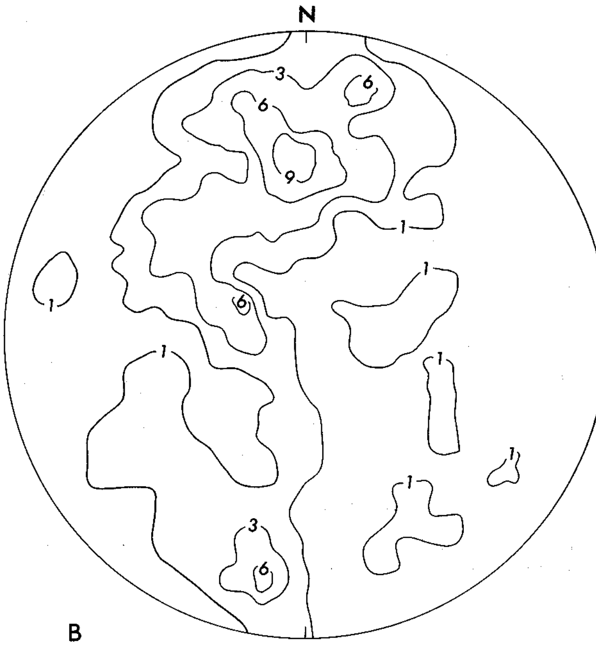


Figure 4. Equal-area projections (lower hemisphere) of some structural elements in the Peters Creek Schist.
A. 134 poles to phyllonitic foliation. Dashed line, best-fit great circle to girdle. Dots, axes of small folds in phyllonitic foliation. B, axes to great-circle girdle. Contours at 12.5, 10, 7.5, 5, 2.5, and 1 percent per 1-percent area.
B. 112 lineations on phyllonitic foliation. Contours at 9, 6, 3, and 1 percent per 1-percent area.

Formation (Figures 2B and 2C). This plot reflects late folding of both units, as these lineations are fold axes or fold-axis-related lineations.

To summarize, phyllonitized Peters Creek Schist has undergone the complete deformational history of the unit that unconformably overlies it. Its phyllonitization, in turn, post-dates polydeformation and high-grade metamorphism. The structural geometry of the phyllonitized Peters Creek Schist verges east with that of the Popes Head Formation, rather than west as it does in the area away from the Clifton syncline. Clifton and Accotink Creek folds can be traced north from the plunging nose of the Clifton syncline to the Potomac River (Drake and others, 1979), where the common structural style is eastward-verging, north- and north-northeast-trending folds, the typical Clifton-Accotink Creek relation. No modern geologic mapping has been done in this part of Maryland, so the extent of these fold phases is unknown, but the structural geometry of the rocks in the District of Columbia and adjacent Montgomery County verges east (Cloos and Cooke, 1953).

TIME OF DEFORMATION

The stratigraphic and structural features of the allochthonous rocks beneath the Popes Head Formation indicate a complex and varied structural history prior to Popes Head deposition. Subsequently, the allochthonous rocks and the Popes Head Formation were thrice folded. At least some of this deformation was reported to predate the emplacement of the Occoquan Granite and the tonalite plutons in northern Virginia (Drake and Lyttle, 1981; Drake and Morgan, 1981). I now conclude that most, if not all, the deformation in northern Virginia can be shown to predate or to be synkinematic with the emplacement of these intrusive rocks.

The contact between the Popes Head Formation and Occoquan Granite can be seen at the Bull Run Marina in the Independent Hill quadrangle (BR on Figure 1). Here, veins and seams of granite occur in the schistosity of the Popes Head, and the foliation (N. 2°E., 80°NW.) in the granite is roughly parallel to the schistosity (due N., 73°W.) in the Popes Head, which suggests that the granite was emplaced syntectonically during Clifton phase-folding of the Popes Head. Syntectonic emplacement of the Occoquan is also suggested by evidence from the Accotink Schist. In this unit, Drake and Lyttle (1981) reported that during a second metamorphism chlorite and garnet porphyroblasts had begun to grow before the formation of a second schistosity but were subsequently rotated by that schistosity. Drake and Lyttle (1981) suggested that the second metamorphic event was thermally related to the emplacement of the Occoquan Granite and the tonalite plutons. The second schistosity in the Accotink Schist is thought to have formed during Clifton folding. This seems to be good evidence for essentially contemporaneous deformation and plutonism.

Mineral and quartz-rod lineations measured in the northern part of the Occoquan Granite batholith and the tonalite plutons are plotted on Figure 5. They plot in two strong maxima: 42°, S. 30°W. (and lesser N. 30°E.-plunging equivalents) and 35°, N. 58°W. The south-southwest plunging lineations certainly appear to be related to Clifton phase folding, and the spread is wide enough to accommodate Accotink Creek folding. The northwest-plunging lineations appear to be related to Fairfax-phase folding.

The map pattern (Figure 1) suggests synkinematic emplacement of the Occoquan Granite and clearly shows that the Occoquan intrusion postdated Tripps Run and Ravenwood folds. The Falls Church pluton and the largest of the tonalite plutons cut the axis of the Ravenwood anticline (Figure 1).

It is possible that the Popes Head Formation was thrust upon the northern Virginia tectonic motifs after Clifton, Accotink Creek, and Fairfax folding and that the rocks of the motifs were not involved in this folding. I noted earlier the theoretical chance that the lower grade, less complexly deformed Popes Head had been tectonically emplaced above these other rocks (Drake and Lyttle, 1981, p. 12,; Drake and Morgan, 1981, p. 494). At that time, I pointed out that no direct evidence supports tectonic emplacement and that the bedding in the Popes Head parallels its contact with the underlying rocks, the classic piece of evidence to support an unconformable relationship. In addition, the Popes Head is so clearly infolded with the underlying rocks, particularly the Peters Creek Schist, that it must have been folded after its emplacement even if it had been originally emplaced by thrusting. One other piece of evidence negates the thrust emplacement of

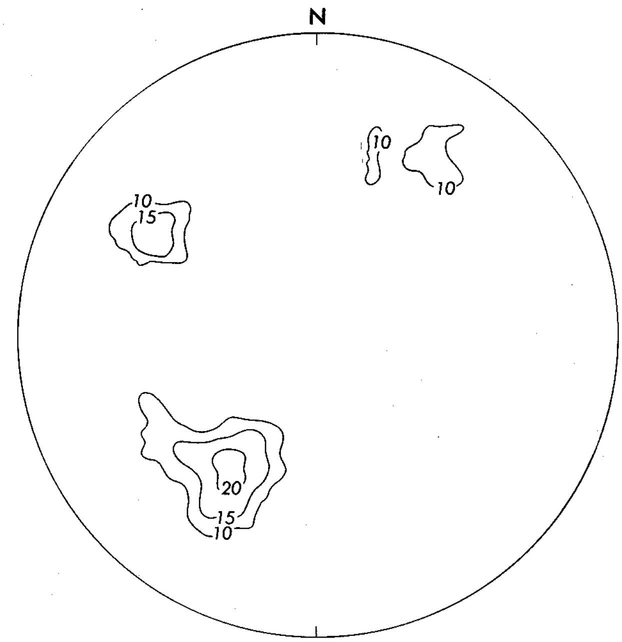


Figure 5. 40 mineral quartz rod lineations in tonalite plutons and northern part of the Occoquan Granite batholith. Contours at 20, 15, and 10 percent per 1-percent area.

an already folded Popes Head. I have shown that the Occoquan Granite was emplaced synkinematically with Clifton phase folding. If the Popes Head were thrust, the Occoquan had to accompany it. The Occoquan Granite, however, also intrudes the Annandale Group, the Sykesville Formation, and the Indian Run Formation. The Popes Head then had to be above these units prior to the emplacement of the Occoquan Granite and the onset of Clifton folding.

These data show that all the deformation events described herein are either synchronous with or predate the emplacement of the Occoquan Granite. Conclusions are that Clifton, Accotink Creek, and Fairfax folding are related to a single deformation synchronous with granite emplacement. The folding can then be dated as either Early (if the zircon dating is correct) or Late (if the Rb-Sr isochron is correct) Cambrian. Pre-Taconian (pre-Arenig) deformation was first recognized in the Appalachians by Neuman (1967) in northern Maine. Neuman termed this orogenic event "the Penobscot disturbance." More recently, Pavlides (1973) recognized a fold phase in the Chopawamsic Formation of Early Cambrian (?) age that is absent in the overlying Quantico Formation of Ordovician age in the

Chopawamsic terrane southeast of Fairfax County, Virginia. He suggested that this pre-Taconic deformation might have resulted from the Penobscot event. Recent work has shown that Cambrian deformation is much more widespread than originally known (Philip Osberg, oral commun., 1983). I believe that the Clifton folding is a manifestation of the Penobscot orogenic period. On the European side of the Appalachian-Caledonian orogen, a major, probably *the* major, phase of the Caledonian orogenic cycle had been termed "Grampian." The Grampian was post-Early Cambrian and pre-Arenig in age and has been isotopically dated as having taken place between 550 and 510 Ma (Rast, Nicholas and Crimes, 1969). Apparently, Cambrian deformation was much more important in the Appalachian-Caledonian orogen than had been recognized previously. I believe that the Penobscotian deserves the full title of "orogeny" to better correlate with the same event, the Grampian orogeny, on the other side of the Appalachian-Caledonian orogen.

The pre-Clifton structural events described above in the Annandale Group and Peters Creek Schist are thought to be pre-Penobscotian (Grampian) and, therefore, probably date from the Late Proterozoic or earliest Cambrian. Such deformation has been termed "Avalonian" (Lilly, 1966) in the northern Appalachians and "Virgilinian" in the southern Appalachians (Glover and Sinha, 1973). Deformation of similar age on the European side of the Appalachian-Caledonian orogen has been termed "Cadomian" by Rast and Crimes (1969) and isotopically dated as having taken place between about 570 and 600 Ma. Although some might choose to use the North American term "Virgilinian" for this deformation, I prefer to follow recent authors (Keppie, 1979; Rast, 1980; and Osberg, in press) and use "Cadomian." This acceptance of Cadomian nomenclature supports the philosophy of the I.G.C.P. in its attempt to standardize terminology on both sides of the orogen.

One final factor bearing on the isotopic age of deformation must be considered—the relation of the small intrusive bodies of Bear Island Granite to the tectonic synthesis of time of deformation presented herein. Two samples of muscovite from the granite have been dated at 469 ± 20 Ma by Rb-Sr techniques (Muth and others, 1979). The Bear Island Granite has been reported as post-dating all but the latest folds (Fisher, 1971), and I have never seen an outcrop of deformed Bear Island Granite. Along the Southern Railway in the Manassas quadrangle, a small crosscutting body of very leucocratic, muscovite-bearing granite, containing 48 percent albite and 5 percent microcline, contains inclusions of Peters Creek phyllonite, so it must postdate that late tectonic event. This granite body is identical with the undeformed bodies of Bear Island granite on Bear Island and in Difficult Run and is believed to be Bear Island Granite. Leucosome that formed during migmatization in the Peters Creek, however, is sheared along the entire strike belt of the phyllonite zone from the Potomac River to the Manassas quadrangle. Hopson (1964) reported that granite sheets concordant to the foliation in the Peters Creek Schist along the Potomac River have participated in the late deformation there, and that more than one generation of granite probably exists. The Bear Island Granite apparently resulted from a thermal event that postdates the deformation.

TECTONIC IMPLICATIONS

Cadomian and Penobscotian deformations have not been recognized in rocks that are considered to be North American by most geologists. Certainly, the North American rocks in the Blue Ridge west of the Culpeper Mesozoic basin and the area of this study were not deformed during those orogenic periods, and, in fact, there is no direct evidence that they were deformed prior to the Alleghanian orogeny (Drake, 1980). The Piedmont rocks described in this paper, then, should be considered part of a suspect terrane or terranes. Williams and Hatcher (1983) have placed the area discussed herein in their Piedmont suspect terrane.

Tectonic relations, however, are probably more complex than that. The rocks beneath the Popes Head Formation constitute different lithotectonic environments, including oceanic (Piney Branch Complex) and two different submarine fans, one of which contains ophiolitic debris (Peters Creek Schist) and one of which does not (Annandale Group). These rocks were deformed during the Cadomian, but all have a different internal structural geology and, therefore, appear to be fragments of different terranes. Subsequently, these fragments were assembled into a stack of tectonic motifs (Drake, 1985) prior to the deposition of the Popes Head Formation and the onset of Penobscotian deformation. The sedimentation of the Popes Head and the Penobscotian deformation combined these units into what is here called the "composite Potomac terrane." To the west, in southern Maryland, the Potomac terrane is in contact with the

Ijamsville Phyllite, which is thought to be a Taconic allochthon (Drake and Lyttle, 1981), and, by the usage of Williams and Hatcher (1983), would itself constitute a suspect terrane. The boundary between the Potomac and Ijamsville terranes in Montgomery County, Maryland, is marked by a zone of phyllonitized Peters Creek Schist. Perhaps this phyllonite zone is the Taconian suture in this part of the Appalachians.

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RELATIONSHIPS OF STRUCTURE TO MASSIVE SULFIDE DEPOSITS IN THE CHOPAWAMSIC FORMATION OF CENTRAL VIRGINIA

James F. Conley

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ABSTRACT

The Chopawamsic Formation in central Virginia consists of a lower unit composed of mafic and intermediate metavolcanic rocks and metagraywacke beds and an upper unit composed of interlayered felsic and mafic metavolcanic rocks. Massive sulfide deposits and ferruginous quartzites (metacherts, which probably represent volcanic exhalants) occur as interlayers in the Chopawamsic in this area.

The Chopawamsic has been deformed by four fold systems, one of which developed prior to deposition of the overlying Arvonia Formation. Size of stratabound massive sulfide deposits in the Chopawamsic Formation is in part controlled

by distance from volcanic centers and in part by polyphase folds. Most prospecting for massive sulfide deposits in the Chopawamsic has been on the limbs of folds; however, the larger ore bodies should occur within the hinge zones of folds. Plastic flowage of massive sulfide deposits in response to stress should result in thickening of these deposits in hinge zones of folds and thinning on the limbs of folds. Cross sections prepared from drill hole data and geologic maps of one of the massive sulfide deposits support this concept that massive sulfides are stretched out along the limbs of folds and have flowed into the hinge zones of these structures.

INTRODUCTION

The Chopawamsic Formation was named by Southwick, Reed, and Mixon (1971) for exposures along Chopawamsic Creek in northern Virginia (Figure 1), where it is composed of metamorphosed felsic and mafic volcanic flows and volcaniclastic rocks with interlayered quartzite, graywacke, and phyllites. Age of the Chopawamsic is considered to be Cambrian, possibly Early to Middle Cambrian (Glover, 1974; Mose, 1981; Pavlides, 1981). It has been mapped southwestward by Pavlides (1976, 1980), into an area just northeast of the present study area. The Chopawamsic forms part of a major northeastward-trending belt of rocks named the "Atlantic seaboard volcanic province" (Higgins, 1972). In central Virginia the Chopawamsic contains massive sulfide deposits and underlies the gold belt mapped by Taber (1913) and the gold-pyrite belt mapped by Lonsdale (1927). Rocks of the Chopawamsic were subdivided by Conley and Johnson (1975) into a lower unit, composed of clastic metasedimentary rocks and an upper unit, composed of metavolcanic rocks. The Chopawamsic Formation has been traced southwest to the limit of mapping shown on Figure 2.

REGIONAL SETTING

In central Virginia, the Chopawamsic Formation overlies the Candler Formation. Although field mapping indicates that the Candler grades upward into the Chopawamsic (Conley and Johnson, 1975), VIBROSEIS data obtained by the U. S. Geological Survey along U. S. Interstate I-64 indicate that a fault could be present in the vicinity of this contact (Leonard D. Harris, personal communication). The base of the Chopawamsic is placed at the lower contact of the first recognizable volcanic layers, which commonly are of metabasaltic composition. The base of the formation approximates the contact between rocks mapped as the lower chlorite-muscovite unit and the middle muscovite unit by Smith, Milici and Greenberg (1964). The top of the Candler contains abundant, thin metagraywacke interlayers especially in the horizon just below the first volcanic bed which marks the base of the Chopawamsic. The Chopawamsic is unconformably overlain by the Middle to Late Ordovician Arvonia Formation (Darton, 1892; Watson and Powell, 1911; Tillman, 1970). Brown (1969) proposed that the Arvonia was overlain by the Buffards Formation, but later mapping (Conley, 1978; Conley and Marr, 1979; Marr, 1980b) shows that the Buffards occupies an antiformal structure and is composed in part of basal Arvonia and in part of rocks of the underlying Chopawamsic Formation.

In central Virginia, metamorphic grade in the Chopawamsic gradually rises to the southeast. The location of the almandine isograd across the Dillwyn quadrangle (Brown, 1969) and across Fluvanna County (Smith, Milici and Greenberg, 1964) indicates that the trend of the isograd is more easterly than the trend of the stratigraphic units. The increase in metamorphic grade caused volcanic textures to be obliterated to the southeast as the rocks were converted to gneisses, schists and amphibolites.

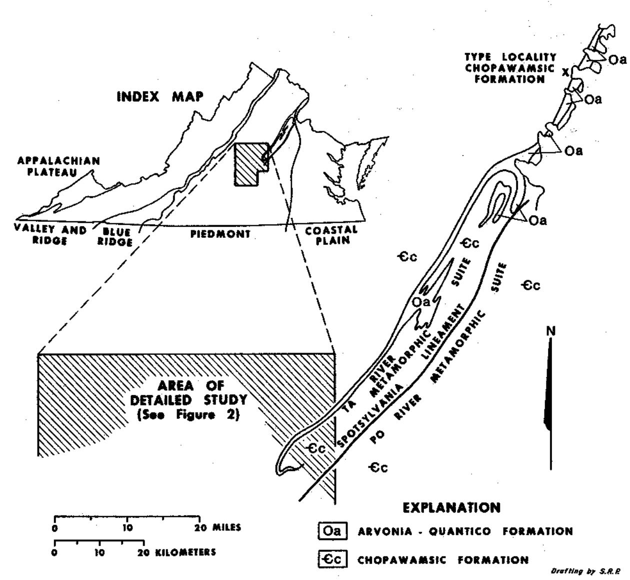

Figure 1. Regional map of the Virginia Piedmont showing area of study.

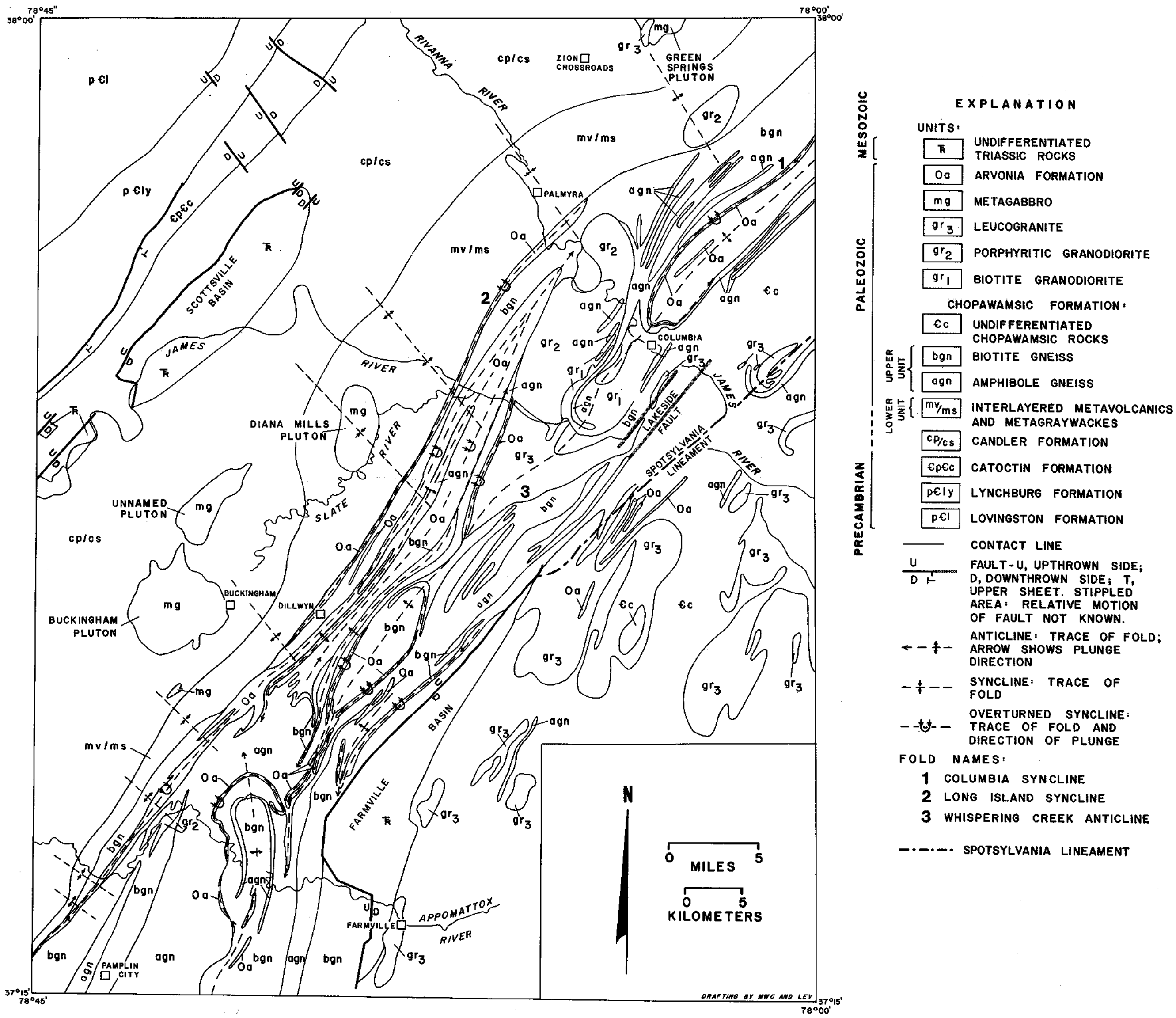

Figure 2. Geologic map of the Central Virginia Piedmont.

Pavlides (1980) mapped the low-rank rocks west of the Quantico syncline in northern Virginia (Figure 1) as the Chopawamsic Formation. He called the high-rank rocks east of this syncline the Fredericksburg Complex. Pavlides (1980) subdivided the Fredericksburg Complex into a western Ta River Metamorphic Suite, composed predominantly of amphibolite gneiss, and an eastern Po River Metamorphic Suite composed chiefly of biotite gneisses. This directional change in composition is also seen in central Virginia (Figure 2). He considered the Ta River Metamorphic Suite to be an eastern facies of the Chopawamsic. Marr (1980a; 1980b), from mapping in the Central Virginia Piedmont agrees that higher rank rocks on the southeast side of the Arvonia syncline are part of the Chopawamsic Formation. These high rank rocks are probably also equivalent to the Hatcher Complex named by Brown (1969), in central Virginia, which is also thought to be a relatively high-grade metamorphic equivalent of the Chopawamsic (Conley, 1978, Marr, 1980a).

The Spotsylvania lineament (Pavlides, 1980) marks the boundary between the Ta River Metamorphic Suite and Po River Metamorphic Suite (Figure 1). The lineament was first recognized by Neuschel (1970) as a linear anomaly evident on both aeroradioactivity and aeromagnetic maps. Neuschel postulated that it was a fault zone and Pavlides (1980) notes that it is a diffuse zone up to 2 km wide and speculates that it might represent a series of en echelon faults or shear zones. In central Virginia the Spotsylvania lineament is traceable southward on aeroradioactivity and aeromagnetic maps into the northern end of the Farmville basin, beyond which it is covered by sedimentary rocks of Triassic age (Figure 2). VIBROSEIS data indicates that the Spotsylvania lineament is the leaning edge of a major thrust sheet (Harris and others, 1982) that separates the Ta River Metamorphic Suite from the Po River Metamorphic Suite. In central Virginia Farrar (1985) has called rocks equivalent to the Po River Metamorphic Suite the Goochland granulite terrane composed of the State Farm gneiss and overlying granulite grade rocks of possible Grenville age. Furthermore those rocks of the Goochland granulite terrane are correlated with rocks of the Raleigh belt along strike to the southwest (Farrar, 1985).

STRATIGRAPHY

In central Virginia the Chopawamsic Formation was subdivided by Marr (1980a, 1980b) into smaller, mappable units (Figure 3). The lower unit of the Chopawamsic Formation is composed predominantly of pink-weathering metasandstones and metasiltstones of graywacke composition. It also contains interlayered felsic and mafic metavolcanic rocks. Purple phyllites, typical of the Candler Formation, occur as interbeds in the lower part of the unit.

To the southeast and presumably up-section, the metagraywackes are replaced by rocks of volcanic origin. The formation is divided into a lower and an upper unit; the boundary is placed where metavolcanic rocks, which increase in abundance upward, comprise 50 percent of the formation. The upward increase in metavolcanic rocks is gradational, but the transitional interval (from over 50 percent metasedimentary rocks to over 50 percent metavolcanic rocks) averages only a few hundred meters across. Felsic metavolcanic rocks predominate over mafic rocks near the base of the upper unit, whereas mafic rocks are more prevalent higher up in the section.

Chopawamsic rocks are at greenschist grade along the northwestern boundary of the unit's outcrop belt (Figure 2). The felsic rocks throughout the formation are composed of quartz, plagioclase, perthitic microcline, muscovite and biotite. Quartz occurs as embayed pseudomorphs, which locally retain the outline of beta phenocrysts. Some of the layers are amygdaloidal metabasalts composed of plagioclase and chlorite with lesser amounts of biotite and epidote, in which amygdules filled with quartz and epidote make up as much as 30 percent of the rock. There are some metavolcanic rocks of intermediate composition; these commonly contain actinolite. Metagraywackes are composed of quartz (which occurs both as granule-size clasts and in the finer-grained matrix of the rock), plagioclase, perthite, muscovite, and biotite. To the southeast where the rocks are at higher metamorphic grade, the mafic rocks are metamorphosed to amphibolites, the intermediate rocks are metamorphosed to amphibole gneisses, the felsic rocks are metamorphosed to quartzofeldspathic gneisses and the metagraywackes are metamorphosed to biotite gneisses.

Ferruginous quartzites are interlayered with the metamorphosed metavolcanic and metasedimentary rocks in the lower unit and lower part of the upper unit of the Chopawamsic. These quartzites were recognized by both Taber (1913) and Lonsdale (1927). Taber (1913) observed gold-bearing veins in these quartzites, and Good, Fordham, and Halladay (1977) suggested that the quartzites may be favorable areas for gold and base-metal exploration. Hodder, Kazda, and Bojtos (1977) were the first to report that sulfide deposits in the Chopawamsic represent volcanic exhalations and Marr (1980a and 1980b) has traced these ferruginous quartzites into massive sulfide bodies. The ferruginous quartzites grade laterally into and also overlie the massive sulfide deposits. The ferruginous quartzites probably represent metamorphosed iron-bearing cherts. Such cherts, precipitated under oxidizing conditions from submarine volcanic vents have been described by Appel (1979). Cherts rich in iron oxide, precursors of the ferruginous quartzites here discussed, are normally deposited at some distance from a volcanic vent (Large, 1977), but during the waning phase of a depositional cycle are deposited closer to the source and may even be deposited on the top of massive sulfide bodies. On the other hand, the massive sulfide bodies are proposed to represent stratabound deposits formed under reducing conditions as a product of quenching near a volcanic vent (Soloman and Walshe, 1979; Appel, 1979); thus they thin rapidly away from the vent. In the study area, the massive sulfide deposits are generally covered by gossan. Primary minerals observed in drill cores of these deposits are: (in order of decreasing abundance)

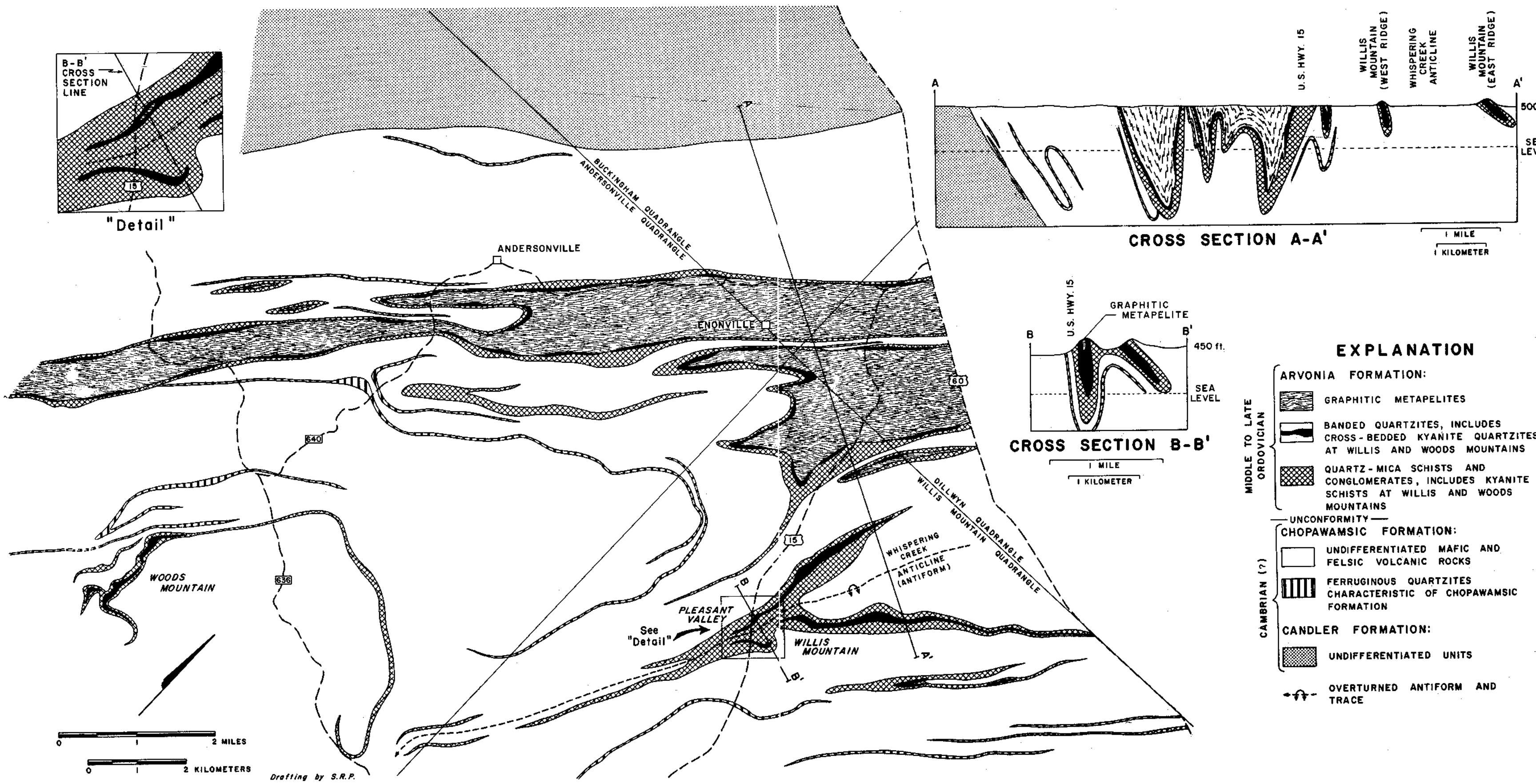

Figure 3. Sketch showing stratigraphic relationships within the Chopawamsic Formation as well as between the Chopawamsic and underlying Candler and overlying Arvonia formations.

pyrite, pyrrhotite, chalcopyrite, sphalerite, and galena (Young, 1981).

STRUCTURE

Four fold systems are recognized in rocks of the Chopawamsic Formation and are defined in the Table. Brown and Griswold (1970) described small rootless isoclinal folds (F_1) in the Chopawamsic that do not occur in the unconformably overlying Arvonia Formation. Similarly in northern Virginia Pavlides (1976) has recognized folds in the Chopawamsic Formation that are not present in the unconformably overlying Quantico

Table. Primary and secondary structures in the Chopawamsic and Arvonia formations of central Virginia.

Feature		Comments
So_a:	Compositional layering in the Chopawamsic Formation.	Transposed bedding.
F_1:	Outcrop and map scale isoclinal folds in Chopawamsic Formation.	Defined by compositional layering in the Chopawamsic Formation; locally the folds are rootless.
S_1:	Foliation parallel to axial surfaces of F_1 folds.	Has transposed primary layering producing rootless folds in the Chopawamsic Formation.
L_1:	Plunging hinges of F_1 folds and stretched mineral grains.	Compositional layers and mineral grains transposed by S_1.
So_b:	Bedding in the Arvonia Formation.	Defined by textural and color changes.
F_2:	Steeply inclined, outcrop-scale isoclinal folds in Chopawamsic and Arvonia formations.	Oldest folds in Arvonia Formation.
S_2:	Foliation developed parallel to axial surfaces of F_2 folds.	The major foliation of both the Chopawamsic and Arvonia formations in the northwestern part of the area.
L_2:	Lineation in the B direction of F_2 folds.	Occurs in the Chopawamsic Formation as mineral streaks and fold hinges and is produced in the Arvonia Formation as a line of intersection of So_b and S_2.
F_3:	Northeastward-trending, map-scale tight upright to slightly asymmetrical folds.	Flanks contain synformal F_2 folds; noses defined by refolded F_2 folds. Extremely elongate folds have folded F_2 folds producing a fishhook-shaped outcrop pattern.
S_3:	Vertical to steeply southeastward dipping crenulation	Becomes better developed to the southeast where it is the major foliation,

Table (cont.).

Feature		Comments
	cleavage that is parallel to axial planes of F_3 folds.	possibly because these rocks are at higher metamorphic grade.
L_3:	Lineation produced by intersection of S_2 and S_3 surfaces.	
F_4:	Broad, upright, north-westward-trending, gentle open warps.	Map-scale folds.
S_4?:	Vertical, widely-spaced, northwestward-trending kink bands that are probably parallel to the axial planes of F_4 folds.	

Formation. These Pre-Arvonia folds are observable in outcrop (Figures 4, 5, 6, 7a, and 7b). Two additional fold systems (F_2, F_3) that occur in both the Chopawamsic and Arvonia formations were recognized by Brown and Griswold (1970). Surfaces of these folds are illustrated in Figures 8, 9, 10, and 11. A fourth set of northwest-trending, broad, gentle open folds (F_4) with wave lengths in the order of several miles occur on a regional scale (Figures 2 and 12). These folds were first recognized by Conley and Toewe (1968) in the southwestern Virginia Piedmont and have since been recognized by Henika (1977) in the area east of Danville and by other workers in the central Virginia Piedmont (Poland, Glover, and Mose, 1979).

Analyses of foliations and lineations measured in the Willis Mountain and Andersonville quadrangles are shown on equal-area, lower-hemisphere diagrams (Figure 12). These figures include data from the Arvonia Formation (Figure

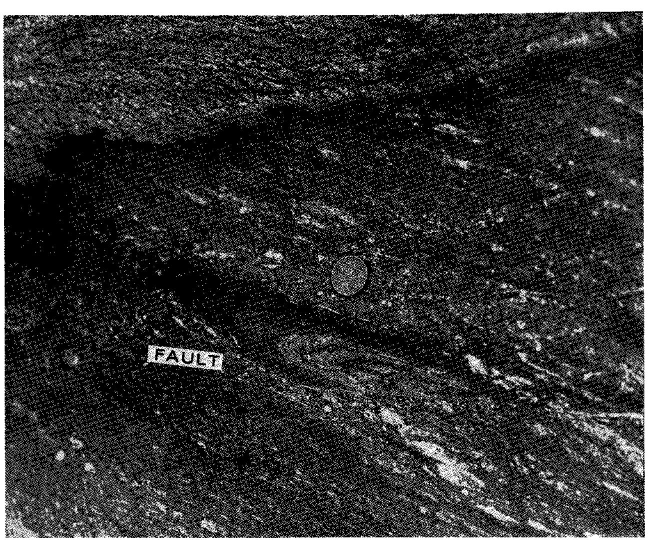

Figure 4. The contact between felsic (light) and mafic (dark) metavolcanic rocks in the Chopawamsic Formation. The contact represents a bedding plane (So_a). Penetrative foliation (S_1) cut the bedding surface at an acute angle and has transposed compositional layers into a series of rootless isoclines. There is a slight refraction of the angle of the S_1 surface between the mafic and the felsic layer. Note that a fault can be traced horizontally across the lower part of the photograph. Locality is along Holiday Creek, about 300 feet (91 m) west-northwest of State Road 614 (Holiday Lake quadrangle).

Figure 5. Rootless isoclines in the Chopawamsic Formation formed by transportation of compositional layers (So_a) by foliation (S_1). Locality is along Holiday Creek about 300 feet (91 m) west-northwest of State Road 614 (Holiday Lake quadrangle).

Table (cont.).

Feature		Comments
	cleavage that is parallel to axial planes of F_3 folds.	possibly because these rocks are at higher metamorphic grade.
L_3:	Lineation produced by intersection of S_2 and S_3 surfaces.	
F_4:	Broad, upright, north-westward-trending, gentle open warps.	Map-scale folds.
S_4?:	Vertical, widely-spaced, northwestward-trending kink bands that are probably parallel to the axial planes of F_4 folds.	

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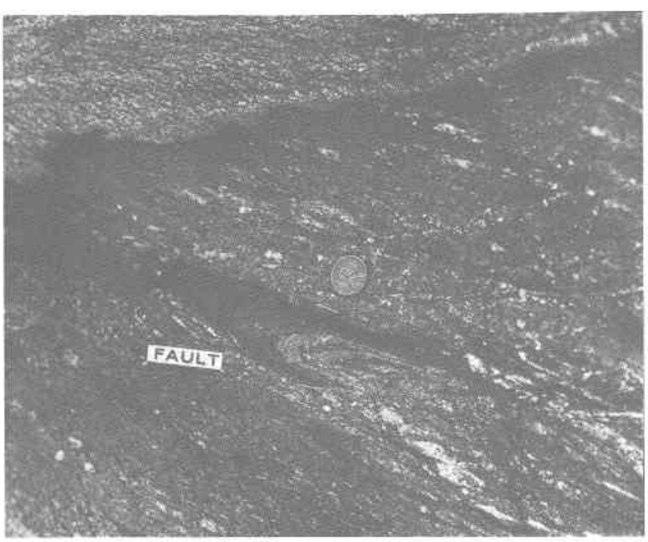

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Figure 6. Rootless isoclines in the Chopawamsic Formation produced by transposition of bedding So_a by penetrative foliation S_1. Second foliation (S_2) is parallel to the stick. Intersection of S_1 and S_2 produces mullions in the lower right hand corner of photograph. A spaced cleavage (S_3) is subparallel to the top and bottom of the photograph. Same locality as Figures 4 and 5.

12c) and from two areas of the Chopawamsic Formation, one northwest of the belt of Arvonia rocks (Figure 12a) and one southeast of this belt (Figure 12b). Both plots of S_1 poles to planes from the Chopawamsic produce a girdle that follows a great circle and support field observations that S_1 foliations were folded by F_2 folds. Figure 12b shows a double girdle, which is interpreted as representing folding of S_1 surfaces by F_2 folds which are asymmetrical and have both shallow and steeply dipping limbs.

Plots of S_2 poles to planes from the Arvonia (Figure 12c), however, form a tight cluster. This concentration of S_2 poles to planes from the Arvonia Formation indicates an axial symmetry for the F_2 folds, and could be interpreted as showing that S_2 foliation in the Arvonia was not markedly reoriented by F_3 folds. The observed symmetry is the result of sampling bias. All the data points for the S_2 plots were collected from the S_2 synform that occupies the northwestern flank of a large, extremely elongate, northeastward-trending F_3 fold in the northwestern part of the Andersonville quadrangle (Figure 2). Because there are no prominent F_3 fold noses that refold this synform, cored by Arvonia Formation, evidence that S_2 surfaces are folded by F_3 folds is not shown by these diagrams.

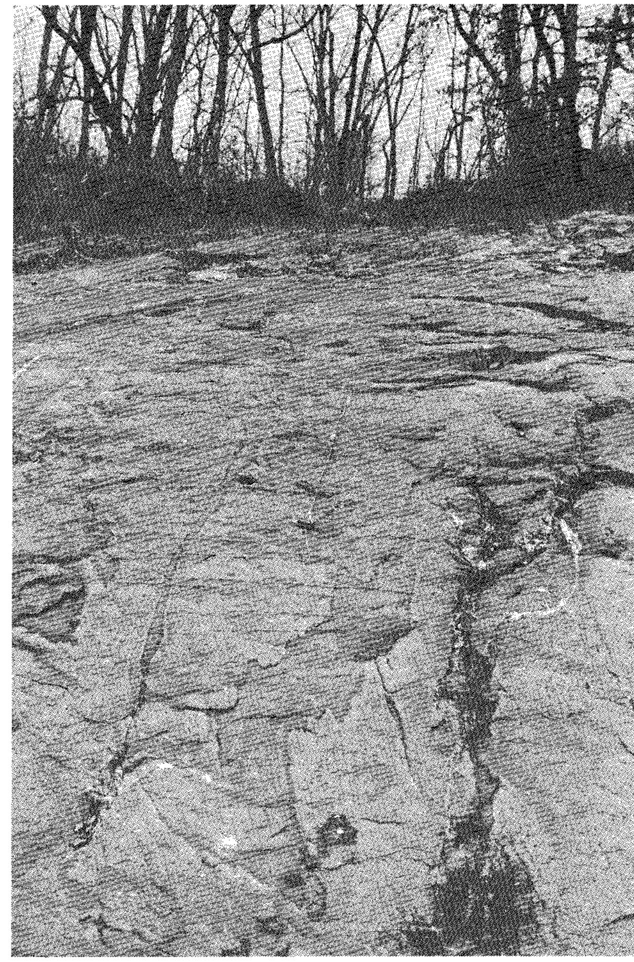
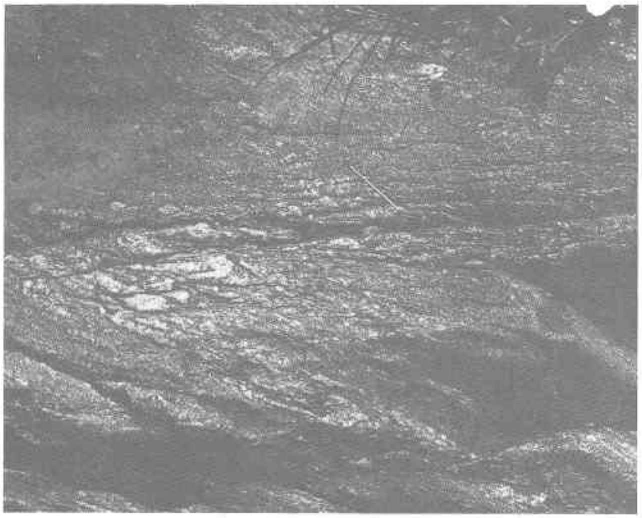


Figure 7a. Large, refolded rootless F_1 isocline of amphibolite refolded by F_2 fold. Sketch shows fold relationships. Locality is on the north bank of the Willis River, where State Road 633 crosses the river, northwest of Arcanum (Andersonville quadrangle).

The concentration of S_1 poles to planes in plots of the Chopawamsic in the area northwest of the outcropping band of Arvonia Formation (Figure 12a) reflects the F_2 axial symmetry seen in the Arvonia Formation (Figure 12c). In the Chopawamsic plot the symmetry is represented as a maximum on a girdle that is located on a great circle, which suggests that these rocks were partially reoriented parallel to S_2 surfaces. In the plots of Chopawamsic lying southeast of the outcropping band of Arvonia this axial symmetry is absent. This diagram represents a much larger area and is the product of warping of S_1 surfaces by the F_2, F_3, and F_4 fold systems.

The plots of mineral lineations in the Chopawamsic (L_1) from both the northwestern and



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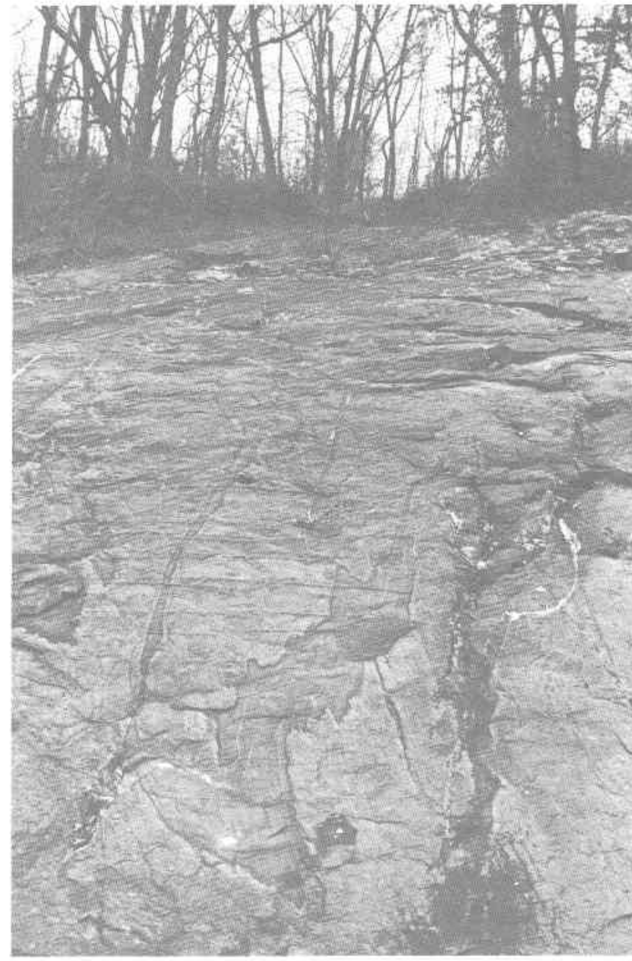

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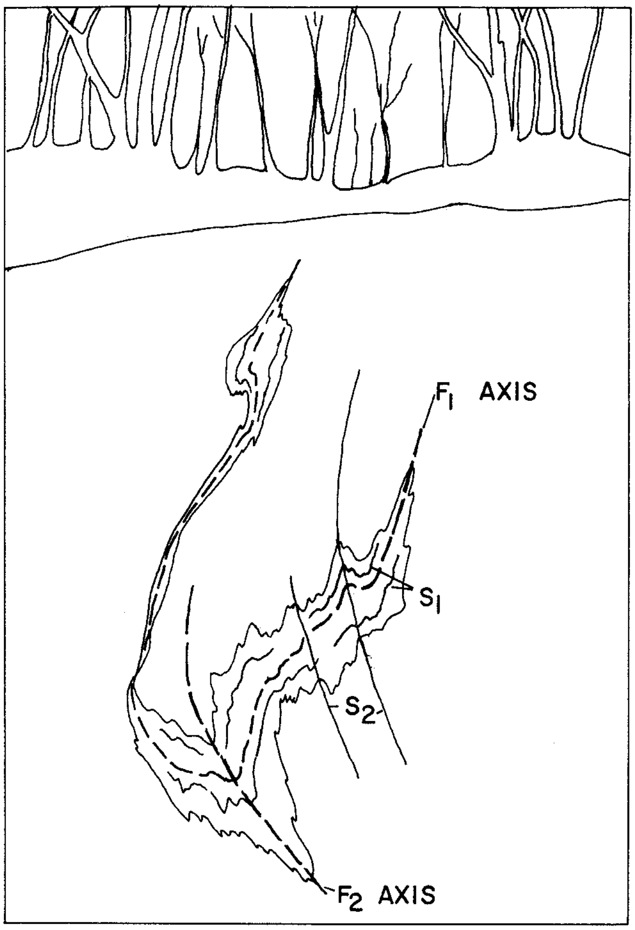


Figure 7b. Tracing of Figure 7a showing F_1 folds refolded by F_2 folds.

southeastern areas (Figures 12a and 12b) show a frequency maximum that is elongated. The spread of the data suggests that L_1 was reoriented by a later fold event, most probably F_2. Contours of plots of L_1 lineations in the Chopawamsic from the rocks southeast of the Arvonia belt (Figure 12b) show plunges to the northeast and to the southwest. This suggests that this lineation was warped during the formation of F_4 folds. Northwest of the belt of Arvonia, the L_1 lineations in the Chopawamsic (Figure 12a) and the L_2 lineations in the Arvonia (Figure 12c) plunge only to the northeast. This apparent conflict in data is also the result of bias in sampling. Data collected from both the Chopawamsic and the Arvonia in the northwest were limited to the northeastern limb of an F_4 fold and those from the area to the southeast were taken from a much larger area that contains both the northeastern

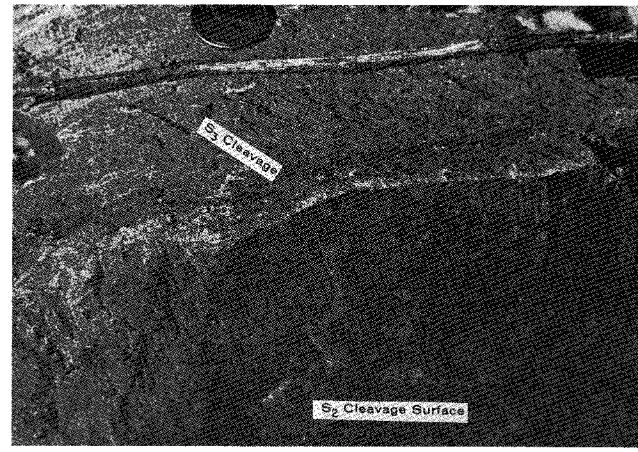
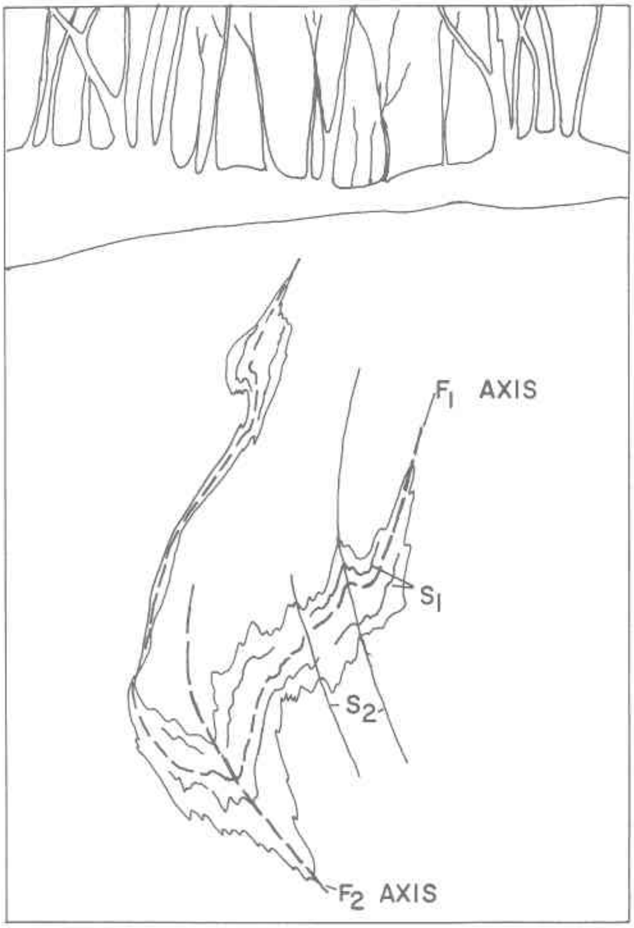


Figure 8. Slaty cleavage (S_2) in Arvonia Formation is parallel to the stick which is laid across the horizontal surface. A spaced cleavage (S_3) oriented from upper left to lower right is outlined by shadows on this surface. The intersection of S_2 and S_3 produces a lineation (L_3) on the vertical surface that dips steeply to the viewer's right. Locality is the Solite Corporation quarry (Arvonia quadrangle).

Figure 9. Slaty cleavage (S_2) surface in Arvonia Formation in northwest wall of Arvonia-Buckingham Slate Company, Inc. quarry. Lineation (L_2) accompanies a slight indentation of the S_2 surface and is oriented from upper right to lower left. The lineation is produced at the acute angle intersection of bedding So_b and slaty cleavage S_2. Nearly vertical spaced cleavage (S_3) at the tip of the knife blade intersects S_2 and produces lineation (L_3) that plunges steeply from upper left to lower right across photograph.



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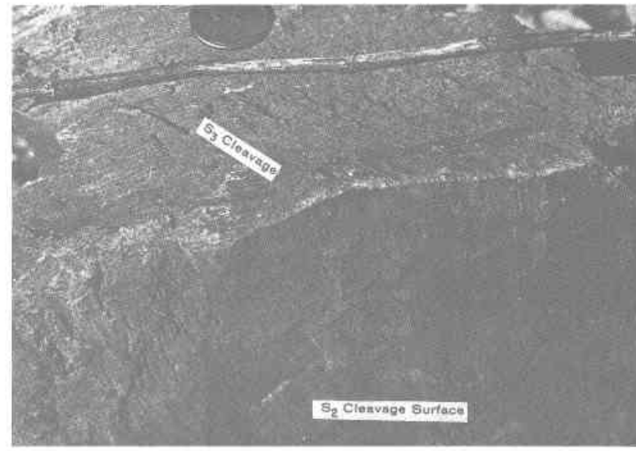

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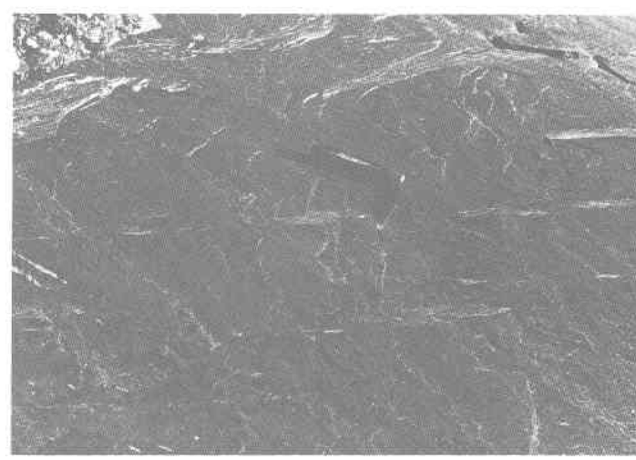

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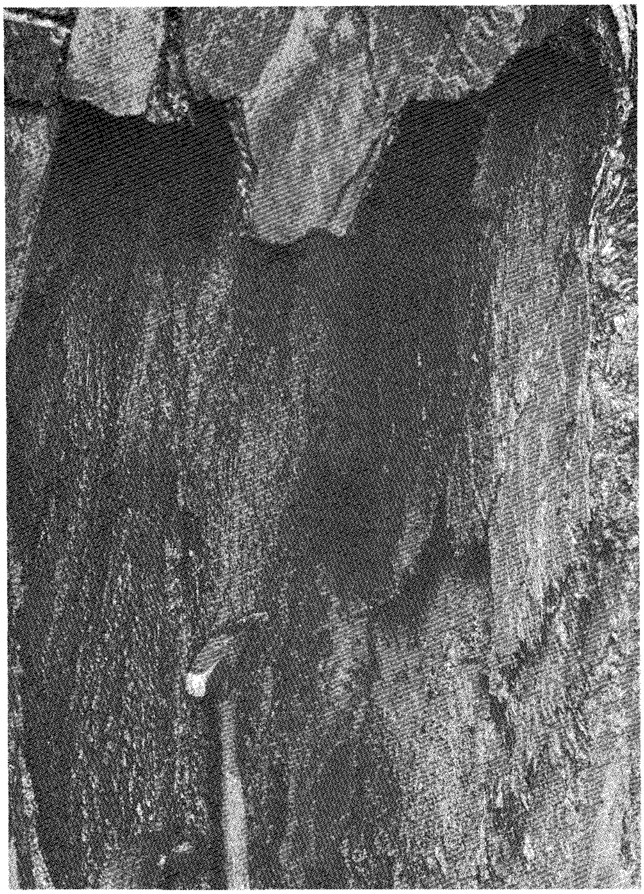


Figure 10. F_2 hinge outlined by So_b bedding in Arvonia Formation exposed in Solite quarry (Arvonia quadrangle). Note the well developed axial plane cleavage, S_2. Hammer is for scale.

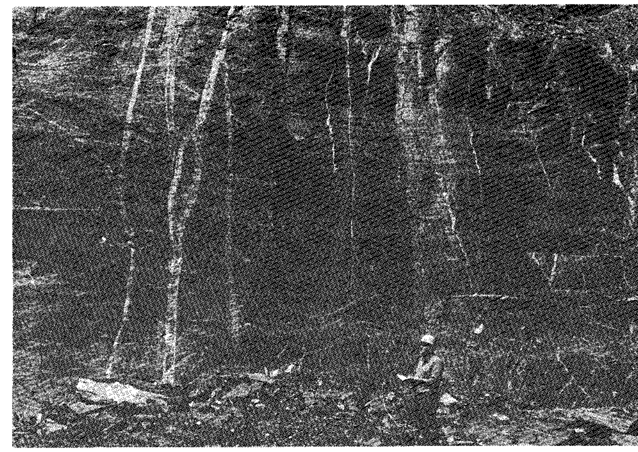


Figure 11. Exposure of slate in the Arvonia Formation exposed in the northwestern wall, Arvonia-Buckingham slate quarry (Arvonia quadrangle). Bedding (So_b) is intersected at extremely acute angle by nearly vertical slaty cleavage (S_2) (which here is parallel to the surface of the northwest wall of the quarry) and produces a lineation (L_{2b}) that can be seen as horizontal lines across the photograph. Non-planar northwest-trending vertical pairs of widely spaced kink bands (S_4?) (light vertical bands) truncate So_b and S_2.

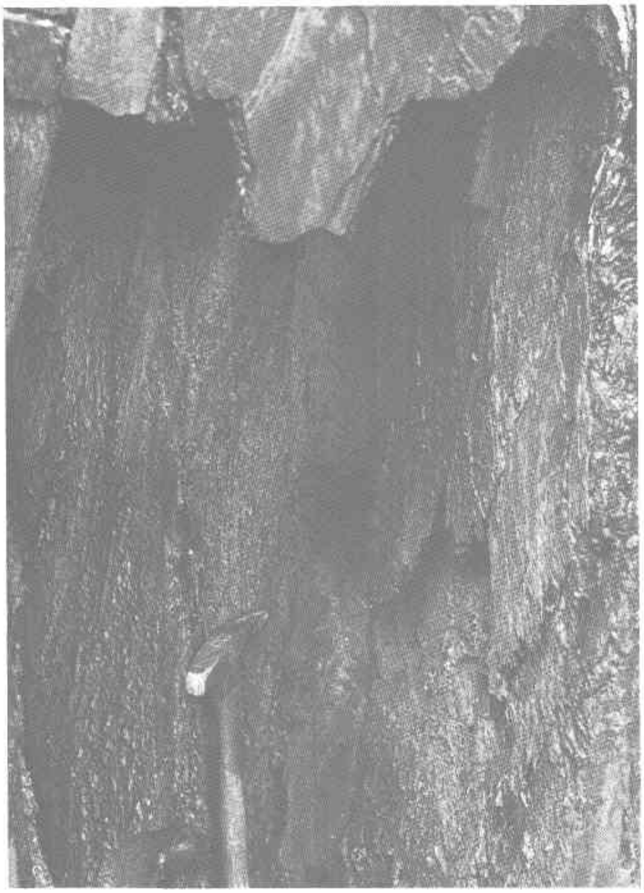
and southwestern limbs of F_4 folds as mapped by Marr (1980a and 1980b).

STRUCTURAL CONTROL OF MASSIVE SULFIDE ORE BODIES

During deposition, stratabound, volcanogenic, massive sulfide bodies commonly develop a primary zonation of minerals. Chalcopyrite and pyrrhotite are concentrated toward the bottom of the body, pyrite and sphalerite are concentrated toward the top, and galena-sphalerite is more abundant in the upper part of the deposit (Sangster, 1972). This zonation may be altered during metamorphism because Carpenter (1974) has found that pyrrhotite can be formed from pyrite during metamorphism. During shearing and deformation accompanied by elevated temperatures capable of producing metamorphism, pyrite and, to a lesser extent, sphalerite tend to deform in a brittle fashion over a wide range of temperatures, confining pressures and strain, whereas chalcopyrite, pyrrhotite and galena deform plastically (Graf and Skinner, 1970; Stanton and Willey, 1970; Sangster, 1972). Under such conditions pyrite and sphalerite would be strung out as thin or irregular bands along the limbs of folds, whereas chalcopyrite, pyrrhotite, and galena would flow into axial zones. Fold hinges might thus contain ore bodies highly irregular in shape and enriched in sphalerite and galena.

Most prospecting for massive sulfides in the Chopawamsic Formation has been on the attenuated limbs of isoclinal folds, because it is on these limbs that the ore bodies generally crop out on the surface and there develop obvious gossans. Drilling of some sulfide deposits in the Andersonville quadrangle indicates that they are composed of several mineral-bearing horizons (Young, 1981). At current market prices, such massive sulfide deposits are too sporadic and too small to be of economic value and the discovery of larger ore bodies will be necessary to support mining in the area.

Because the massive sulfide bodies are deformed by one pre-Arvonia and three post-Arvonia fold events their original geometry has been



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Because the massive sulfide bodies are deformed by one pre-Arvonia and three post-Arvonia fold events their original geometry has been extremely altered. Stresses producing F_1 isoclines greatly transposed and attenuated S_{0a} beds. These stresses must have caused flattening and stretching of the stratabound massive sulfide deposits in the Chopawamsic along F_1 fold limbs. During this deformation chalcopyrite, pyrrhotite, and galena would tend to flow toward the axes of folds and the massive sulfide body would thicken in the hinge zones of these locally rootless F_1 folds similar to those shown in Figures 4, 5, and 6. It follows that the ore bodies in the Chopawamsic would be concentrated in hinges of both locally rootless F_1 folds and in F_2 folds (similar to, but on a larger scale than the fold defined by the amphibolite shown in Figure 7) and that intersecting hinges of super-imposed folds are prime areas for the occurrence of massive sulfide deposits.

This concept that massive sulfide bodies in the Chopawamsic Formation are thickened by plastic flowage into fold hinges can be tested by analysis of existing geologic and drill hole data. The more particular concept that the greatest thicknesses of these bodies occurs at the intersections of hinges of different fold systems is not readily tested from these data. The test data include a geologic map of the gossan and surrounding rocks prepared by Marr (1980b) and drill-core logs of Young (1981) for the massive sulfide body which Young designated as Zone 24, located in the extreme east-central part of the Andersonville quadrangle (Figure 13). The massive sulfide body

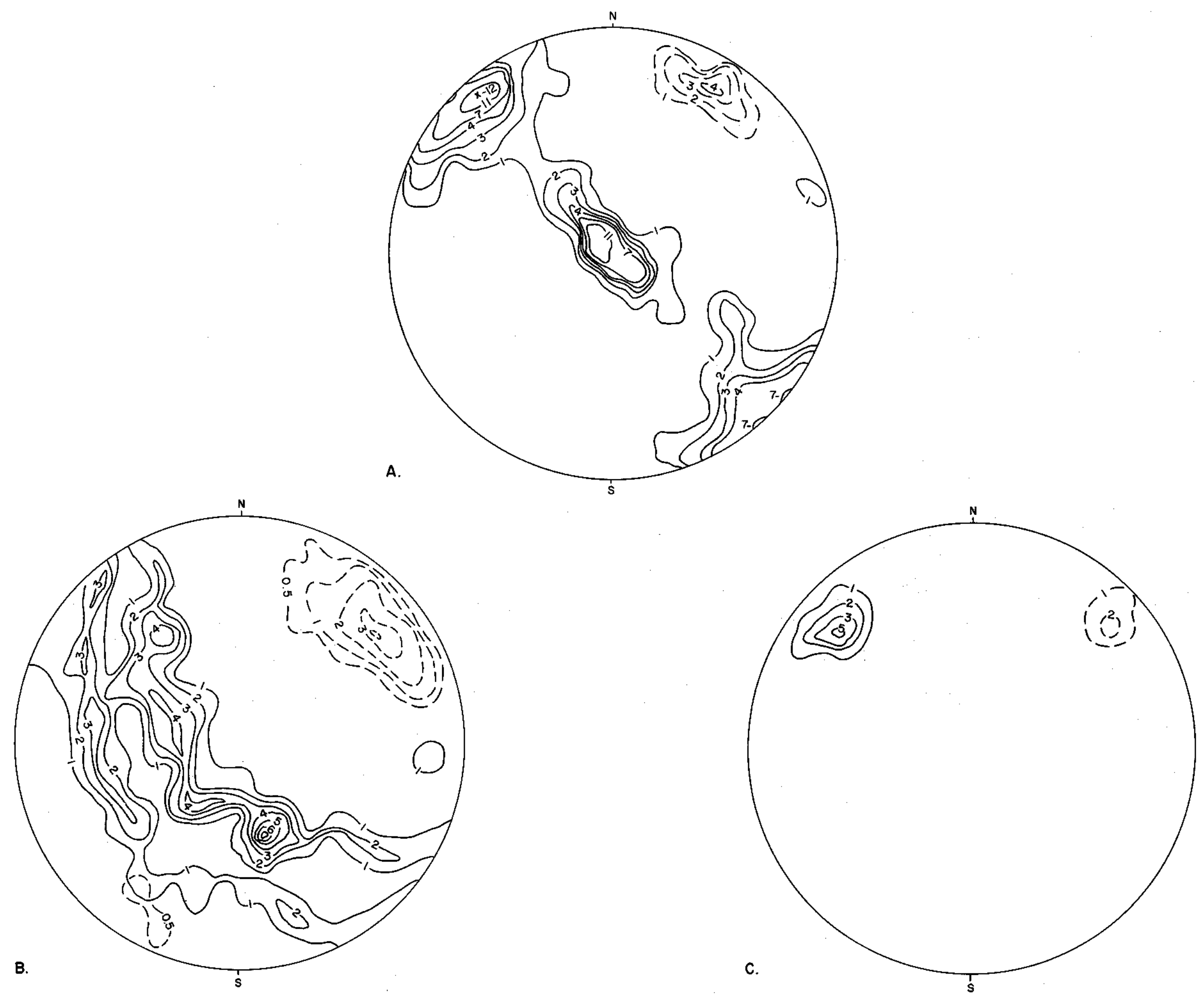

Figure 12. Equal area contour diagrams of poles to foliations and of lineations based on data from the Andersonville and Willis Mountain quadrangles. Poles to foliations are contoured in solid lines; lineations are contoured in dashed lines. Values are of contour lines in percent.

A. Chopawamsic Formation northwest of outcropping band of rocks of the Arvonia Formation. Poles to S_1 and L_1 lineations.

B. Chopawamsic Formation southeast of outcropping band of rocks of the Arvonia Formation. Poles to S_1 and L_1 lineations.

C. Arvonia Formation. Poles to S_2 and L_{2b} lineations.

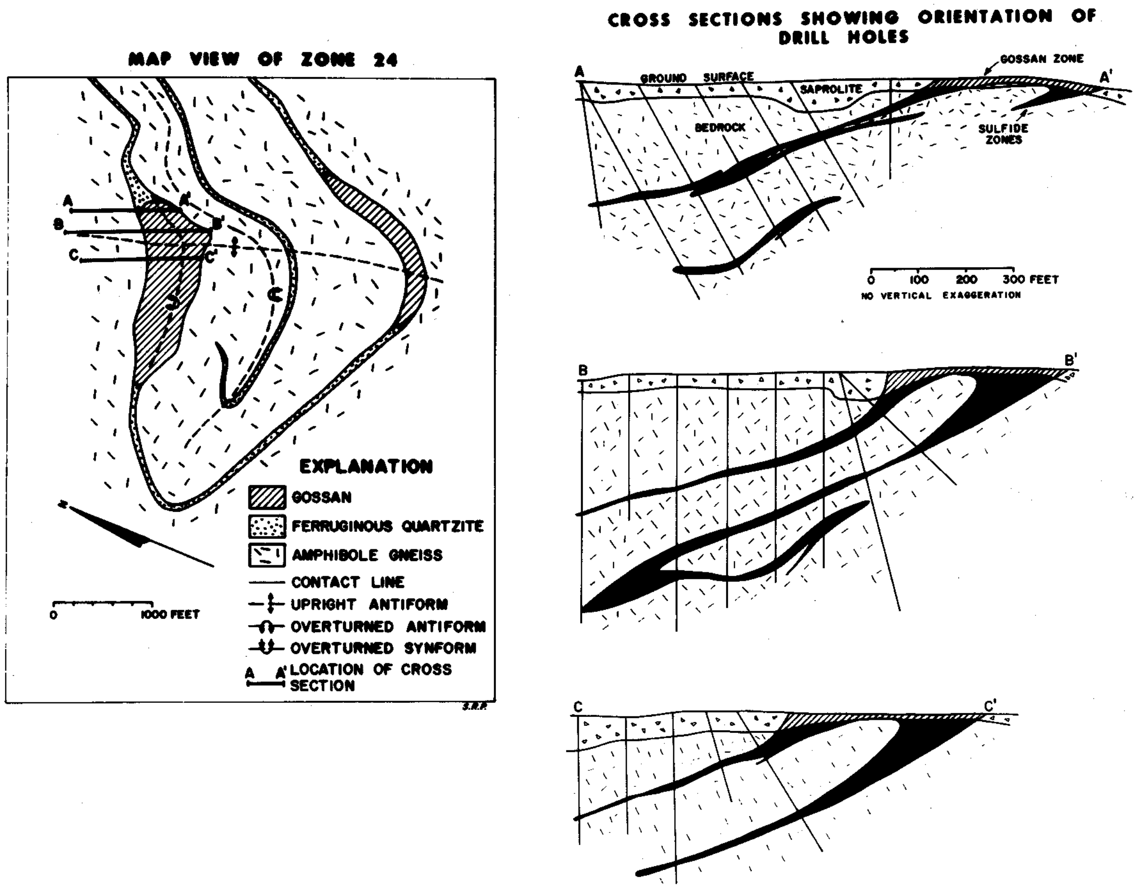

Figure 13. Geologic map and interpretive cross sections of massive sulfide zone 24. East-central part of the Andersonville quadrangle.

occurs in amphibole gneiss. The outcrop of this body is marked by the largest gossan zone in the quadrangle and is contained within the hinge zone of a fold. At Zone 24, Young (1981) found a fairly consistent stratigraphic sequence in the drilled area which indicated that it was in the crestal portion of a partially eroded antiform. The structure of the area includes an F_2 synform lying just southeast of the gossan which has been refolded by a northwest-trending F_4 fold, the axis of which can be traced approximately through the center of the gossan zone (Figure 13).

The writer's interpretation of the structure is based on geologic contacts mapped by Marr (1980b) and core data from Young (1981). The interpretation is shown in the cross sections A-A' through C-C' of Figure 13. In this illustration the core holes are superimposed on the cross sections.

Massive sulfide horizons shown in cross sections B-B' and C-C' clearly outline the limbs of an antiformal structure that is probably complementary to the F_2 synform recognized by Marr (1980b). The massive sulfide body is thickest in the hinge zone of the antiformal fold, which is exposed at the surface and produces and large gossan. Cross section A-A' is located on the northwestern edge of the gossan and shows two rootless F_1 folds, one on each limb of the antiform. In addition, extreme attenuation along the limbs of the antiform has proceeded to the extent that the lower limb is totally pulled apart. These cross sections support the idea that the massive sulfides are concentrated in the axes of folds and indicate that the understanding of fold geometry of the area is an important factor in the location of economic massive sulfide bodies.

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LITHOSTRATIGRAPHY AND BIOSTRATIGRAPHY OF THE MARTINSBURG FORMATION IN SOUTHWESTERN VIRGINIA, WITH DESCRIPTIVE SECTIONS

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ILLUSTRATIONS

ABSTRACT

The lithologies and fossils of the Martinsburg Formation have been examined at twenty-seven localities in southwestern Virginia. Sections at four of these localities are most completely exposed and served as control sections for the study. The Martinsburg (450 to 250 m) is comprised of upper and lower tongues of terrigenous clastic sediments, which are separated by a thin shaley limestone interval in easternmost exposure belts. In westernmost exposures the formation is predominantly carbonate rocks with some shale in the upper half. Sandstone is the dominant lithology in the top 50 meters in the eastern half of the study area. All these lithologies are characterized by distinctive thin-to-medium scale bedding and sedimentary structures believed to have resulted from contrasting storm and fair weather processes in a marine-shelf depositional environment.

The Martinsburg Formation is abundantly fossiliferous, including brachiopods, bryozoans, bivalve molluscs, and trilobites. However, the relative abundance of faunal elements is strongly related to lithologic facies. Biostratigraphic markers are rare, the most reliable is the disappearance of *Sowerbyella curdsvillensis*. Bento-

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nites near the base of the formation provide a reliable time-stratigraphic boundary.

INTRODUCTION

The Upper Ordovician Martinsburg Formation in southwestern Virginia is a richly fossiliferous unit that consists of well-bedded limestone, siltstone, and sandstone, interbedded with shale. The formation is exposed in the Appalachian Valley and Ridge province in a series of imbricate thrust sheets (Figure 1). This belt of rocks in which the Martinsburg exposures occur is bounded on the west by Upper Paleozoic rocks of the Appalachian Plateaus and is in fault contact with Precambrian metamorphic rocks of the Blue Ridge province to the east (Figure 1). Lithologies and fauna of the Martinsburg Formation in southwestern Virginia were described by the authors (Kreisa, 1980) (Springer, 1982). Because the formation is thick (300+m) and relatively nonresistent to weathering, it is nowhere completely exposed. Four nearly complete sections are suitably located within the study area (Figure 1) to serve as control sections. These four sections were measured and sampled in detail (Figure 1, Appendix I). Twenty-three partially exposed, intermediate sections were used to reconstruct facies relationships (Figures 1 and 2).

The authors are grateful to R. Sue Murley, Brian Cooper, and Francis Plants Whitehurst for field assistance. Our thanks, also, to Richard K. Bambach and J. Fred Read for field visits and many helpful discussions. We gratefully acknowledge partial financial support from the Virginia Division of Mineral Resources, the Department of Geology, VPI & SU., the American Association of Petroleum Geologists, Sigma Xi, and the Geological Society of America.

LITHOSTRATIGRAPHY

The Martinsburg Formation was named for exposures near Martinsburg, West Virginia (Geiger and Keith, 1891). These exposures lie in the Great Valley which extends from Virginia to New York along the eastern margin of the Valley and Ridge province. Although the formation name is applied in mapping throughout most of southwestern Virginia (Butts, 1940; Calver and Hobbs, 1963), the formation there is lithologically different from the Martinsburg in the type area. In the Great Valley, the Martinsburg Formation consists predominantly of a basal dark gray to

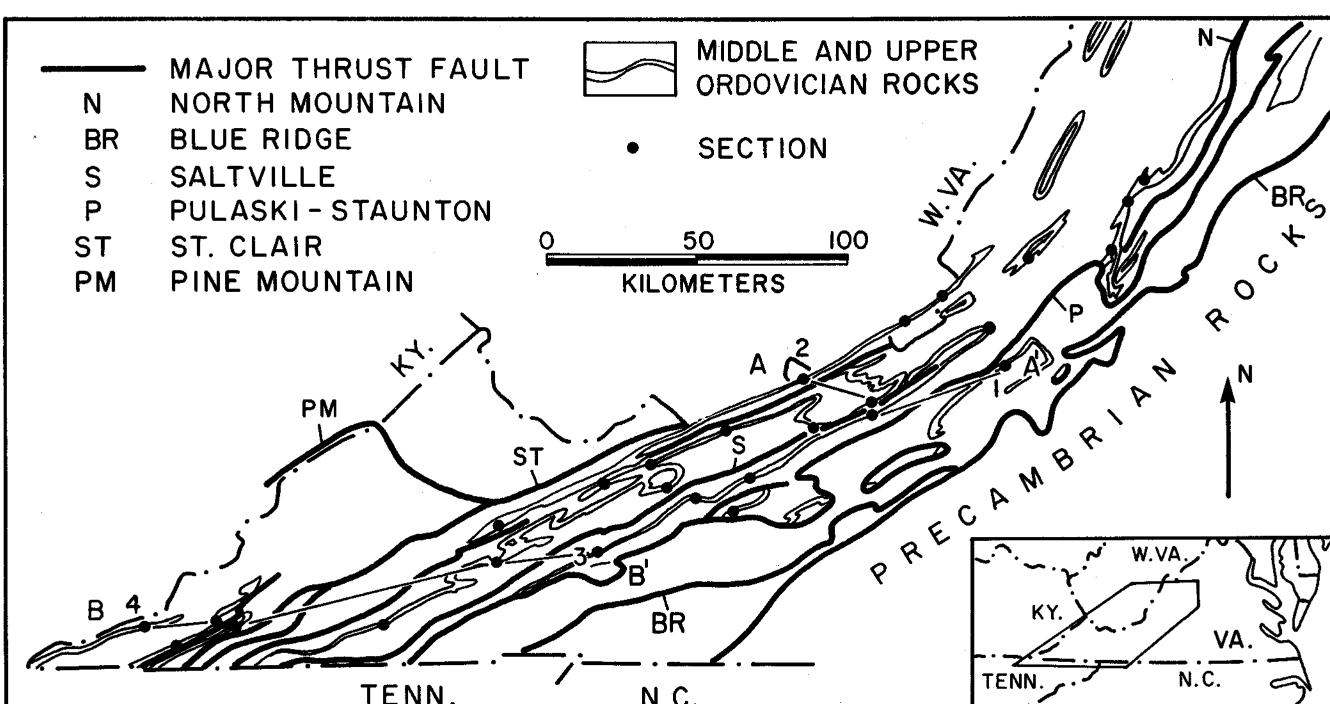

Figure 1. Generalized geologic and location map of the study area. Numbered sections are: 1 - Catawba Mountain; 2 - Narrows; 3 - Walker Mountain; 4 - Hagan (See Appendix I).

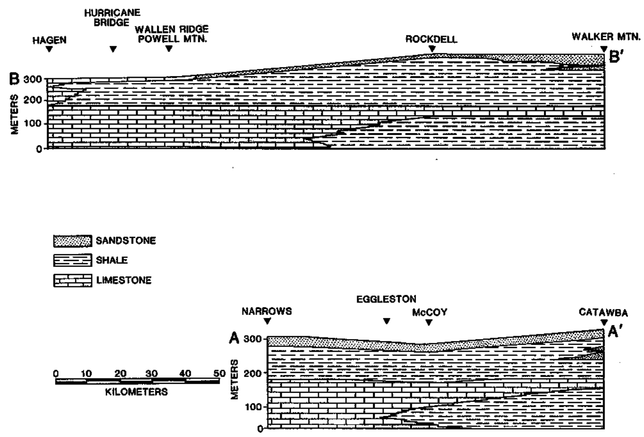

Figure 2. Lithologic cross sections, Martinsburg Formation and equivalents; palinspastic base. Patterns shown are portions of the section containing greater than 50 percent sandstone, shale, and limestone respectively. See Figure 1 for location of measured sections.

black shale and overlying interbedded graywacke and shale (flysch) with a meager fauna consisting mostly of graptolites (McBride, 1962; Rader and Henika, 1978). Stephens and Wright (1981) have described the lithologic and biostratigraphic relationships of the flysch facies of the Martinsburg Formation in the Great Valley of Pennsylvania. In southwestern Virginia and in exposures in the western anticlines of west central Virginia (Rader and Ryan, 1965) the Martinsburg Formation is composed of skeletal, intraclast and peloidal limestone, siltstone and sandstone, interbedded with shale. The formation contains an abundant benthic marine fauna (Walker, 1967, 1970; Plants, 1977; Bambach, Kreisa and Whitehurst, 1978). In Lee County, Virginia, equivalent lithologies contain less terrigenous sediment and the "Trenton" Formation and "Reedsville" Shale are considered Martinsburg equivalents (Butts, 1940; Miller and Brosgé, 1954; Calver and Hobbs, 1963).

The Martinsburg Formation in southwestern Virginia has commonly been divided into "Trenton", "Eden", and "Maysville" "members" based largely on fossil taxa (Butts, 1940; Cooper, 1944; Woodward, 1932; Rader and Ryan, 1965). Walker (1967, 1970) departed from this practice and defined three lithologic "facies" (Walker, 1970, p. 216-217). A similar three-fold lithologic subdivision of the Martinsburg and its southwestern equivalents has been recognized in this study, though facies descriptions and distributions differ somewhat from Walker's. The Formation has been divided into three lithofacies which contain more than 50 percent limestone, shale, or sandstone, respectively (Figure 2). The limestone classification used in this study is that of Dunham, 1962.

The lower shale tongue (Figure 2) contains a few thin beds of very fine-grained sandstone and siltstone near the base of sections in the eastern part of the area, and thin and very thin beds of skeletal packstone, and terrigenous, silty peloidal limestone throughout. Limestone becomes more abundant through the upper half of the lower shale tongue and undergoes a very gradational transition to the overlying limestone-dominated facies. To the west, very thin medium-bedded skeletal, intraclast and peloidal limestone dominates the entire lower portion of the section (Figure 2). This lateral transition from the lower shale tongue is also gradual. In the far western part of the area, only 5 to 15 percent of the lower strata ("Trenton" Formation) is shale.

Transition to the laterally extensive upper shale-dominated wedge (Figure 2) is sharp, occurring over less than 10 meters vertically and commonly in only one to two meters. The upper shale wedge contains thin and very thin beds of skeletal and peloidal limestone and calcareous siltstone at the base. Fine- and very fine-grained sandstone gradually increases in abundance upward, becoming the most abundant lithology as thin and thick beds in the upper portion of the Martinsburg in the western and central parts of the study area (Figure 2).

The Martinsburg Formation and its equivalents in southwestern Virginia conformably overlie a variety of gray limestones (Eggleston Formation) in the western and northern part of the study area, and red mudrock and sandstone facies (Bays and Moccasin formations) in the central, southern and eastern parts of the area (Hergenroder, 1966; Kreisa, 1980; Read, 1980). The base of the formation is placed usually at the lowest fossiliferous skeletal packstone. This lithologic marker is sharp. Non-fossiliferous and sparsely-fossiliferous beds similar to those which occur below it never recur above. In addition, it falls near the top of the zone of abundant bentonites ("cuneiform group" of Rosenkrans, 1936) in all measured sections. More precisely, throughout most of the region, it typically occurs three to ten meters above bentonite "V-7" of Rosenkrans (1936); Cooper, (1944, p. 99); Miller and Fuller, (1954, p. 131); Hergenroder, (1966, Plate 12);

Kreisa, (1980). The bentonite horizon, always close to the lower contact is probably a reliable time marker and is used as the stratigraphic datum for cross sections in this paper.

The top of the formation is placed below the first abundant reddish-gray sandstone or mudrock of the overlying Juniata or Sequatchie formations. Throughout most of southwestern Virginia the Martinsburg Formation is overlain by the Juniata Formation. The Martinsburg-Juniata contact occurs at or near the top of a 10- to 15-meter zone of highly bioturbated lithologies containing abundant *Lingula* and *Orthorhynchula linneyi*. This contact is probably time-transgressive, becoming younger to the west and southwest. Locally along the southeastern margin of the study area (for example, Catawba Mountain, Figure 1, Section 1) the Juniata Formation has been removed by late Ordovician erosion and the Martinsburg is unconformably overlain by Silurian sandstones. In Lee County at the far southwestern tip of Virginia, the Martinsburg ("Reedsville") Formation contains abundant limestone near the top (Appendix I, Hagan Section) and *Lingula* and *Orthorhyncula* are rare or absent. Here the Martinsburg grades upward to red and greenish-gray calcareous shale, mudstone and limestone of the Sequatchie Formation.

The regional thickness pattern of the Martinsburg Formation (Figure 3) is not well known due to structural deformation and poor exposure. Thickness of the formation is especially difficult to determine on the North Mountain fault block and the isopachous contours of Figure 3 are generalized in that area. In southwestern Virginia the Martinsburg thickens gradually from approximately 300 meters at Virginia's southwestern tip, to approximately 440 meters eastward and northeastward near Roanoke, and then thickens markedly into the flysch facies in the Great Valley.

Regardless of the relative abundance of sandstone, shale, or limestone present in the Martinsburg Formation, these lithologies nearly always occur in distinctly segregated beds (Figure 4). This bedding and associated sedimentary structures are described and interpreted in detail elsewhere (Kreisa, 1981), and are treated only briefly here. Individual beds commonly occur in fining-upward sequences (Figures 4 and 5) ranging from less than one to more than 80 cm in thickness, which typically have a sharp or scoured base and a gradational or burrowed upper bed surface. Where they are complete, fining-upward sequences consist of a couplet of whole-fossil

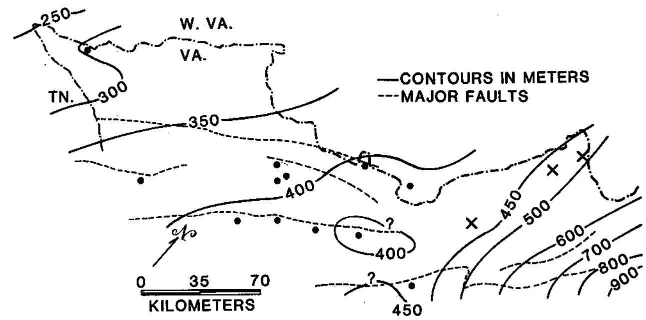

Figure 3. Thickness of Martinsburg Formation and equivalents; palinspastic base. Sources: this report (black dots); Rader and Ryan, 1965, (X's); published maps and reports.

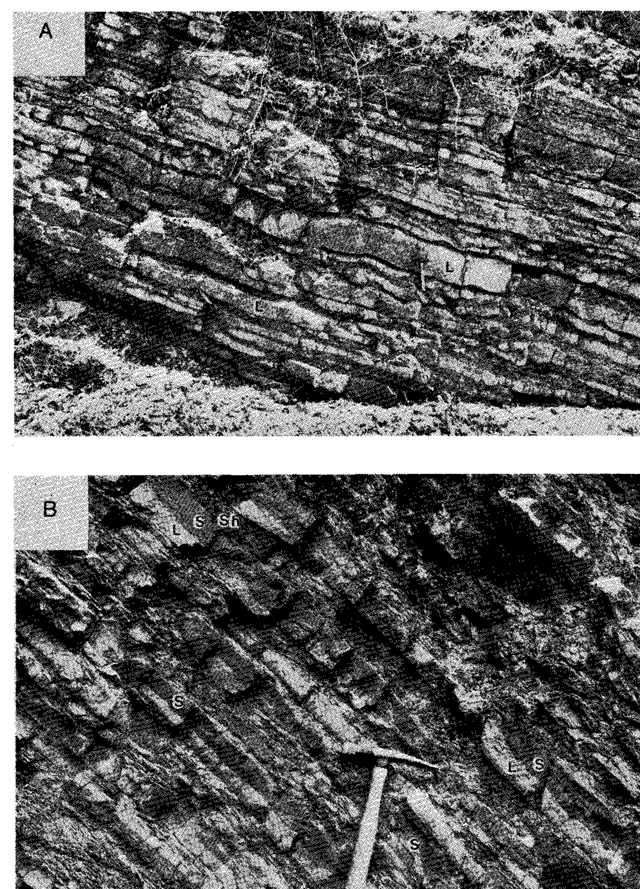

Figure 4. Bedding and sedimentary structures in the Martinsburg Formation.
A) Interbedded shale and limestone (L), U. S. Highway 460 at Narrows. Note distinct segregation of lithologies.
B) Interbedded shale (sh) and very fine-grained sandstone (s), Narrows. Note that several sandstone beds are fining-upward sequences of basal sandy limestone (L), sandstone, and shale.

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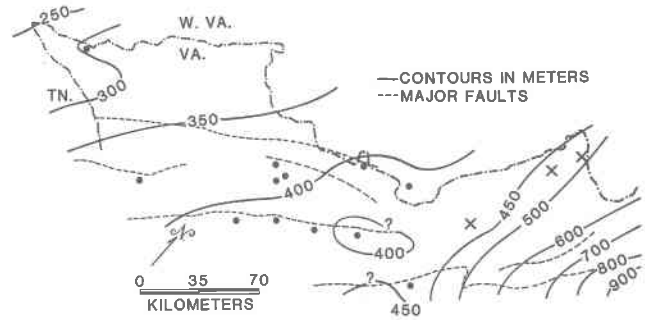

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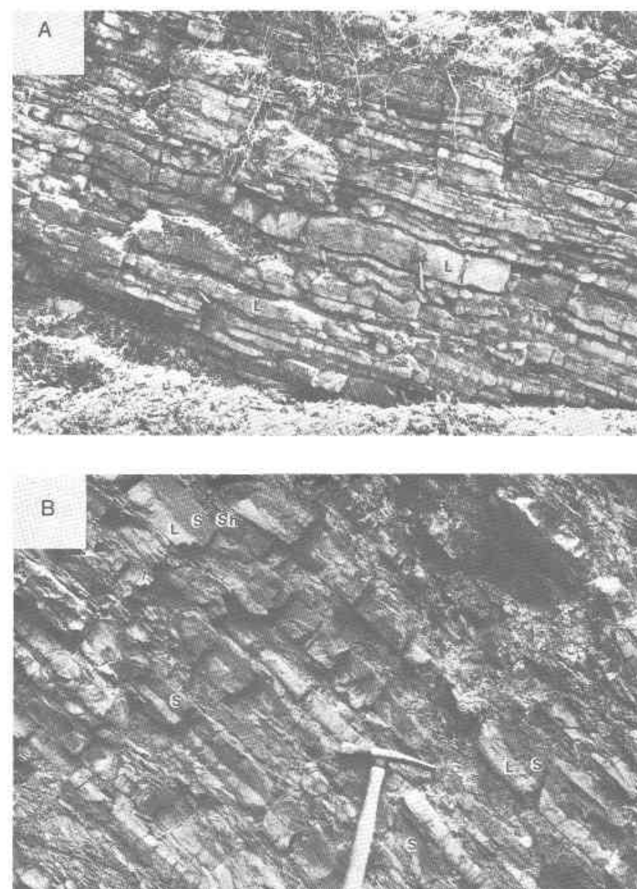

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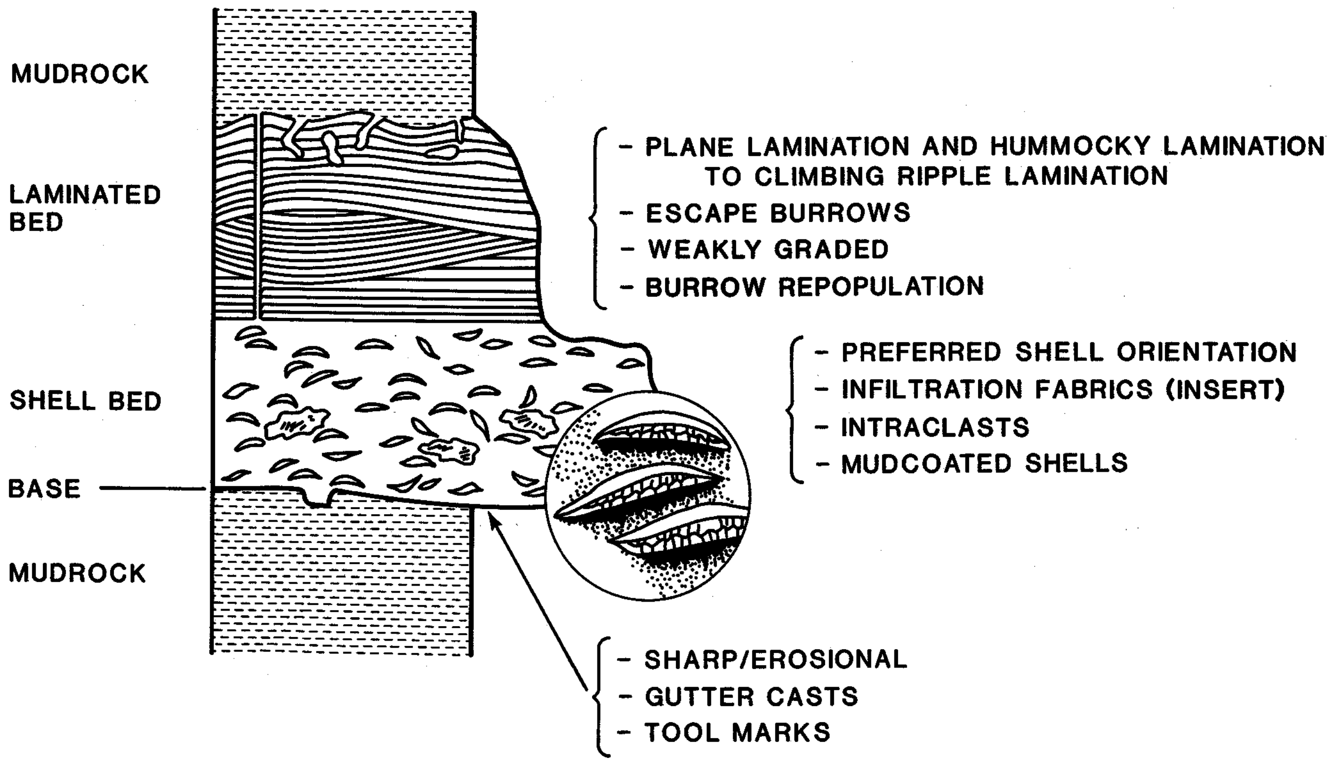


Figure 5. Idealized sequence of lithologies and sedimentary structures from the Martinsburg Formation.

packstone and laminated sandstone, siltstone, or carbonate, which in turn grades up to shale (Figures 4, 5, and 6). This bedding style is interpreted to have formed because of contrasting storm and fair-weather processes. Major storms that occurred over the Martinsburg marine shelf produced long period waves, which disturbed the bottom sediment well below "normal" wave-base. Coarse sediment, such as shells and ripped-up intraclasts, formed winnowed lag deposits whereas sand and mud sized sediments were suspended by the wave energy and deposited in laminated beds and shale layers when the storms waned. Much of the shale probably also represents normal hemipelagic deposition. Sedimentary structures such as "hummocky cross-stratification" (Figure 6) (Harms and others, 1975) and "climbing wave-ripple lamination" (Kreisa, 1981) are believed to be important clues to the origin of storm-generated bedding. These sedimentary structures are abundant in the Martinsburg Formation and their occurrence is noted in the descriptive sections (Appendix I). Although only between 10 and 40 percent of the beds occur as complete couplets of packstone-laminated sandstone, most beds are interpreted to have formed in a similar way. Only in the upper 10 to 15 meters of the Martinsburg Formation is such bedding largely absent. Here it has been destroyed by intense bioturbation.

Most measured sections of the Martinsburg Formation contain several zones (up to seven) that have been influenced by soft sediment deformation (see Appendix I). These zones range in thickness from a few decimeters to 8 meters and are most common in the sand-dominated facies near the top of the formation. Some zones consist of ball-and-pillow structures and slightly deformed bedding within which individual disjointed or contorted beds can be traced across the outcrop. Others are comprised of scattered, chaotically arranged blocks of sandstone, sandy packstone, and other lithologies imbedded in zones of massive, matrix-rich, poorly-sorted fine-grained sandstone. These deformational features are attributable to loss of strength within the sediment not long after it was deposited (Kreisa, 1980). Liquefaction, resulting from increased pore-fluid pressure resulting in a temporary fluid-like mixture of sand, clay and water is probably responsible for the massive, matrix-rich sandstones. Similar soft sediment deformational phenomena

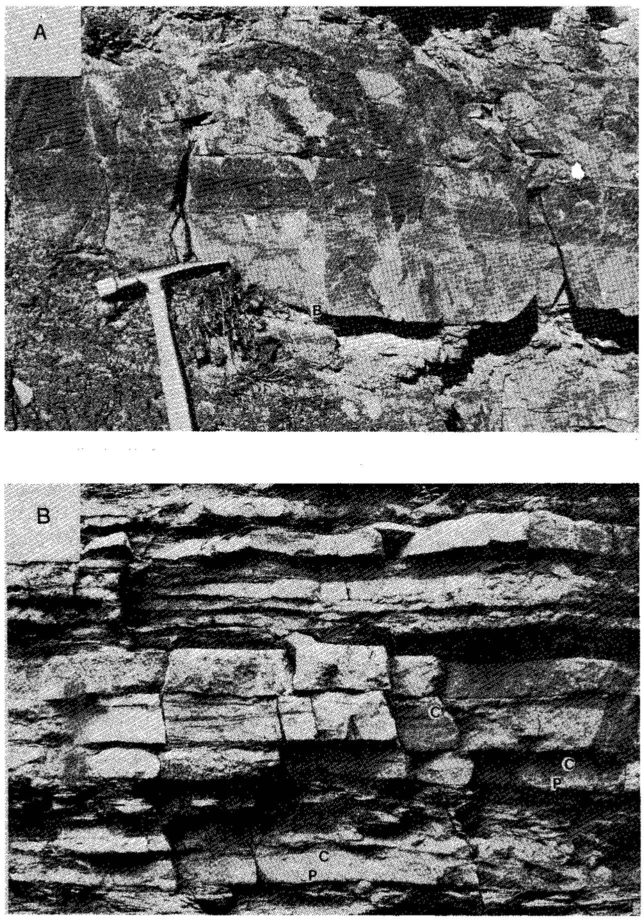


Figure 6. Sedimentary structures in the Martinsburg Formation.
A) Hummocky cross-stratified sandstone, U. S. Highway 311 at Peters Mountain Virginia - West Virginia. Note also irregular scoured base (B).
B) Interbedded limestones and shale, Narrows. Note hummocky cross-stratified limestone (calcarenite) near scale (15 cm long), scoured basal bed contacts, and beds which are couplets of packstone (P) and calcarenite (c).

have been documented to result from seismic shock and from cyclic loading associated with the passage of large ocean-waves. Martinsburg soft-sediment deformation zones show little evidence of lateral movement (i.e. slumping; Appendix I; Kreisa, 1980).

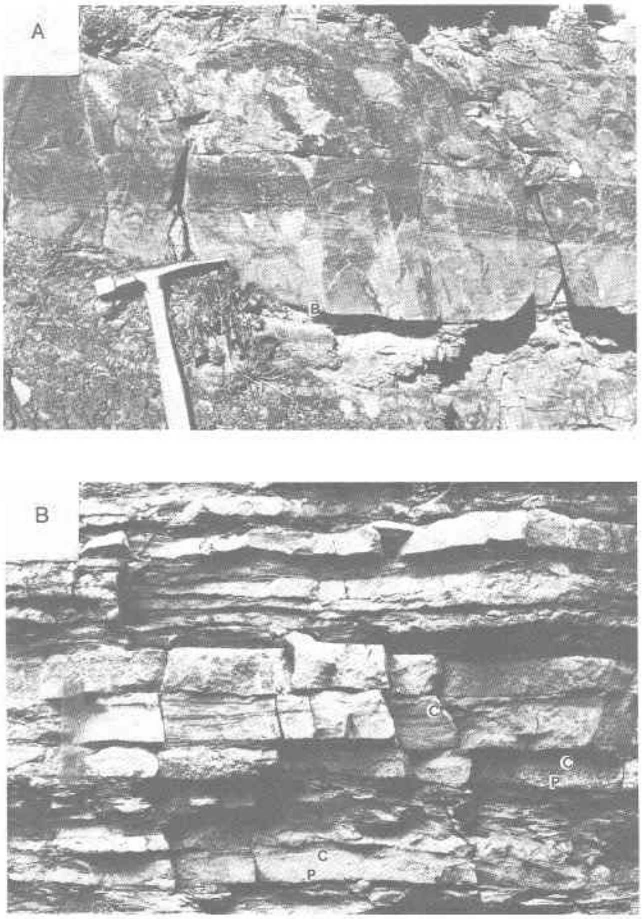
The composition of Martinsburg Formation lithologies has been documented using thin sections, insoluble residues, and X-ray diffraction (Kreisa, 1980). Martinsburg sandstones are lithic and sublithic arenites (Appendix II). They contain an average of 17 percent chloritic and illitic matrix, much of which has apparently formed through diagenetic crushing of micaceous rock fragments. Carbonate and silica cements are subequally abundant, comprising a total of about 20 percent of the rock volume. Discounting minor constituents, cement, and matrix, Martinsburg sandstones average 66 percent quartz, 26 percent micaceous rock fragments and 8 percent feldspar. Based on clay minerals identified by X-ray diffraction for five shale samples, the shales of the Martinsburg Formation consist dominantly of white mica (35 to 50 percent), kaolinite (up to 30), and vermiculite (up to 45), with lesser amounts of montmorillonite, chlorite, and mixed-layer chlorite/vermiculite (up to 15 of each). Both the sandstones and the shales are typically light- to medium gray and light-olive gray.

Limestones of the Martinsburg Formation include abundant skeletal packstones, laminated peloidal calcarenites and calcisiltites, and lime mudstone. Coarse, whole-fossil packstones are abundant and contain whole and broken brachiopods, molluscs, echinoderms, trilobites, and bryozoans, in addition to lime mud (micrite) (Appendix II). Sand and silt-sized carbonates typically include abundant peloids and fine-grained skeletal fragments.

BIOSTRATIGRAPHY/BIOFACIES

The Martinsburg Formation fauna is moderately diverse, composed primarily of strophomenid and orthid brachiopods, trepostome bryozoans, bivalve molluscs, and isotelid trilobites. Rhynchonellid brachiopods, pleurotomariacean and bellerophontid gastropods, crinoids, and nautiloid cephalopods are locally important. A listing of common Martinsburg species identified in this and other studies is found in Appendix III. Preservation commonly occurs as internal, external or composite molds. Original shell material is preserved in many brachiopod and bryozoan specimens.

Biostratigraphic markers are rare in the Martinsburg. Rust (1968) examined the sparse conodont fauna and correlated the Martinsburg Formation with Trenton and younger rocks in New York State. Walker (1967) placed the Trentonian-Edenian boundary with the Rafinesquina alternata/Dalmanella bassleri zone (Martinsburg species identified as *Dalmanella* are now considered synonymous with *Onniella*, G. Chandlee, personal communication, 1981). This zone occurs between 335 and 366 meters above the base of the section at Catawba and Walker mountains, between 274



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The disappearance of *Sowerbyella curdsvillensis*, the larger of two species of *Sowerbyella* found in the Martinsburg, probably represents the most reliable biostratigraphic marker in the formation. *S. curdsvillensis* disappears between 45 and 60 meters above the base of the sections at Hagan, Walker Mountain, and Narrows, and is replaced by the smaller species, *S. rugosa*. The species change is not directly observable at Catawba Mountain, where exposures are poor or nonexistent between 9 and 73 meters above the base. *S. curdsvillensis* was collected immediately below the covered interval. *S. rugosa* was collected immediately above the covered interval, indicating that the species change-over occurs within that 60-meter portion of the section, a stratigraphic position not inconsistent with that in the other three sections. (Ranges of major Martinsburg taxa are recorded in Figure 7).

The *Orthorhynchula* zone (Bassler, 1919), approximately the upper 30 meters of Martinsburg and Reedsville formations in southwestern Virginia, has been considered a local range zone (hence, of biostratigraphic significance) (Butts, 1940) and a facies-controlled faunal zone (Butts, 1945; Woodward, 1951; Walker, 1967). Species from this zone include abundant nuculanacean, modiomorphid, and pterioid bivalves, pleurotomariacean and bellerophontid gastropods, and linguloid brachiopods, as well as *O. linneyi*, the large rhynchonellid brachiopod for which the zone is named. The *Orthorhynchula* zone fauna occur predominantly in fine- to very fine-grained calcareous sandstones within the study area. The fauna has been found as well in argillaceous limestones, thick-bedded limestones, and siltstones near the top of the Reedsville and Martinsburg formations through the Central Appalachians (Bretsky, 1970).

Rocks in which *Orthorhynchula* and its associated fauna occur contain abundant sedimentologic and strataigraphic evidence of deposition in near-shore environments. The *Orthorhynchula* zone represents a fauna tolerant of high stress conditions, particularly salinity fluctuations and bottom instability, which may be preserved in a number of different lithologies. (Springer and Bambach, 1980; Springer, 1982).

The *Orthorhynchula* zone is diachronous (younger to the west), reflecting shifting shoreline position during final filling of the late Ordovician Martinsburg depositional basin (Kreisa, 1980, 1981). It is therefore a local range zone, but cannot be considered an accurate time-line.

GEOLOGIC HISTORY

The Martinsburg Formation and its equivalents in southwestern Virginia were deposited on an open-marine platform as evidenced by the abundant shelly fauna found in these strata. Limestones and red beds of the underlying Eggleston, Moccasin, and Bays formations were deposited in a variety of peritidal and deltaic environments (Hergenroder, 1966; Kreisa, 1980; Read, 1980). Open-marine conditions of the Martinsburg Formation were established during an initial transgression that drowned these underlying environments. Bentonite stratigraphy suggests that the transgression was nearly synchronous throughout the study area. The marine platform extended the entire width of present exposure belts in southwestern Virginia, southward into eastern Tennessee (Rodgers, 1953), and north of the study area through west-central Virginia (Rader and Ryan, 1965) at least as far as central Pennsylvania (Thompson, 1972). The platform to the north lay northwest of a flysch basin that extended along the eastern margin of the Valley and Ridge province from central Virginia to southern New York (McBride, 1962; Rader and Henika, 1978). The precise geometry of the platform in southwestern Virginia is not clear due to limited exposure, but it appears that after initial transgression the platform sloped gently to the northeast and was thus a ramp as defined by Ahr (1973).

Fine terrigenous clastic sediments deposited on the southeastern part of the Martinsburg platform during early Martinsburg time (the lower shale tongue, Figure 2) were shed from tectonic uplands to the southeast. The uplands were associated with the waning stages of an early phase of the Taconic Orogeny (the Blountian phase; Rodgers, 1953, 1971; Kreisa, 1980). Terrigenous influx gradually slackened, giving way to dominantly carbonate deposition across most of the platform (Figure 2) during mid-Martinsburg time.

The change from clastic to carbonate-dominated facies is accompanied by shifts in relative abundance of several faunal elements. *Onniella sp. 1* and *R. alternata* decrease in relative abundance; *S. curdsvillensis* disappears, to be replaced by *S. rugosa* in the carbonate-rich tongue near the middle of the Martinsburg (Figure 2).

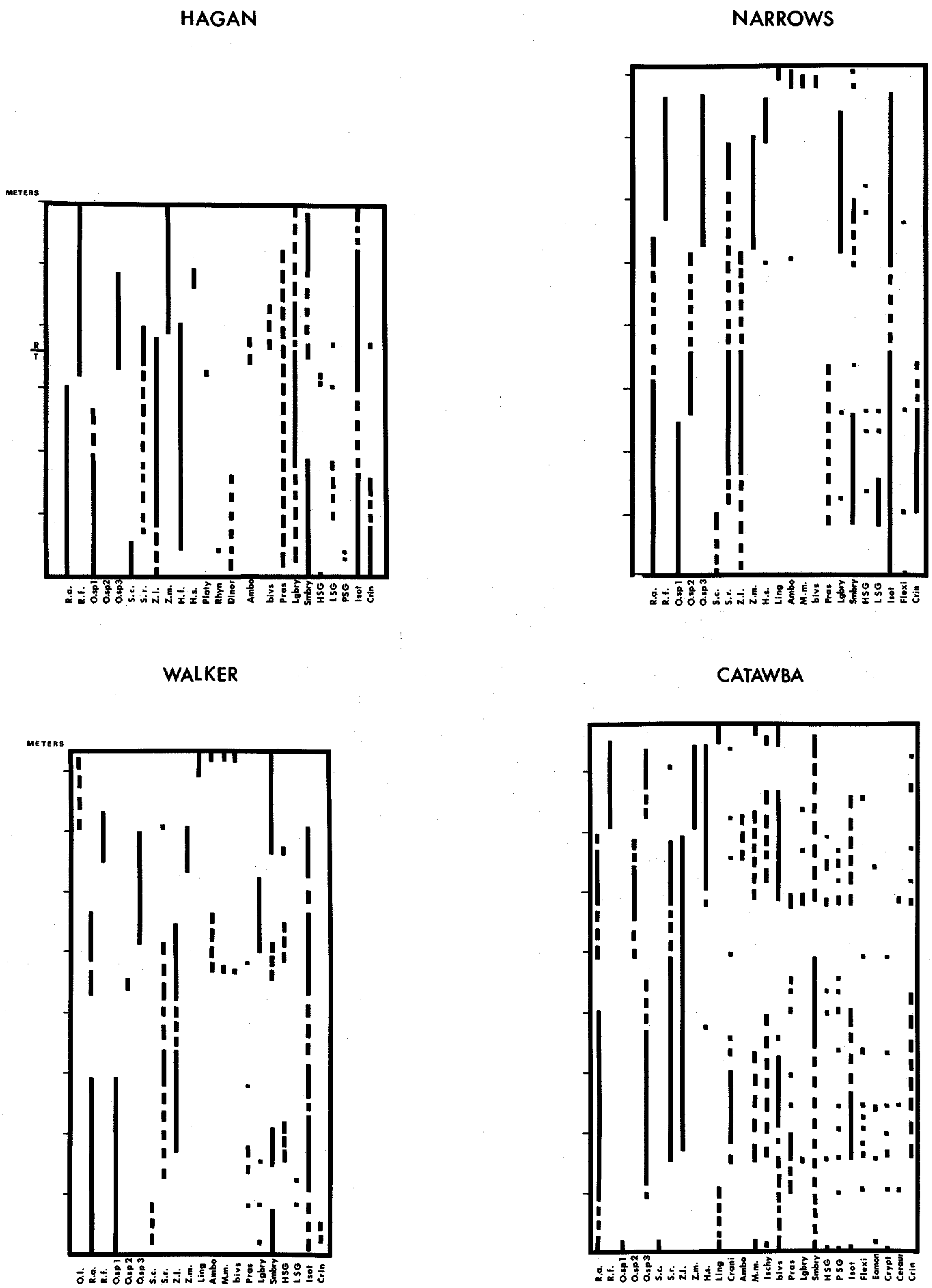

Figure 7. Ranges of important Martinsburg taxa at Hagan, Walker Mountain, Narrows, and Catawba Mountain sections, southwestern Virginia. Refer to Appendix III for taxa.

The upper half of the Martinsburg Formation reflects renewed influx of terrigenous clastic sediments accompanying the main or "Hudson Valley" (Rodgers, 1971) phase of the Taconic orogeny in the southeastern source area. This sequence of strata coarsens upwards and fine-grained terrigenous sediments spread farther to the west than during deposition of the lower half of the formation (Figure 2). The fauna of the upper portion of the Martinsburg reflects the return to the clastic-dominated depositional regime. *Onniella* and *Rafinesquina* again become abundant; *S. regosa*, unable to tolerate the increase in frequency of disturbance associated with final filling of the basin, ceases to be an important member of the fauna.

The intensely burrowed beds at the top of the Martinsburg Formation, which contain abundant *Orthorhynchula linneyi*, *Lingula*, and clams, were deposited in very nearshore, shallow water environments, at or just below the lower limit of fair-weather wave base. This burrowed horizon is directly overlain by beach and deltaic facies of the Juniata Formation (Kreisa, 1980, 1982) and is very similar to highly bioturbated zones found at the transition between the offshore and shoreface environments in modern depositional settings (Howard and Reineck, 1972; Reineck and Singh, 1975). Little terrigenous sand was transported as far as Lee County, either during the time of deposition of the upper Martinsburg Formation or the time of deposition of the Sequatchie Formation. There and farther to the southwest in Tennessee the marine platform gradually shoaled to sea level and red mudstone and carbonate of the overlying Sequatchie Formation were deposited largely on peritidal mud flats (Kreisa, 1980; Milici and Wedow, 1977).

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APPENDIX I: STRATIGRAPHIC SECTIONS

Descriptions of four lithologic sections which are representative of the Martinsburg Formation in southwestern Virginia are presented here. Sections were measured with Jacob staff, except where noted for certain covered intervals. Grain size for sandstones was determined in the field by comparison with a grain-size chart distributed by the Geologic Specialty Company, and colors of all units are from the Geological Society of America Rock Color Chart (Goddard and others, 1951). Bedding thickness terms are those defined by Ingram (1954) and limestones are classified according to Dunham (1962). In addition to the descriptions presented here, details of sedimentary structures and the relative abundances of lithologies were documented with thirty sequences, from one to seven meters thick, which were measured and described in centimeter-by-centimeter detailed (Kreisa, 1980). "HCS" and CWRL" refer to the sedimentary structures "Hummocky Cross-Stratification" and "Climbing Wave-Ripple Lamination" respectively (see text).

Catawba Mountain Section: Measured along U.S. Highway 311, on the northwest slope of the mountain; beginning at the crest of the mountain approximately 1.5 km east of the Catawba Post Office.

Tuscarora Formation-Not measured

Juniata Formation-Absent, probable unconformity

Martinsburg Formation-Total thickness 440.13 meters

UNIT		THICKNESS *(meters)*	CUMULATIVE THICKNESS
1	Sandstone, fine to very fine grained, grayish red and some light olive gray, thin bedded, but bedding largely destroyed by burrowing; phosphatic nodules. In this section, phosphatic nodules (0.2 to 2 cm) are generally closely associated with *Lingula* shells.	2.44	2.44
2	Sandstone, fine to very fine grained, light olive gray, thin to medium bedded, but largely destroyed by burrowing; phosphatic nodules.	4.85	7.29
3	Not exposed	1.58	8.87
4	Sandstone and minor shale; sandstone very fine grained, very thin bedded, highly burrowed; phosphatic nodules.	6.13	15.00
5	Sandstone and minor shale; sandstone thin to thick bedded, HCS, CWRL, plane laminated; many beds have shale intraclasts and shells concentrated at their base; beds variable in thickness laterally.	7.50	22.50
6	Sandstone and mudstone; beds disrupted by soft sediment deformation; sandstone as isolated, rotated ball-and-pilow-structure, 1 to a few cm long (load-cast ripple cross-lamination); medium light gray.	0.21	22.71
7	Highly fossiliferous sandstone.	0.21	22.92
8	Shale and siltstone, medium gray, laminated to thin bedded.	1.58	24.50
9	Poorly exposed and deeply weathered, with strong cleavage; possible fault.	2.59	27.09
10	Sandstone (45%), shale (45%) and sandy skeletal packstone; sandstone very fine grained to siltstone, medium light gray, very thin and thin bedded, plane laminated, CWRL; shale very thin to thin bedded, medium gray; skeletal packstone with abundant interstitial sand, to fossiliferous sandstone; beds have sharp erosional bases; packstone/laminated bed couplets common.	5.00	32.09
11	Poorly exposed, unit apparently consists of sandstone and shale; sandstone very fine grained to siltstone, medium light gray, thin to very thin bedded, CWRL; beds variable in thickness laterally; fossil concentrations at base of several beds.	3.39	35.48

Appendix I (cont.).

UNIT		THICKNESS (meters)	CUMULATIVE THICKNESS
12	Sandstone, very fine grained, medium light gray, disrupted by soft sediment deformation: ball - and - pillow structures near base, massive homogeneous in upper portion.	1.71	37.19
13	Not exposed	13.84	51.03
14	Shale (75%), sandstone (20%) and skeletal packstone (5%); sandstone very fined grained to siltstone, medium gray to medium light gray, thin to very thin bedded, sedimentary structures same as unit 10-11; skeletal packstone sandy.	7.13	58.16
15	Not exposed	2.53	60.69
16	Shale (70%), sandstone (20%) and skeletal packstone (10%); shale locally bioturbated, locally sandy; sandstone very fine grained to fine grained, very thin to medium bedded, sharp planar to undulatory bases, plane laminated, CWRL, HCS; skeletal packstone with sharp erosional bases, very thin to medium bedded, sandy.	6.95	67.64
17	Shale (80%) and sandstone (20%); sandstone very fine grained to siltstone, thin bedded, CWRL, but bedding generally not present due to intense bioturbations.	14.23	81.87
18	Sandstone, fine grained, medium gray, load casts.	0.67	82.54
19	Sandstone, shale, and minor skeletal packstone; sandstone very fine grained, thin to medium bedded with sharp erosional bases, CWRL. HCS, plane laminated, some beds variable in thickness, thin beds commonly lenticular; skeletal packstone sandy, commonly filling shallow scours at base of laminated sandstone beds.	15.36	97.90
20	Muddy very fine-grained sandstone, massive and homogeneous except for isolated, rotated blocks of laminted sandstone; due to soft sediment deformation (liquefaction).	3.05	100.95
21	Sandstone (50%), shale (45%), and skeletal packstone (5%); very fine to fine grained, medium light gray, thin to thick bedded (up to 55 cm) with some lenticular, very thin-bedded; plane laminated, CWRL, HCS, wave ripple cross - laminated, sharp erosional bases, shale interclasts locally at base; shale locally highly burrowed; skeletal packstone sandy, commonly filling shallow scours at base of sandstone beds.	6.37	107.32
22	Sandstone, fine to very fine grained, with abundant clay matrix; homogeneous and massive due to soft sediment deformation (liquefaction); mulluscs abundant and well preserved, apparently uniformly distributed throughout; rotated at all angles and "floating" within the unit; sandstone blocks form 5 cm to meter long, 0.5 meter thick.	8.06	115.38
23	Sandstone (55%), shale (35%), and skeletal packstone; sandstone fine grained to siltstone, medium light gry, very thin to thick bedded, plane laminated, CWRL, HCS, very thin beds lenticular, others variable in thickness; skeletal packstone generally sandy, commonly at base of laminated sandstone beds; scoured bases with shale intraclasts common.	11.95	127.33
24	Laminated sandstone and skeletal packstone in the form of ball-and-pillow structures and rotated beds in a matrix-rich sandstone (due to soft sediment deformation).	0.91	128.24
25	Shale (70%), sandstone (20%) and skele packstone (10%); shale sandy, in beds up to 90 cm thick, burrowed locally; sandstone very fine grained to siltstone, thin bedded and some medium bedded, plane laminated, CWRL; packstone commonly at base of laminated sandstone beds.	9.11	137.35

Appendix I (cont.).

UNIT		THICKNESS (*meters*)	CUMULATIVE THICKNESS
26	Shale (70%), limestone (20%) and siltstone (10%); limestone is skeletal packstone, plus thin skeletal accumulation several shell layers thick (0.5 cm), and a few calcisiltities, very thin bedded; other lithologies and sedimentary structures same as unit 25.	9.48	146.83
27	Silty, calcareous mudstone with abundant whole fossils, homogeneous due to soft sediment deformation; with rare rotated beds of skeletal limestone (15 cm); grades laterally to interbedded shale, limestone and siltstone similar to unit 26.	0.91	147.74
28	Shale (75%), limestone (20%), and siltstone (5%), shale moderately bioturbated; limestone is skeletal packstone, thin and very thin bedded, 0.5 cm thick skeletal accumulations, and laminated calcisiltite, silty; siltstone to very fine-grained sandstone calcareous, very thin bedded.	5.91	153.65
29	Not exposed	4.48	158.13
30	Shale and limestone; limestone mostly skeletal packstone in thin beds.	1.13	159.26
31	Not exposed, float suggest shale, thin siltstone beds, thin skeletal packstones.	40.54	199.80
32	Limestone (80%) and shale (20%); limestone is skeletal packstone, laminated silty calcisiltite, and argillaceous, peloidal lime mudstone; colors medium dark gray; bedding thin and very thin; calcisiltites plane laminated, CWRL; packstone-calcisiltite/lime-mudstone couplet common; beds have scoured bases and intraclasts.	26.24	226.04
33	Not exposed		
34	Same as unit 32.	4.63	230.67
NOTE:	Units 35 through 38 measured along old U.S. Highway 311 which winds through pasture north of present highway.		
35	Limestone (60%) and shale (40%); very thin to medium bedded, mostly thin bedded; lithologies and sedimentary structures same as unit 32.	21.34	252.01
36	Limestone (70%) and shale (30%); same as unit 35.	6.74	258.75
37	Shale (60%) and limestone (40%); shale is weakly fissile, calcareous; limestone is mostly lime mudstone and plane laminated, CWRL calcisiltite; common thin skeletal accumulations (0.5 cm thick); color medium dark gray.	18.59	277.34
38	Shale (90-95%) and limestone (5-10%); shale is medium gray, moderately to poorly fissile, with abundant whole fossils, calcareous to extremely calcareous (some originally logged as arqillaceous limestone but insoluable residues indicate most is calcareous shale); limestone is skeletal packstone, very thin bedded and some thin-bedded, thin skeletal accumulations (0.5 cm, abundant), peloidal lime mudstone, and laminated calcisiltite; packstone-laminated calcisiltite or lime mudstone couplets present but not abundant (less than 10%) of beds), some lime mudstones are nodular; slabbed samples reveal that at least some nodular bedding has a stylolitic fabric.	88.27	365.61
NOTE:	Units 39 through 41 measured along new U.S. Highway 311.		
39	Mostly not exposed; float suggests shale and very thin-bedded limestones similar to unit 38; thin bedded siltstones toward base.	67.51	433.12
40	Shale, light olive gray, calcareous.	3.05	436.17
41	Sandstone and shale; sandstone very fine grained to siltstone; both sandstone and shale highly burrowed; concentrations of *Lingula* in sandy packstones.	3.96	440.13

Bays Formation

Appendix I (cont.).

Narrows Section: Measured along U.S. Highway 460 approximately 1.5 km west of Narrows, Virginia. Units 1-20 of Martinsburg Formation measured along westbound lane.

Juaniata Formation

Martinsburg Formation - total thickness 407.82 meters

UNIT		THICKNESS *(meters)*	CUMULATIVE THICKNESS
1	Sandstone, fine grained, medium light gray, massively bedded due to bioturbation; black phosphatic modules.	6.83	6.83
2	Sandstone, fine grained, medium light gray, massively bedded due to bioturbation with few medium beds, sharp bases and burrowed upper contracts; apparent small displacement (?) fault between unit 1 and 2; black phosphatic nodules.	3.26	10.09
3	Sandstone, fine grained, calcareous, thin to medium bedded, laminated, interbedded with burrowed fine-grained sandstone; skeletal concentrations at base of a few beds; black phosphatic nodules.	1.46	11.55
4	Sandstone, fine to very fine grained, calcareous, light gray, thin to medium bedded, interbedded as laminated and bioturbated beds; abundant bryozoans; upper half with black phosphatic modules; nodules and skeletal accumulations at base of beds, and in burrow disrupted concentrations.	4.66	16.21
5	Sandstone and shale; sandstone fine grained, medium light gray, slightly calcareous, thin to medium bedded, variable in thickness to lenticular, CWRL and "hummocky" upper bed surfaces, sharp erosional bases; scattered bryozoan and brachiopod-rich sandstone; shale silty, moderately bioturbated.	2.56	18.77
6	Shale and sandstone; sandstone (40%) fine grained, light gray, calcareous, thin to medium beded, plane laminated and CERL, wave ripple cross-lamination, "hummocky" upper bed surfaces, beds have sharp and erosional bases, variable in thickness laterally, common bryozoan plus crinoid and brachiopod skeletal accumulations at bases; shale (60%) in thin to medium interbeds, light gray, sandy, calcareous, with some very thin lenticular sandstone.	6.28	25.05
7	Shale (59%), sandstone (40%) and skeletal packstone (10%); sandstone fine to very fined grained, light gray, plane laminated, wave ripple cross-laminated, HCS, CWRL, hummocky and ripple-marked tops, lenticular, often with sharp irregular bases and a few fossil fragments at bases; packstone thin to thick bedded, with erosional bases and many grading abruptly to laminated sandstone; shale in medium to thick beds, medium light gray, silty and sandy, clacareous.	4.90	29.95
8	Sandstone (50%) and shale (50%) sandstone fine to very fine grained, light gray, thin to medium-bedded, CWRL, hummocky tops, sharp erosional bases, some with thin fossil concentrations; shale thin to medium interbeds, calcareous.	2.01	31.96
9	Shale (75%), sandstone (25%); sandstone fine grained to siltstone, medium light gray, calcareous, thin and very thin bedded, plane laminated, CWRL, variable in thickness and lenticular; a few highly fossiliferous zones at bases of beds, erosional lower surfaces; shale medium gray, calcareous, with some very thin-bedded lenticular siltstone beds, partly bioturbate.	6.19	38.15
10	Siltstone to very-fine grained sandstone (60%), shale (30%) and skeletal packstone (10%); sandstone medium light gray, thin to medium bedded, plane laminated and CWL, hummocky tops, variable in thickness, scoured bases; skeletal packstone to highly fossiliferous sandstone, thin to medium bedded, generally grading abruptly upward to siltstone/sandstone.	3.02	41.17

Appendix I (cont.).

UNIT		THICKNESS (*meters*)	CUMULATIVE THICKNESS
11	Shale (65%), skeletal packstone (20%) and siltstone to very fine-grained sandstone (15%); packstone with erosional bases, extremely variable in thickness (including lenses), up to 0.3 m thick; siltstone with sedimentary structures as in unit 10, thin-bedded.	6.83	48.00
12	Siltstone to very fine-grained sandstone (50%), shale (40%) and skeletal packstone (10%); siltstone in thin to medium beds sedimentary structures as in unit 10, calcareous.	2.32	50.32
13	Sandstone, very fine grained, plane laminated, thins laterally to 9 cm.	0.30	50.62
14	Shale (60%), siltstone to very fine-grained sandstone (30%), and skeletal packstone (10%); siltstone thin bedded, sedimentary structures as in unit 10; common couplets of packstone and laminated siltstone.	3.38	54.00
15	Shale (70%) and sandstone (25%); sandstone very fine grained to siltstone, calcareous, medium light gray, medium to thin bedded, variable in thickness, plane laminated, CWRL, wave ripple cross-laminated, ripple marks and hummocky tops; skeletal concentrations at bases of beds and in shallow and scoured pockets, (up to 9 cm relief); minor calcarenite (5%); thin medium interbeds of silty shale.	8.29	62.29
16	Shale (55%), siltstone (35%) and skeletal packstone (10%); siltstone calcareous, thin bedded, plane laminated and CWRL; packstones include some which are dominantly bryozoans and many which are lenticular.	3.23	65.52
17	Skeletal packstone, medium light gray.	0.24	65.76
18	Shale (60%), siltstone to very fine-grained sandstone (20%),calcarenite (10%) and skeletal packstone (0%); shale medium gray to light olive gray, variably calcareous, thin to thick beds, locally highly burrowed, siltstone very thin to medium bedded, sedimentary structures same as units 15-16; packstone mostly thin bedded; couplets of packstone grading to laminated lithologies common.	29.29	95.05
19	Shale (70%), siltstone (15%), and limestone (15%); shale, light olive gray, in thin to thick beds; siltstone variably calcareous to silty calcarenite and calcisiltite; medium light gray, thin and very thin bedded, sedimentary structures same as unit 18; skeletal packstones 5%; two 0.3 meter thick horizons of chaotic soft sediment deformation similar to unit 21.	7.25	102.30
NOTE:	Units 20-27 exposed on eastbound lane of U.S. Highway 460 (lower road).		
20	Shale (60%), siltstone (30%), and skeletal packstone (10%); and shale fissile to moderately bioturbated, non-calcareous; siltstone variably calcareous to calcisiltite; sharp erosional bases, especially of packstone, packstone/laminated siltstone couplets; other sedimentary structures same as unit 19.	5.27	107.57
21	Structureless mudstone containing rotated and contorted thin beds of siltstone and skeletal packstone; due to soft sediment deformation; some isolated fragments of beds and other beds continuous but highly contorted.	1.52	108.82
22	Same as unit 20, except shale dominantly burrowed.	2.90	111.72
23	Soft sediment deformation; chaotic bed-and ball-and-pillow; lower 1 meter contains chaotically arranged fragments of siltstone in structureless mudstone, gradational to underlying unit; upper 0.5 meter ball-and-pillow structure composed mostly of one thick siltstone bed.	1.52	113.24
24	Shale (65%), siltstone (30%), and skeletal packstone (5%); shale generally fissile and nonburrowed; siltstone, thin and very thin bedded, variably calcareous, plane laminated, CWRL, wave ripple cross-laminated, two thick siltstone beds as ball-and-pillow structures; skeletal packstone thin to medium bedded, mostly in lower half of unit.	7.16	120.40

Appendix I (cont.).

UNIT		THICKNESS (*meters*)	CUMULATIVE THICKNESS
25	Soft sediment deformation; consists dominantly of 0.6 to 1 meter long, 0.3 to 0.45 meter thick rounded blocks of very fine-grained sandstone, discontinuous but apparently the same bed, "loaded" into underlying unit; lamination within blocks is contorted; discontinuous skeletal packstone at base of sandstone; upper part is chaotic, broken and rotated calcisiltite beds.	0.61	121.01
26	Shale (60-80%), skeletal packstone (10-30%), and calcisiltite (15%); shale variably calcareous, common 2-4 mm. Diameter horizontal and randomly oriented burrows; some unreworked concentrations of bryozoans; skeletal packstone, commonly with intraclastics (siltstone), very thin and thin beds, sharp erosional bases, many as couplets of packstone-calcisiltite; some packstone dominated with bryozoan concentrations in shale; calcisiltite thin bedded, terrigeneous silt abundant, some calcarenite, horizontal plane lamination, CWRL, and some wave ripple cross-lamination, bedding commonly lenticular.	28.65	149.66
NOTE:	Units 27-43 measured along westbound lane of U.S. Highway 460.		
27	Mostly not exposed; apparently mostly shale with thin and very thin beds of calcisiltite and silty calcisiltite weakly laminated to massive; several intervals dominated by bryozoans; thickness by compass and tape; shale locally poorly fissile due to bioturbation.	29.60	179.26
28	Poor exposure; lithologies and structures generally similar to units 29 and 30; shale apparently about 35% at base, increasing to greater than 50% at top; limestones are skeletal packstone, calcisiltites, weakly laminated to massive, and some calcarenite; colors medium dark gray to medium light gray, thin and very thin bedded, some nodular bedded.	34.05	213.31
29	Limestone (65%) and shale (35%); shale fissile (non-burrowed), calcareous; limestone in subequal amount of skeletal packstone and laminated calcarenite to calcisiltite, dominantly thin bedded.	7.77	221.08
30	Limestone (80%) and shale (20%); limestone is skeletal packstone (30%), skeletal/intraclast packstone (15%) in beds up to 0.3 meters thick, and calcarenite (35%) plus minor calcisiltite; colors medium to medium light gray; bedding thin to medium, thinner at top; plane laminated, CWRL, with hummocky upper bed surfaces; intraclasts irregular to tabular in shape, up to 20 cm; shale extremely calcareous in thin and very thin beds.	14.78	235.86
31	Limestone (80%) and shale (20%); limestone is skeletal packstone (25%), calcarenite (30%) and calcisiltite (25%), commonly in packstone/laminated bed couplets; medium to medium light gray, very thin to medium beds, plane laminated CWRL, hummocky tops and ripple marks; some calcisiltites indistinctly laminated to massive, some nodular; several intraclast/skeletal packstones; shale in thin and very thin beds.	7.56	243.42
32	Limestone (70%) and shale (30%); limestone is skeletal packstone, calcarenite and calcisiltite, commonly in couplets; sedimentary structures same as unit 31; some thin calcisiltites nodular, some calcarenites massive; bedding thin and very thin.	35.36	278.78
33	Not exposed.	13.38	292.16
34	Limestone (80%) and shale (20%); limestone is skeletal packstone (40%), and calcarenite (40%), commonly in couplets; thin to medium bedded; CWRL lamination and hummocky upper bedding planes common.	22.71	314.87
35	Limestone (60%) and shale (40%); limestone is skeletal packstone (20%) and subequal amounts of calcisiltite and calcarenite (total 40%); medium light gray, very thin and thin bedded, plane laminated and CWRL but some calcisiltite is massive, some nodular bedded.	15.67	330.54

Appendix I (cont.).

UNIT		THICKNESS (meters)	CUMULATIVE THICKNESS
36	Limestone (50%) and shale (50%) limestone is skeletal packstone, calcarenite and calcisiltite, commonly in couplets; medium to medium light gray thin bedded to thin and very thin bedded toward top of unit, some laminated beds contain appreciable amounts of terrigenous silt and sand; some very thin calcisiltites nodular; shale in beds up to 1.0 m thick.	20.67	351.21
37	Limestone and shale; limestone dominant, mostly as whole fossil skeletal packstone in medium to thick beds; unit includes small covered interval at culvert.	15.30	366.51
38	Limestone (80%) and shale (20%); limestone is skeletal packstone and calcarenite, commonly in couplets; thin and medim bedded but several calcarenites up to 0.3 m thick; horizontal plant laminated, CWRL, calcarenites include abundant laminae of terrigenous silt and sand; several percent laminated siltstone to very fine-grained sandstone; shale has good fissility and horizontal branching burrows but few body fossils; packstones include intraclasts of calcarenite.	18.84	385.35
39	Limestone (85%) and shale (15%); limestone is skeletal packstone (40%) and calcarenite (45%), commonly in couplets; medium to thick bedded, plane laminated, HCS and CWRL; approximately 20% terrigenous sand and silt in calcarenites; one thin zone (0.3 m) of soft sediment deformation (ball-and-pillow and apparent slump folding); unit starts at 0.3 m thick shale bed and ends at small tight anticline.	11.58	396.93
40	Limestone (85%) and shale (15%), limestone dominantly calcarenite (70%) and lesser skeletal packstone (15%); thin to medium bedded and some thick bedded, sedimentary structures same as unit 39; terrigenous sand and silt abundant in calcarenite, plus a few siltstone to very fine-grained sandstone beds.	4.69	401.62
41	Clay (bentonite)	0.21	401.83
42	Limestone (65%), shale (32%), and siltstone to very fine grained sandstone (3%). Limestone as in unit 40, subequal amounts of skeletal packstone and calcarenite, medium-bedded; siltstone thin bedded, light olive gray.	2.41	404.24
43	Poorly exposed; dominantly limestone as above with shale and minor siltstone to very fine-grained sandstone.	3.41	407.65

Eggleton Formation

Walker Mountain Section: Measured along Virginia Highway 16, beginning about 100 meters from the crest of Walker Mountain and continuing down the northeast facing slope, Smyth County, Virginia. The measured section is compiled from numerous segments, in part overlapping, exposed along "switch-backs" of Highway 16.

Juniata Formation

Martinsburg Formation - total thickness 418.22 meters.

UNIT		THICKNESS (meters)	CUMULATIVE THICKNESS
1	Sandstone, fine to very fine grained, calcareous, medium gray, medium to thick bedded; plane laminated beds with skeletal concentrations at base, grading upward to highly burrowed tops; phosphatic nodules.	3.51	3.51
2	Sandstone fine to very fine grained, medium gray, thick to very thick bedded due to intense bioturbation; vague bedding planes defined by concentrations of *Orthorhynchula*, *Lingula*, and phosphatic nodules.	6.86	10.37

Appendix I (cont.).

UNIT		THICKNESS *(meters)*	CUMULATIVE THICKNESS
3	Sandstone, fine to very fine grained, medium to thick bedded due to intense bioturbation; occasional thin to medium bedded plane laminated beds with erosional bases, fossil concentrations at bases, and burrowed tops; phosphatic nodules.	5.21	15.58
4	Sandstone and shale; sandstone fine to very fine grained, medium gray to light olive gray, thin to medium bedded, plane laminated, HCS, CWRL, variable in thickness to lenticular, sharp erosional bases with fossils concentrated at base of many beds, gradational or burrowed tops; shale sandy, medium gray to light olive gray, highly burowed in places.	4.75	20.33
5	Sandstone, very fine grained, medium gray, thick bedded due to intense bioturbation.	1.10	21.43
6	Sandstone, shale, and minor skeletal packstone; sandstone very fine grained, thin to thick bedded, plane laminated, HCS, CWRL, bedding thickness variable laterally to lenticular; fossils concentrated at base of beds and locally in lenses within sandstone beds; skeletal packstone at base of some sandstone beds and as packstones enclosed in shale; some filling shallow scours.	30.18	51.61
7	Shale (60%), sandstone (30%), and skeletal packstone (10%); shale locally sandy, locally highly burrowed, strong cleavage locally, biovalves common; sandstone very fine grained to siltstone, very thin to thin bedded, rarely medium bedded, plane laminated and CWRL; skeletal packstone very thin to thin bedded.	7.47	59.08
8	Sandstone (50%) and shale (50%); sandstone very fine grained, medium gray, thin and some medium bedded, with fossils concentrated at base of some beds; sedimentary structures as in unit 7.	8.23	67.31
9	Shale (70%), sandstone (20%) skeletal packstone (10%); shale medium light gray, in beds up to 0.67 meters thick; sandstone, very fine grained to siltstone, thin and very thin bedded, calcareous; skeletal packstone commonly in packstone/sandstone couplets; other sedimentary structures same as unit 7.	24.54	91.85
10	Shale (55%), sandstone (35%) and skeletal packstone (10%); shale medium gray to medium light gray in beds up to 0.5 meters thick; sandstone very fine grained to siltstone, thin bedded, rarely medium bedded and very thin bedded; skeletal packstone includes bryozoan packstone, shale intraclasts; packstone/laminated bed couplets and sedimentary structures same as unit 7.	16.92	108.77
11	Shale (50%), sandstone (35%), and skeletal packstone (15%) lithologies and sedimentary structures same as unit 10.	26.94	135.72
12	Shale (55%), sandstone (25%), and limestone (20%); lithologies and sedimentary structures same as unit 11, but limestone includes a few percent sandy, laminated calcarenite.	32.77	168.48
13	Shale (80%) and limestone (20%); shale medium gray to light olive gray in beds 0.15 to 1 meter thick, includes several percent very thin beds of siltstone toward the top of unit; limestone is skeletal packstone and laminated calcarenite to calcisiltite, both with abundant terrigenous silt, medium light gray, very thin- to medium-bedded, commonly in packstone/laminated bed couplets, plane laminated and CWRL, rare tabular ripple cross-laminated in sets 10-20 cm long.	32.16	200.64
14	Not exposed.	57.00	257.64
15	Limestone (70%) and shale (30%); limestone is skeletal packstone (30%), calcarenitee (20%) and calcisiltite (20%), thin and very thin-bedded; packstone/laminated bed couplets common; plane laminated and CWRL.	31.09	288.73

UNIT		THICKNESS *(meters)*	CUMULATIVE THICKNESS
16	Not exposed.	15.24	303.97
17	Shale (55%) and limestone (45%); shale ranging from 60% at base of unit to 50% at top; limestone is skeletal packstone (15%) and calcisiltite and minor calcarenite (total 30%), with abundant terrigenous silt (few beds of siltstone); very thin to medium-beded, packstone/laminated bed couplets common; plane laminated band CWRL, but many calcisiltites appear structureless.	62.50	366.44
18	Not exposed.	10.00	376.47
19	Partially covered; shale (50%) and limestone (50%); lithologies and sedimentary structures same as unit 17.	13.41	389.88
20	Partially covered, folded; shale, siltstone and limestone; siltstone very thin bedded. calcareous to silty calcisiltite; thin fossil concentrations at base of beds; skeletal packstone (10%) in thin and very thin beds and thin concentrations in shale, several shell layers in thickness.	12.91	402.07
21	Mostly not exposed; thickness by compass and tape; shale and siltstone; siltstone calcareous, very thin bedded, medium gray to olive gray; shell in thin concentrations, several shell layers thick.	16.15	418.22

Bays Formation

Hagan Section: Measured along the Louisville and Nashville Railroad siding at the Village of Hagan, Virginia, 1.5 km north of U.S. Highway 58.

Sequatchie Formation

"Reedsville" Formation - Total thicknes 124.06 meters.

UNIT		THICKNESS *(meters)*	CUMULATIVE THICKNESS
1	Limestone (80%) and shale (20%); shale extremely calcareous, burrowed, poorly fissle; limestone is skeletal packstone and minor peloidal-skeletal calcarenite to calcisiltite; limestones are irregularly bedded with gradational lower and upper bedding contacts, apparently burrowed.	21.64	21.64
2	Limestone (60%) and shale (40%); limestone is skeletal packstone and argillaceous to sandy laminated calcarenite and calcisiltite; packstone/laminated bed couplets common; packstones have sharp erosional bases, thin bedded; laminated beds plane laminated, CWRL.	6.10	27.74
3	Skeletal packstone with argillaceous partings, thick bedded.	3.05	30.79
4	Shale an minor limestone; shale very calcareous, burrowed; limestone is skeletal packstone and calcisiltite, very thin-bedded.	2.13	32.92
5	Limestone (95%) and shale (5%); limestone is skeletal packstone and coarse calcarenite, thin bedded near base grading to thick-bedded at top; shale in thin interbeds in lower 2/3 unit.	3.35	36.27
6	Limestone (65%) and shale (35%); shale medium gray, generally fissile, but burrowed locally; limestone is skeletal packstone and laminated calcarenite to calcisiltite commonly as packstone/laminated bed couplets; bedding thin with some very thin and medium; laminated limestones contain abundant terrigenous silt and sand and some are extremely calcareous siltstones and very fine-grained sandstone; CWRL, plane laminated.	7.77	44.04
7	Same lithologies and structures as unit 6; shale 30%, limestone 70%; thin and medium bedded; spans minor fault.	18.14	62.18
8	Same lithologies and structures as unit 7; shale 40-50%, limestone 50-60%; dominantly thin bedded; load casts at base of several beds; one bed as ball-and-pillow structure.	51.82	114.00

Appendix I (cont.).

UNIT		THICKNESS (*meters*)	CUMULATIVE THICKNESS
9	Shale (70%), and skeletal packstone (10%), and calcareous siltstone to silty calcisiltite (20%); shale calcareous, medium gray to olive gray, minor bioturbation.	10.06	124.06
"Trenton" Formation - Total thickness 181.66 meters.			
1	Skeletal packstone and minor coarse skeletal calcarenite (grainstone), medium light gray, medium to thick bedded.	4.88	4.88
2	Limestone (85%) and shale (15%); limestone is skeletal packstone (25%) and laminated calcarenite to calcisiltite (60%), commonly in couplets, thin and very thin bedded, plane laminated and CWRL; bed thickness variable laterally to lenticular; shale 10% in lower 2/3 of unit, increasing to 25% in upper 1/3; lenses and nodules of chert locally.	25.91	30.79
3	Not exposed.	48.77	79.56
4	Limestone (90%) and shale (10%); limestone is skeletal packstone (50%) and laminated calcisiltite, medium light gray, thin and medium-bedded, with zones very thin-bedded; packstones commonly with scattered intraclasts, variable in thickness laterally, calcisiltites plane laminated, CWRL.	59.44	139.00
5	Skeletal packstone (90%) and shale (10%); packstones commonly with scattered intraclasts medium light gray, thin to medium bedded.	4.57	143.57
6	Poorly exposed; float and scattered exposures suggest same as units 5, 7, and 8.	16.15	159.72
7	Skeletal packstone (60%) and shale (40%); packstone medium light gray, mostly thin-bedded, variable in thickness laterally; shale in very thin and thin beds; bentonite reported by Miller and Brosge (1954) at top of this interval, no longer exposed.	9.14	168.86
8	Skeletal packstone with shale partings (5%) also minor coarse skeletal calcarenite; medium light gray, thin to medium bedded, bed thickness variable laterally.	12.80	181.66
Eggleston Formation			

APPENDIX II: PETROGRAPHY

Sandstones

Constitutent:	Average (%)	Range (%)
Quartz	33.7	21.0 - 46.2
Rock fragments	14.0	4.2 - 21.8
Plagioclase	3.4	0.7 -7.2
K-feldspar	0.9	0 - 3.6
Chert	0.8	0 - 3.6
Mica	0.7	0 - 2.1
Heavy minerals	0.1	0 - 0.7
Opaques	0.7	0 - 2.8
Galuconite	Trace	0 - 0.8
Phosphatic nodules	0.9	0 - 9.5
Shells		
Calcite	3.4	0 - 11.8
Collophane	0.6	0 - 5.1

Appendix II con't.

	Average (%)	Range (%)
Matrix	16.5	4.2 - 43.2
Cement:		
CaCO	9.9	0 - 22.2
SiO	10.1	6.9 - 17.7
Phosphate	1.5	0 - 10.9
Feldspar	Trace	0 - 0.6
FeO/FeCa (CO)	0.2	0 - 1.5
Framboidal pyrite	0.7	0 - 2.4
Unknown:	1.9	0 - 3.6
(16 samples)		
Major Constituents:		
Quartz	51.2	25.3 - 73.6
Feldspar	6.4	1.0 - 14.8
Rock fragments	19.6	8.3 - 28.7
Matrix	22.8	8.3 - 49.3
Q/F/R		
Quartz	65.5	50.0 - 82.1
Feldspar	7.9	1.3 - 16.8
Rock Fragments	26.5	9.1 - 44.4
Brachiopod	12.8	2.2 - 19.9
Mollusk	7.8	Tr - 19.9
Echinoderm	4.4	0.6 - 27.0
Trilobite	2.1	Tr - 9.6
Bryozoan	2.5	0 - 11.6
Ostracod	0.4	0 - 11.7
Unidentified skeletal	3.9	0.7 - 10.5
Peloids	6.8	Tr - 29.0
Intraclasts	0.5	0 - 4.7
Terrigenous		
Quartz	2.6	Tr - 17.4
Feldspar	0.1	Tr - 2.0
Rock fragments	0.9	Tr - 9.9
Heavy minerals	Trace	Tr - 0.7
Matrix (mostly lime mud)	24.2	3.4 - 69.0
Cement:		
CaCO	27.3	8.5 - 49.6
SiO_2	1.6	0 - 2.7
Phosphate	Trace	0 - 1.8
Dolomite	0.9	0 - 0.9
Framboidal pyrite	0.5	0.7 - 1.4
Unknown:	0.9	0 - 2.5
(22 samples)		

APPENDIX III: COMMON MARTINSBURG TAXA

BRACHIOPODA: ARTICULATA
Rafinesquina alternata
R. fracta
Onniella sp. 1 (formerly identified as Dalmanella fertilis)
Onniella sp. 2 (formerly identified as Dalmanella multisecta)
Onniella sp. 3 (formerly identified as Dalmanella bassleri)
Sowerbyella curdsvillensis
S. rugosa
Zygospira lebanonensis
A. modesta
Hebertella sinuata
H. frankfortensis
Platystrophia sp.
Rhynchotrema sp.
Dinorthis sp.
Orthorhynchula linneyi

BRACHIOPODA: INARTICULATA
Lingula sp.
Craniops sp.

MOLLUSCA: BIVALVIA
Ambonychia 'praecursa'
Modiolopsis modiolaris
Ischyrondonta sp.
Ctenodonta sp.
Trancrediopsis sp.
Praenucula sp.
Cyrtodonta sp.
Pterinea sp.
Cuneamya sp.

MOLLUSCA: GASTROPODA
Plectonotus sp.
Bucania sp.
Loxoplocus sp.

ANTHROPODA: TRILOBITA
Isotelus spp.
Cryptolithus sp.
Ceraurus sp.
Flexicalymene sp.
Eomonorachus sp.

BRYOZOA
Prasopora spp.
Monticulipora sp.
Dekayia sp.
Hallopora sp.

ENCHINODERMANA: CRINOIDEA
Crinoid columnals

MAFIC ROCKS IN THE ALLIGATOR BACK FORMATION, THE UPPER UNIT OF THE LYNCHBURG GROUP, IN THE SOUTHWESTERN VIRGINIA PIEDMONT: AN OPHIOLITE SEQUENCE?

James F. Conley

CONTENTS

ILLUSTRATIONS

ABSTRACT

The Lynchburg Group in the southwestern Virginia Piedmont is composed of a lower Ashe Formation and an upper Alligator Back Formation. The Alligator Back, the upper formation of the group, locally contains a basal sequence composed of metamorphosed ultramafic rocks, overlain by metagabbros, which are in turn overlain by metabasalts. These ultramafic and mafic plutonic and volcanic rocks are overlain by metasedimentary rocks that are predominantly metagraywackes. These metagraywackes may locally contain interlayers of impure marbles; mica schists; graphite-quartz-mica schists, graphite-mica schists, and calcareous quartzites. This sequence of ultramafic-mafic rocks, overlain by metasedimentary rocks, is repeated at least once and possibly twice in the stratigraphic section and is interpreted to be ophiolites, on which were deposited deep marine sediments, all of which have been obducted onto continental crust. It is believed that this ophiolite sequence was obducted during a period of previously unrecognized plate closing in late-Precambrian time. This period of plate closing occurred prior to the great outpouring of the Catoctin continental basalts in central and northeastern Virginia at the beginning of continental fracturing, which preceded formation of the Iapetus Ocean in latest Precambrian time.

INTRODUCTION

In the southwestern Virginia Piedmont the Lynchburg Group (Figure 1) has been mapped at 1:24,000 scale by various workers (Conley and Henika, 1970; Henika, 1971; Conley and Henika, 1973, McCollum, in preparation; and Conley, Piepul, Robinson and Lemon, in preparation). The Lynchburg Group, a highly variable lithologic unit, was named the Lynchburg Formation by Jonas (1927) for exposures near the City of Lynchburg, which lies about 50 miles (80 km) on strike to the northeast of the area discussed in this paper. Furcron (1969) renamed the unit the Lynchburg Group. Rankin (1970) and Rankin and others (1973) recognized this sequence of rocks and subdivided it into a lower unit which they named the Ashe Formation and an upper unit which they named the Alligator Back Formation. In the southwestern Piedmont Conley (1985) recognized a lower Ashe Formation and an upper Alligator Back Formation and, using the terminology of Furcron proposed that these two formations composed the Lynchburg Group. The Lynchburg Group crops out on the southeastern limb of the Blue Ridge anticlinorium from Culpeper County in north-central Virginia to the North Carolina boundary. Metabasalts, metagabbros, and ultramafic rocks occur in the Lynchburg Group and especially in the upper unit, the Alligator Back Formation. These and other mafic and ultramafic rocks of the Southern Appalachians have been reviewed in detail by Misra and Keller (1978).

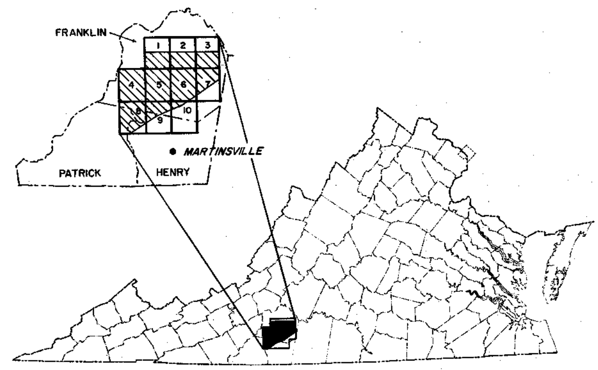


Figure 1. Index map of area. (1) Boones Mill quadrangle; (2) Redwood quadrangle; (3) Moneta S. W. quadrangle; (4) Ferrum quadrangle; (5) Rocky Mount quadrangle; (6) Gladehill quadrangle; (7) Penhook quadrangle; (8) Philpott Reservoir quadrangle; (9) Bassett quadrangle; (10) Snow Creek quadrangle.

In the southwestern Virginia Piedmont (Figure 2) the Lynchburg Group is in the footwall of the Bowens Creek fault, a southeastward dipping thrust that separates rocks of the Blue Ridge anticlinorium to the northwest from rocks of the Smith River allochthon to the southeast (Conley, 1985). In the region under discussion metamorphic grade ranges from greenschist to lower amphibolite.

STRATIGRAPHY

The basal part of the Alligator Back Formation is composed of a persistent band of amphibolite and mafic and ultramafic rocks (Figure 3). Up section from these basal rocks, the formation also contains abundant amounts of ultramafic rocks, metagabbros and metabasalts with associated massive, fine-grained quartzites and ferruginous quartzites. Although not commonly mapped as faults, the basal contacts of these plutonic-vol-

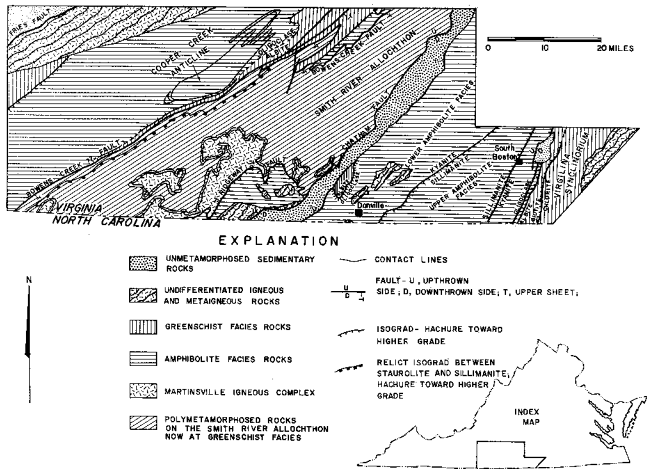


Figure 2. Regional geologic map.

Figure 3. Contact between Ashe Formation and Alligator Back Formation on the southeastern limb of the Cooper Creek anticline exposed on the north side of State Road 718 (northwestern corner of the Snow Creek quadrangle). Gneiss of the Ashe Formation at bottom of picture is overlain by amphibolite at the base of the Alligator Back. Rocks are folded by northwest trending open folds.

canic sequences are generally sheared. The remainder of the Alligator Back, which makes up about 60 percent of the unit, is a metamorphosed clastic sequence. These clastic rocks are metagraywackes and metagraywacke conglomerates interlayered with graphitic schists, graphitic quartz schists, graphitic quartzites, quartzites, muscovite-sericite schists, and impure marbles. Stratigraphic horizon markers are not traceable over great distances and lithologic units are lenticular.

Ultramafic Rocks

The ultramafic rocks consist of metamorphosed dunites and metapyroxenites. Three of the largest ultramafic bodies in the area are the Grassy Hill sequence in the northwestern part of the area (Rocky Mount quadrangle), Jacks Mountain sequence in the northeastern part of the area (Penhook quadrangle) and Brier Mountain sequence in the core of the Cooper Creek anticline (Rocky Mount quadrangle) in the south-central part of the area (Figures 1 and 4). The ultramafic bodies that make up Grassy Hill and Jacks Mountain both have a northeastward trend, parallel to regional structure, whereas Brier Mountain has an east-west trend across the axis of the Cooper Creek anticline. All three of these bodies contain both ultramafic rock and metagabbro. The Brier Mountain and Jacks Mountain bodies contain metabasalts, especially around their perimeters, thus indicating a close association among the ultramafic rocks, the metagabbros and the metabasalts. The ultramafic rocks contain a few relict olivine (forsterite) and large hypersthene and augite grains, but these rocks are now almost totally altered to serpentine, chlorite, talc, and tremolite.

The margins of the ultramafic bodies are sheared and altered to chlorite that is 1-2 m (3-6 feet) thick; the chlorite in places is further altered to vermiculite. The ultramafic bodies are locally cut by veins of anthophyllite that are a few centimeters thick. Opaque minerals (magnetite and possible chromite) occur in accessory amounts in all the ultramafic rocks.

Metagabbro

Metagabbro occurs as interlayers in the ultramafic rocks and also as discontinuous to relatively continuous concordant bands at the base of the sequences of metabasalts in the area. They are not as altered as the ultramafic rocks. A few metagabbro bodies, like those west and southwest of Rocky Mount, are isolated masses that are surrounded by metasedimentary rocks. Some of these bodies are cored by ultramafic rocks, which suggests that they might be overturned sequences. The metagabbros locally exhibit compositional layering. Perimeters of metagabbro, espe-

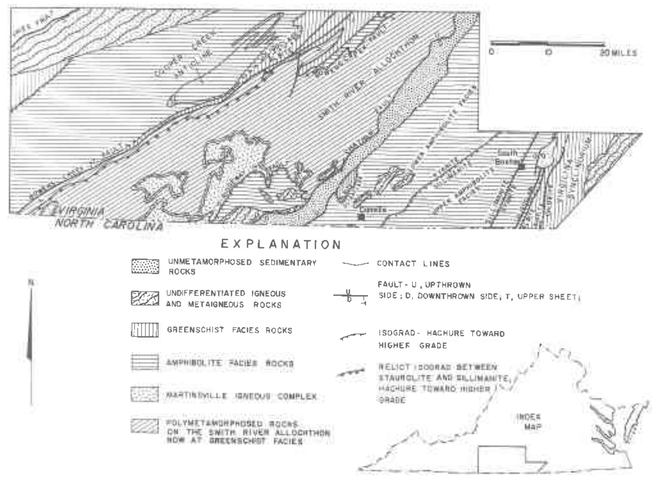


Figure 2. Regional geologic map.

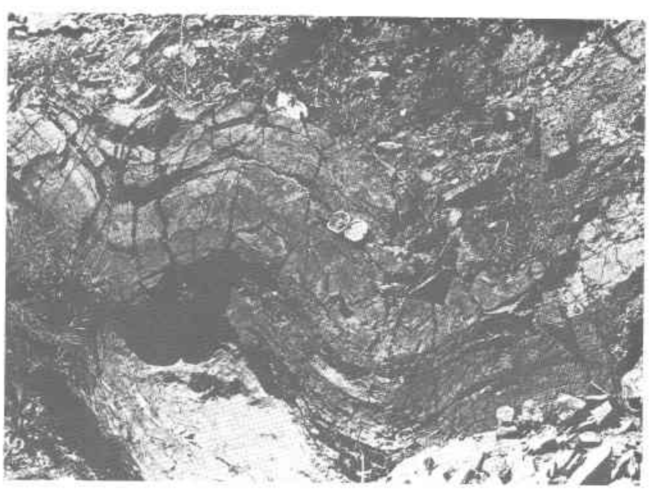


Figure 3. Contact between Ashe Formation and Alligator Back Formation on the southeastern limb of the Cooper Creek anticline exposed on the north side of State Road 718 (northwestern corner of the Snow Creek quadrangle). Gneiss of the Ashe Formation at bottom of picture is overlain by amphibolite at the base of the Alligator Back. Rocks are folded by northwest trending open folds.

canic sequences are generally sheared. The remainder of the Alligator Back, which makes up about 60 percent of the unit, is a metamorphosed clastic sequence. These clastic rocks are metagraywackes and metagraywacke conglomerates interlayered with graphitic schists, graphitic quartz schists, graphitic quartzites, quartzites, muscovite-sericite schists, and impure marbles. Stratigraphic horizon markers are not traceable over great distances and lithologic units are lenticular.

Ultramafic Rocks

The ultramafic rocks consist of metamorphosed dunites and metapyroxenites. Three of the largest ultramafic bodies in the area are the Grassy Hill sequence in the northwestern part of the area (Rocky Mount quadrangle), Jacks Mountain sequence in the northeastern part of the area (Penhook quadrangle) and Brier Mountain sequence in the core of the Cooper Creek anticline (Rocky Mount quadrangle) in the south-central part of the area (Figures 1 and 4). The ultramafic bodies that make up Grassy Hill and Jacks Mountain both have a northeastward trend, parallel to regional structure, whereas Brier Mountain has an east-west trend across the axis of the Cooper Creek anticline. All three of these bodies contain both ultramafic rock and metagabbro. The Brier Mountain and Jacks Mountain bodies contain metabasalts, especially around their perimeters, thus indicating a close association among the ultramafic rocks, the metagabbros and the metabasalts. The ultramafic rocks contain a few relict olivine (forsterite) and large hypersthene and augite grains, but these rocks are now almost totally altered to serpentine, chlorite, talc, and tremolite.

The margins of the ultramafic bodies are sheared and altered to chlorite that is 1-2 m (3-6 feet) thick; the chlorite in places is further altered to vermiculite. The ultramafic bodies are locally cut by veins of anthophyllite that are a few centimeters thick. Opaque minerals (magnetite and possible chromite) occur in accessory amounts in all the ultramafic rocks.

Metagabbro

Metagabbro occurs as interlayers in the ultramafic rocks and also as discontinuous to relatively continuous concordant bands at the base of the sequences of metabasalts in the area. They are not as altered as the ultramafic rocks. A few metagabbro bodies, like those west and southwest of Rocky Mount, are isolated masses that are surrounded by metasedimentary rocks. Some of these bodies are cored by ultramafic rocks, which suggests that they might be overturned sequences. The metagabbros locally exhibit compositional layering. Perimeters of metagabbro, espe-

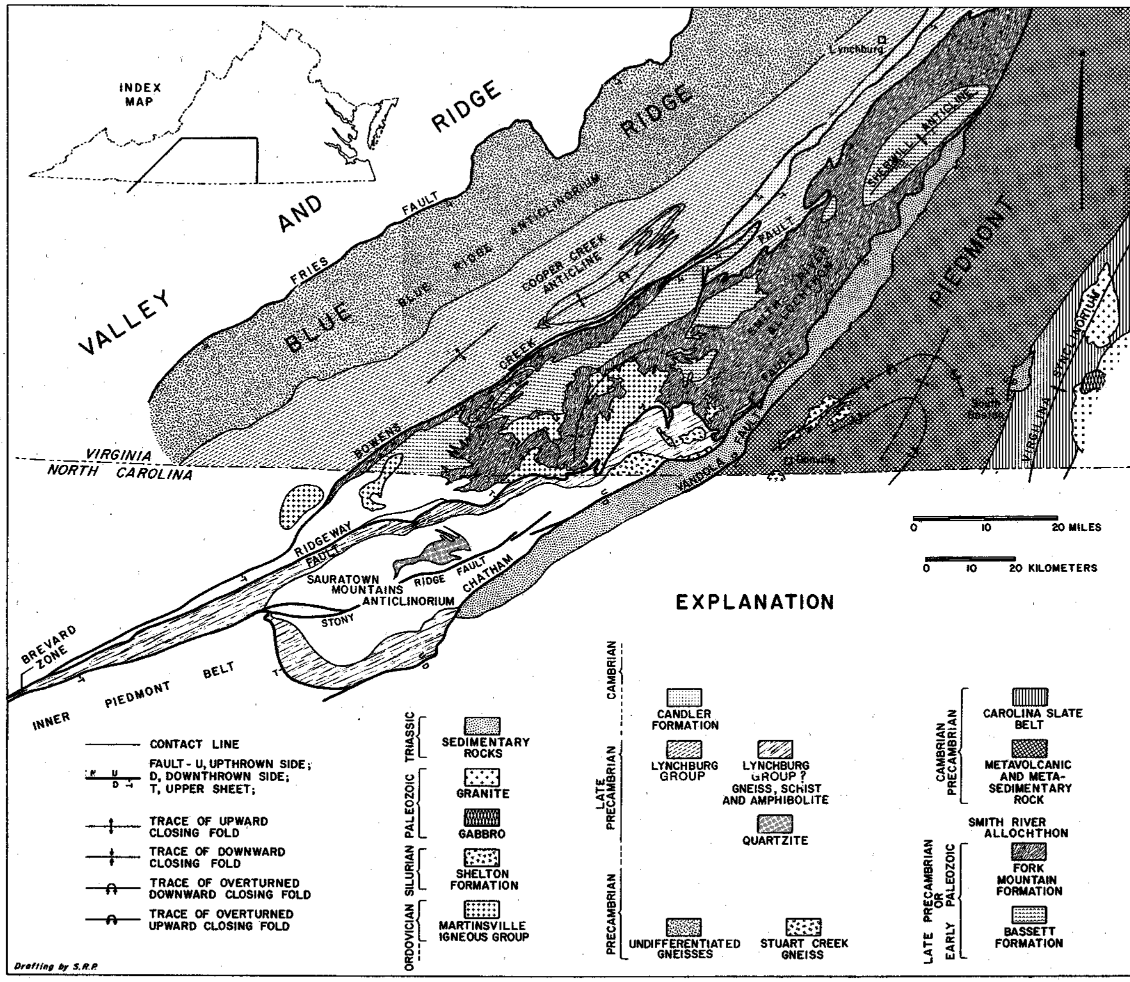


Figure 4. Geologic map of parts of the Blue Ridge anticlinorium and Smith River allochthon, southwestern Piedmont of Virginia.

cially at the base of metagabbro-metabasalt sequences, generally have a cataclastic texture and are locally sheared and altered to actinolite schists and chlorite schists. The ultramafic rocks and the metagabbros have not produced contact metamorphism in country rocks (Figure 5). Although Conley and Henika (1970) report a dense, gray hornfels-like rock at the top of the large metagabbro body north of Fairy Stone State Park. Examination of thin sections of this rock indicate that it is a highly indurated graphitic quartzite. No growth of contact metamorphic minerals or development of hornfels texture could be detected even as relict grains or altered masses. Thin, massive, fine-grained quartzite and ferruginous quartzites are in contact with the metagabbros in the area west of Rocky Mount, and along the northwest shore of Fairy Stone Lake in Fairy Stone State Park. Magnetite deposits are associated with the quartzites and ferruginous quartzites in contact with the metagabbros and were mined for iron ore on Stuarts Knob, a hill underlain by metagabbro located in the northern part of Fairy Stone State Park. Similar occurrences of magnetite were also mined in areas west and southwest of Rocky Mount.

The metagabbros are coarse grained and have a relict ophitic texture. At greenschist facies they are composed of actinolite (that in part is uralitic amphibole), albite, epidote, chlorite, and minor amounts of zircon, quartz, sphene, ilmenite and magnetite. Relict augite and calcic plagioclase, partially altered to epidote, were identified in thin section. At lower amphibolite facies oligoclase and hornblende are present in the rock.

Metabasalt

The metabasalt is a dark-greenish-black, fine-grained, schistose rock composed of actinolite

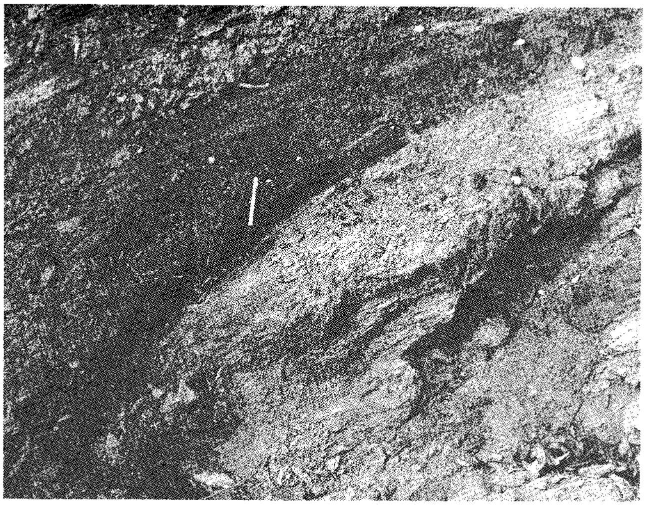


Figure 5. Contact (at pen) between altered ultramafic rock and metasedimentary quartz mica schist, both belonging to the Alligator Back Formation on the southeastern flank of Grassy Hill, just north of the intersection of State Road 919 and U. S. Highway 220 (Boones Mill quadrangle). There is no contact metamorphism of the underlying mica schist by the ultramafic rock.

nematoblasts and variable amounts of epidote, chlorite, albite and quartz. Where metagabbros are present, the metabasalts directly overlie the metagabbros; where metagabbros are absent, the metabasalts are separated by a shear zone of varying thickness from the underlying units. The tops of the metabasalts are generally interlayered with overlying metasedimentary rocks (Figures 6 and 7). The metabasalts locally contain minute amygdules (1-2 mm), now filled with epidote, and show a primary texture, that may possibly represent transposed flow lines or flattened and attenuated pillow structures (Figure 8) (Conley and Henika, 1970).

Metagraywacke

The metagraywackes in the Lynchburg are gray rocks that range in grain size from sandstone to conglomerate. The predominant rock generally shows graded bedding and is composed of granule to sand-size clasts of quartz and feldspar in a finer grained quartz-feldspar-mica matrix. This rock generally contains thin mica schist interlayers and may grade, by a gradual decrease in quartz and increase in mica, into schists (Figures 9 and 10). The metagraywackes occur as thick sequences, and as thin interlayers and lenses in finer grained mica schists and graphitic

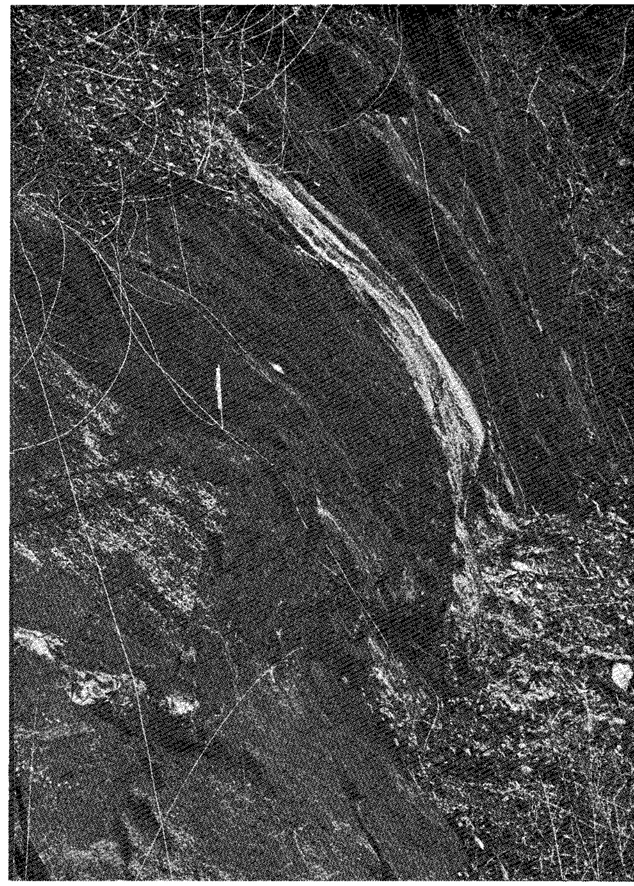
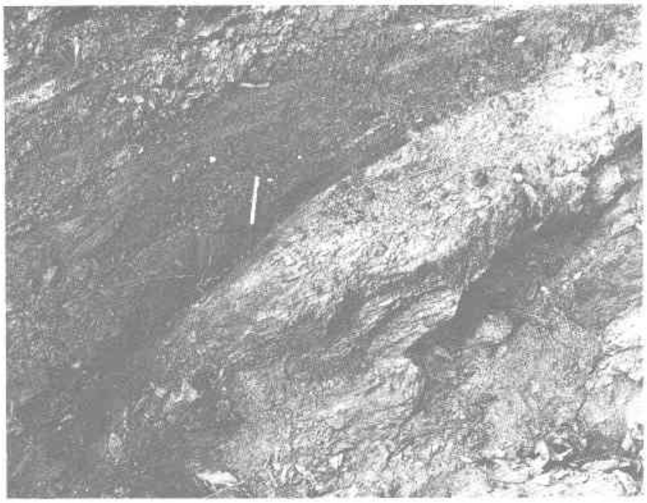


Figure 6. Alligator Back Formation exposed on east bank of U. S. Highway 220 Business Route, at the southern city limits of Rocky Mount (Rocky Mount quadrangle). Metagabbro exposed in lower left hand corner of photograph is overlain by metabasalt (contact just above pen) that contains interlayers of metasedimentary rock (light colored rock).

schists. A metaconglomerate that shows graded bedding and contains rock fragments up to a foot in length is exposed in Town Creek on the southeast limb of the Cooper Creek anticline (Henika, 1971). In addition, this metaconglomerate contains angular fragments (rip-up clasts) of metaclaystones and graphitic metaclaystones (Figure 10). At this locality, bedding is not only graded but also repetitive. The basal part of one of these beds is a coarse, structureless, polymictic conglomerate. Above the base the bed grades upward into finer material. Some graded beds are overlain by a layer of finely laminated sandstone that in turn may be overlain by a thin cross-bedded sequence. The rocks exposed in Town Creek are very similar to the units A through C of Bouma sequence rocks found in proximal turbidite dep-



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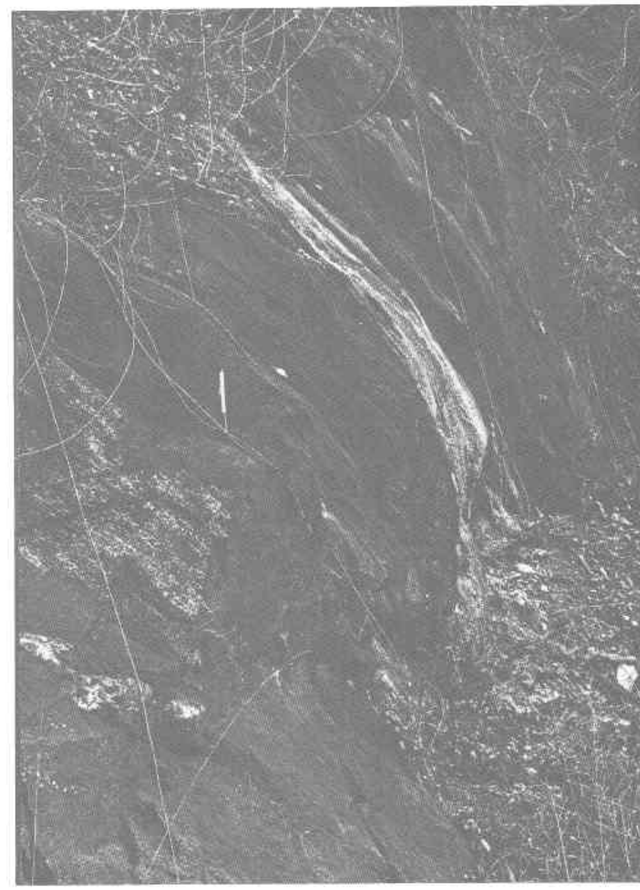

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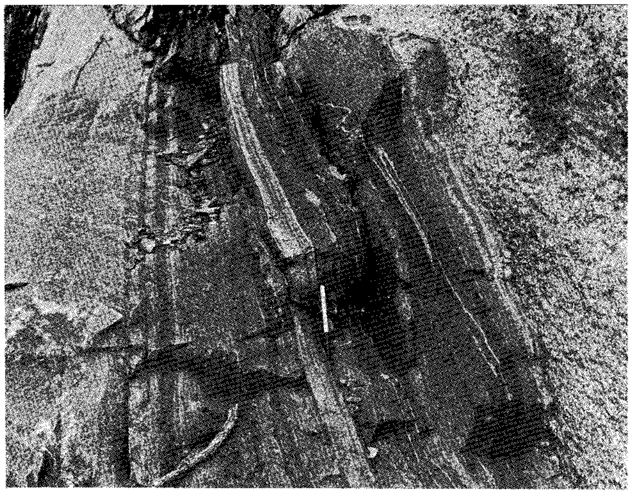


Figure 7. Thin metabasalt layer in metagraywacke exposed on the east side of State Road 1008, just south of its intersection with U. S. Highway 220 (Boones Mill quadrangle). Metabasalt layer (at pen) lays between coarse, conglomeratic metagraywacke on the right and metagraywacke that shows graded bedding on the left. Top of beds is to the left. This outcrop is at the top of a thick section of metabasalt located along U. S. Highway 220 just north of Rocky Mount.

Figure 8. Metabasalt in the Alligator Back Formation on the southeastern flank of Bald Knob, east of Rocky Mount (Gladehill quadrangle). Transposed and isoclinally refolded lens-shaped masses are thought to be primary structures, possibly transposed flow banding or flattened and attenuated pillow structures.

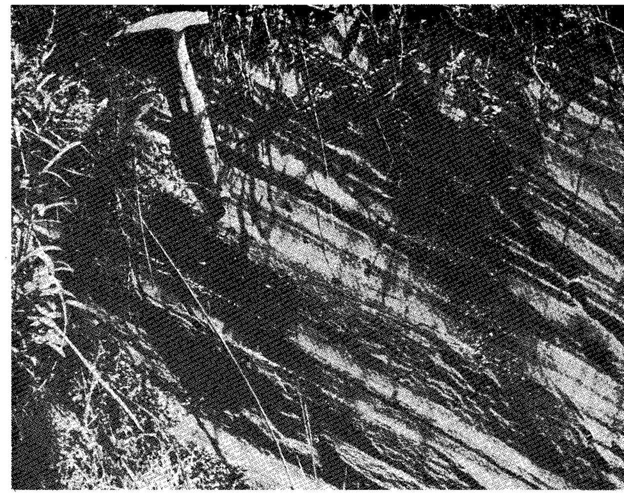


Figure 9. Thin-bedded metagraywacke of Alligator Back Formation on U. S. Highway 220 on the south bank of Big Chestnut Creek (Snow Creek quadrangle).

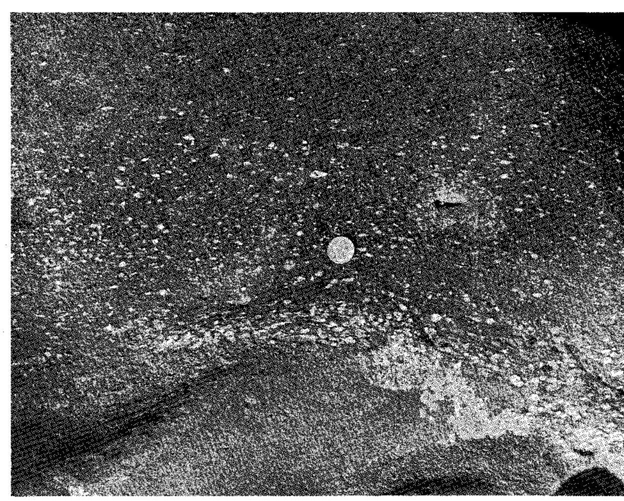
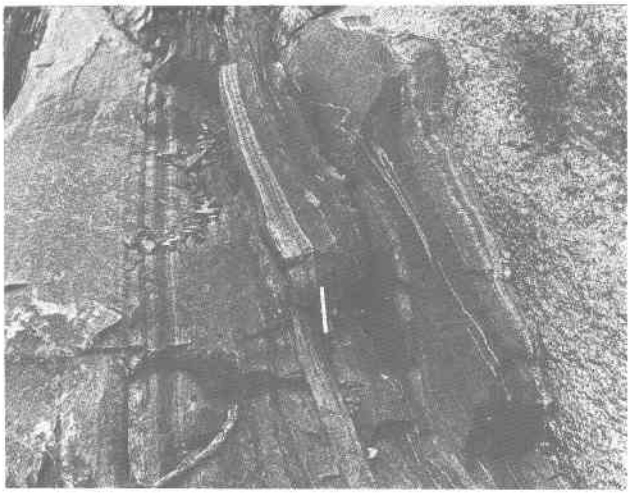


Figure 10. Lithic conglomerate showing graded bedding in Alligator Back metagraywacke on Town Creek one mile north of Henry (Bassett quadrangle). Note channel cut into top of the massive rock now filled with pebble conglomerate (lower left and lower right corners of photo). A claystone pebble that is recessed due to differential weathering (note black shadow at base) to the upper right of coin.

osits. These metaconglomerates are probably deposits of distributary channels on the inner to mid part of a submarine fan at or near the mouth



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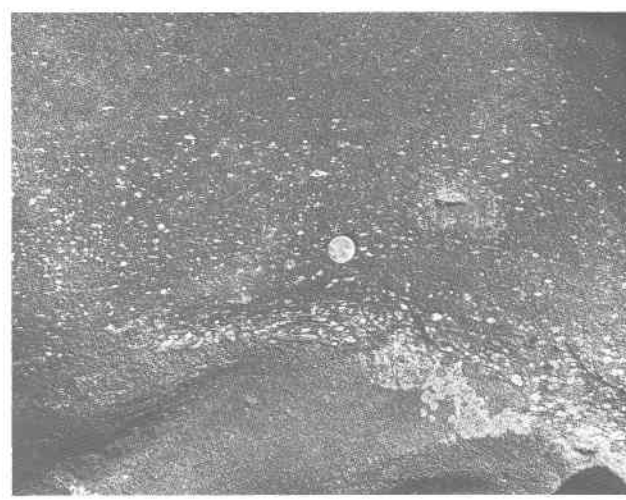

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osits. These metaconglomerates are probably deposits of distributary channels on the inner to mid part of a submarine fan at or near the mouth of a submarine canyon (Middleton and Hampton, 1973; Walker, 1978).

The metagraywackes are composed of quartz, microcline, perthite, plagioclase, muscovite and biotite. Other minerals in accessory amounts less than 10 percent) include epidote, sphene, ilmenite, calcite, garnet, chlorite, zircon and magnetite. Graphite occurs sparingly throughout the metagraywackes as disseminated material and as distinct graphite-bearing-schist layers on top of graded beds. These graphite-rich rocks might represent fine-grained, hemipelagic layers that originally contained organic matter. Blue quartz, a common constituent in the metagraywackes, is also a common constituent of pre-Grenville layered granulites and in Grenville age plutons in the core of the Blue Ridge anticlinorium. If, indeed, these blue quartz clasts in the metagraywackes were derived from an exposed pre-Grenville and Grenville age basement terrane, it would suggest that the source for the sediments which make up the metagraywackes was from the west, off of the North American craton. This is in marked contrast to source direction indicated by a series of cross beds in the previously described Bouma sequence exposed along Town Creek. These cross beds have little or no tectonic overprint and all have a dip to the northwest, indicating a source to the southeast. Both the cross beds and the graded bedding (shown in Figure 9) indicate that the rocks exposed along Town Creek are right-side up.

Mica Schist

Silvery-gray, quartz-mica schists occur as interlayers in the metagraywackes and also form a distinct mappable unit near the base of the Alligator Back Formation around the nose of the Cooper Creek anticline. This mica schist lies directly on and is interlayered with the metabasalts at the base of the Alligator Back.

The mica schists are composed of varying amounts of biotite, muscovite and quartz. They contain fine trains of graphite and rare graphite-rich layers. Other minerals present in accessory amounts are tourmaline, garnet, plagioclase, microcline and ilmenite.

Graphite Schist

Graphite-bearing rocks, ranging from graphite schists to quartz-graphite schists, occur throughout the Alligator Back Formation. Thin graphite schist and quartz-graphite schist bands are interlayered with the metagraywackes and may be in contact with metabasalts locally.

The graphitic schists are gray to grayish black and are composed of varying amounts of quartz, muscovite and graphite and range from a quartz-rich rock that has a flaggy foliation and a pronounced pencil-rodding (Figure 11) to mica-rich rock with a well-developed schistosity. Locally these quartz-rich and mica-rich graphitic rocks are interlayered. Graphite schists are composed of up to 50 percent quartz and varying amounts of graphite and mica. Pyrite is abundant in these rocks. The more schistose rocks have proportionately larger amounts of mica. Accessory minerals include chlorite, biotite, ilmenite, and garnet. Leucoxene occurs as an alteration of ilmenite. Thin calcareous zones and thin discontinuous quartzite beds occur in the graphite schist in the southeastern part of Fairy Stone State Park (Conley and Henika, 1970).

The gray graphite schists were likely hemipelagic sediments containing organic matter that were deposited in a restricted basin under reducing conditions.

Marble

White, pinkish-orange, and gray fine to coarse crystalline marbles occur interlayered with me-

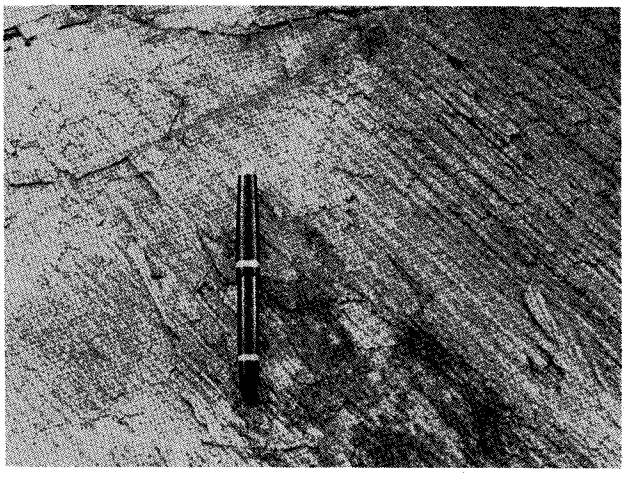

Figure 11. Graphitic quartz schist that occurs in Alligator Back Formation at entrance to Ryans Branch Campsite, north shore of Philpott-Reservoir, just northeast of Union Church Bridge (Philpott Reservoir quadrangle). A pronounced rodding is produced by intersection of transposed bedding and schistosity. Kink band is located just above pen.

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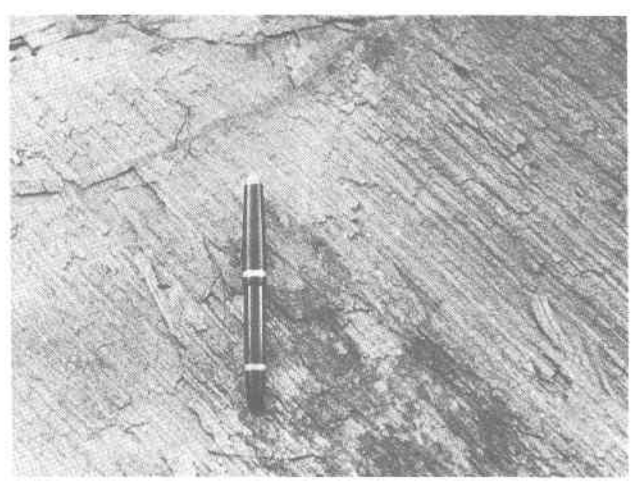

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The presence of marbles containing numerous thin quartz-bearing interlayers and mica schist partings in a metagraywacke sequence is anomalous. The marbles could be of clastic origin and represent reworked shelf carbonates deposited as part of a turbidite sequence.

Quartzite

Impure micaceous quartzites occur as interlayers in the metagraywackes and graphite schists of the Alligator Back Formation. These quartzites are composed primarily of quartz, with lesser amounts of muscovite, biotite and accessory amounts of microcline, plagioclase, magnetite, sphene, ilmenite, zircon and apatite. Quartz-bearing layers are commonly separated by thin sericite layers. Graphite schist fragments and biotite films occur along foliation surfaces. Thin, cross-bedded, isoclinally folded quartzite beds were observed as interbeds in the graphite schists along State Route 58 at the eastern boundary of Fairy Stone State Park (Figure 13) and in a metagraywacke unit in exposure along the right-of-way of the Norfolk and Western Railway along Town Creek (at the Franklin-Henry County boundary on the northwestern edge of the Philpott Reservoir quadrangle).

A different quartzite occurs in contact with, and probably stratigraphically overlies, the metagabbros on the northwest side of Fairy Stone Lake in Fairy Stone State Park (Figure 14). This massive quartzite is a white to yellowish-gray

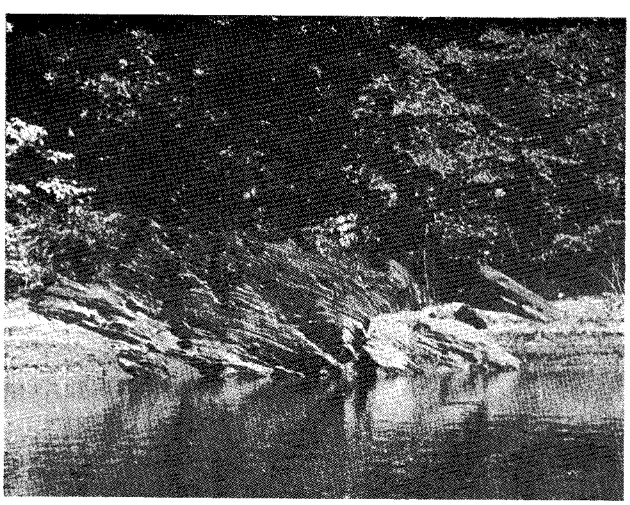

Figure 12. Finely laminated, impure marble on southeastern limb of the Otter Creek anticline, south shore of Philpott Reservoir on Rennett Bag Creek at boat ramp, 0.5 mile NW of the Smith River (Philpott Reservoir quadrangle). Quartzose layers are accentuated by differential weathering.

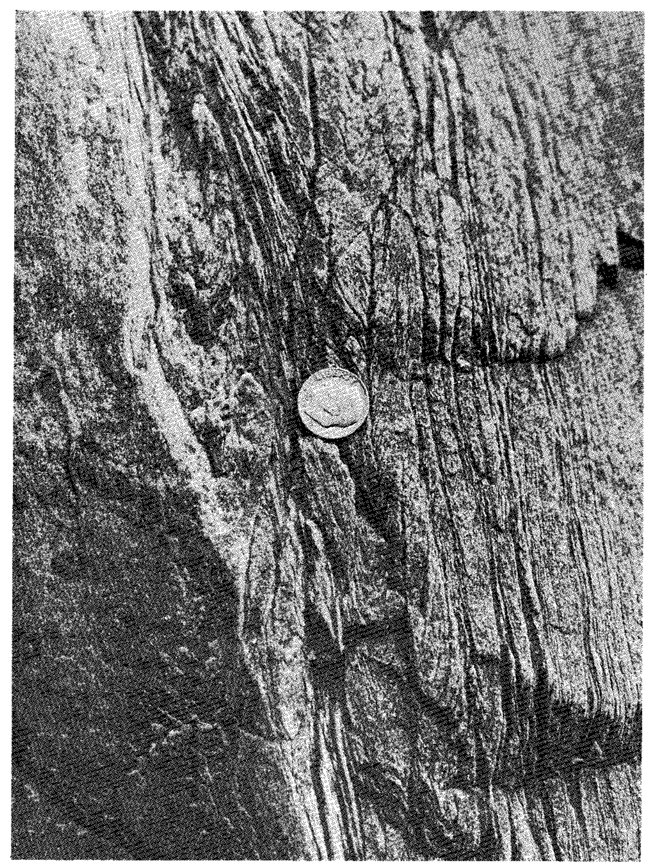

Figure 13. Thinly laminated quartzite interlayered with graphite schists on State Highway 58, on the eastern boundary of Fairy Stone State Park (Philpott Reservoir quadrangle). Early set of isoclinal folds F_1 plunge steeply toward viewer, and are cut by spaced cleavage lying parallel to axial surfaces of a later set of isoclinal folds (probably F_2) that have subhorizontal hinge surfaces. Fine crossbeds are present but not apparent.

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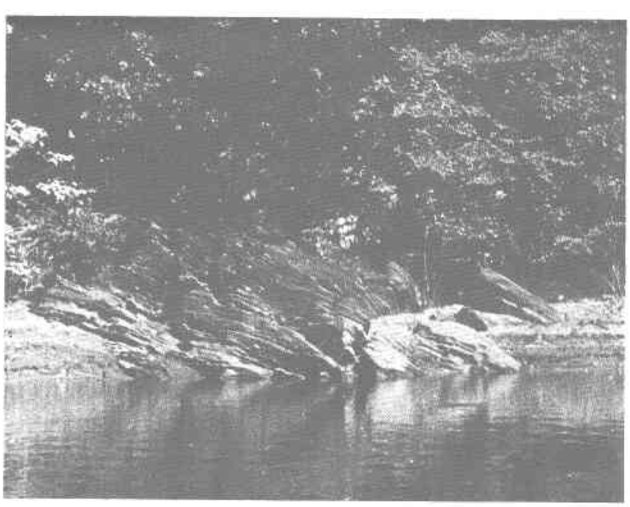

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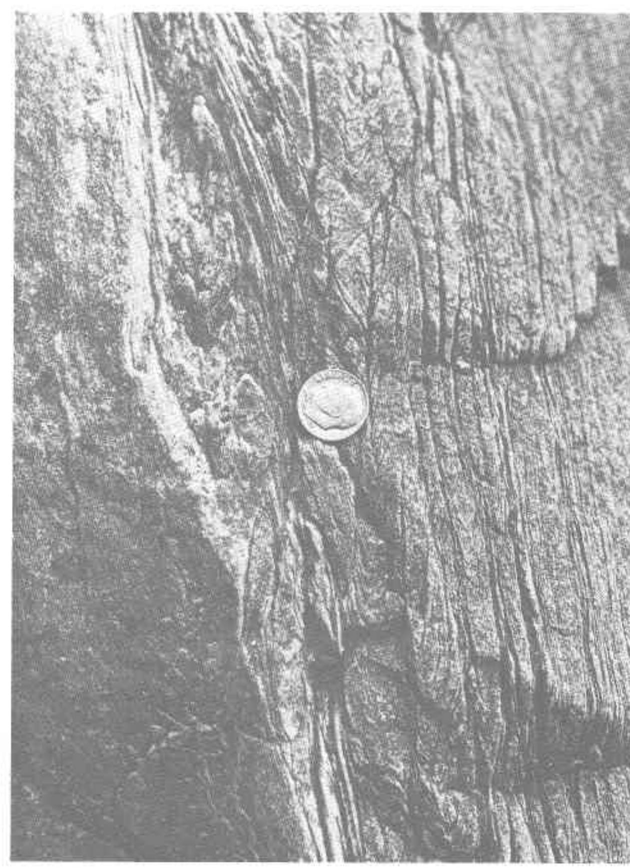

Figure 13. Thinly laminated quartzite interlayered with graphite schists on State Highway 58, on the eastern boundary of Fairy Stone State Park (Philpott Reservoir quadrangle). Early set of isoclinal folds F_1 plunge steeply toward viewer, and are cut by spaced cleavage lying parallel to axial surfaces of a later set of isoclinal folds (probably F_2) that have subhorizontal hinge surfaces. Fine crossbeds are present but not apparent.

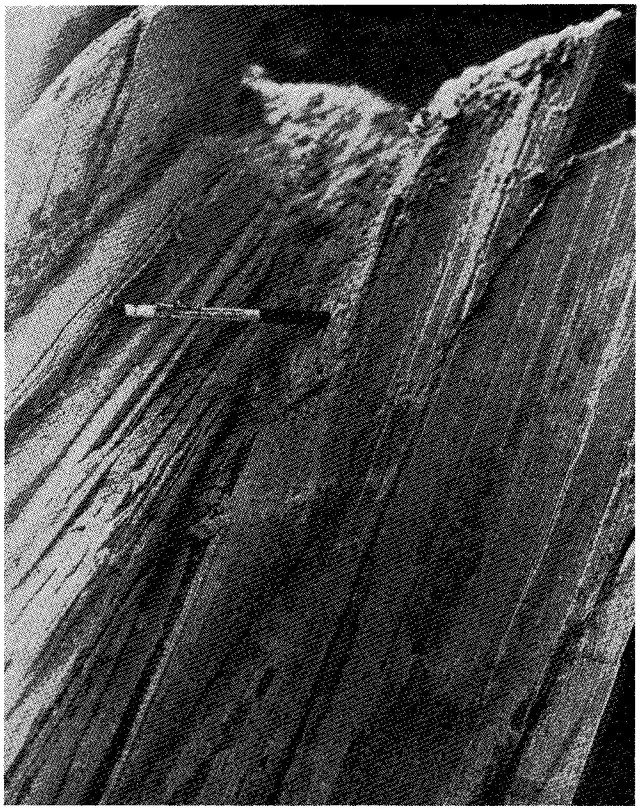

Figure 14. Fine-grained massive quartzite showing fine foliation planes (at top of pen). Locality is on north side of Fairy Stone Lake directly across from the swimming beach, Fairy Stone State Park (Philpott Reservoir quadrangle).

rock that is composed of very-fine-grained saccharoidal quartz with sphene as the single accessory mineral. It has a rude tectonic foliation but all primary textures, such as cross-bedding and relict clasts, are absent. Similar quartzites containing abundant magnetite, occur in association with metagabbros and metabasalts southwest of Rocky Mount.

The cross-bedded quartzites interlayered with metagraywackes and the massive quartzites associated with metagabbros are probably of different origin. The thin, fine-grained quartzites that overlie the metagabbros west of Rocky Mount and along the northwest shore of Fairy Stone Lake in Fairy Stone State Park might represent recrystallized volcanogenic cherts, because they are so extremely massive and do not contain recognizable primary structures. They are associated with the mafic igneous rocks and could be at least in part precipitates from waters enriched in silica by volcanic and associated fumerolic emanations.

STRUCTURAL RELATIONSHIPS

The Alligator Back Formation contains at least two and possibly three intervals of metabasalt—each is locally underlain by either metagabbro or metagabbro and ultramafic rock. These metabasalt intervals are (1) rocks at the base of the formation, which lie directly upon the Ashe Formation in the core of the Cooper Creek anticline and (2) a higher interval which makes up the thin, basal metagabbro and thick metabasalt sequence south of Ferrum. The metagabbro-metabasalt sequence at Rocky Mount is possibly higher in the section and thus (at least structurally) overlies the thick metabasalt sequence south of Ferrum (Figure 4). Grassy Hill seems to be formed of an isolated, ultramafic-metagabbro mass. The isolated mass of ultramafic rocks, metagabbros and metabasalts, which makes up Brier Mountain, overlie plagioclase gneiss of the Ashe Formation in the core of the Cooper Creek anticline and could represent a klippe from the base of the Alligator Back. The suite of rocks that compose Brier Mountain is quite similar to rocks that compose Jacks Mountain, which are in the base of the Alligator Back and also overlie the Ashe Formation.

The repeated sequences of mafic and ultramafic metaigneous rocks overlain by metasedimentary rocks within the Alligator Back seem to suggest that they are made up of imbricate fault slices. Many of the sequences contain only metabasalts or metabasalts underlain by discontinuous layers of metagabbro, which suggests that these sequences are incomplete. The previously mapped (Conley and Henika, 1970) shear zones of varying thicknesses developed within and at the base of these metaigneous sequences supports this conclusion. The lower part of the metabasalt-metagabbro-ultramafic sequence that lies at the base of the Alligator Back is sheared, as is the top of the underlying Ashe Formation. Shear zones are also observed at the base of most of the ultramafic-metagabbro-metabasalt sequences that occur in stratigraphic positions above the base of the Alligator Back.

An idealized package of a single fault slice would consist of, from bottom to top: ultramafic rock; gabbro, in many places cut by mafic dikes; quartzite (metachert, probably a precipitate from waters enriched in silica by volcanic emanations); metabasalt containing metagraywacke interlayers at the top; and metagraywacke, locally containing interlayered marbles and thin graphitic quartz-mica schist layers (hemipelagic

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EVIDENCE FOR AN OPHIOLITE SEQUENCE

Definition and Description of an Ophiolite

An ophiolite was defined by the Penrose Conference on ophiolites (anonymous, 1972) as a distinctive assemblage of mafic and ultramafic rocks that should be used neither as a rock name nor as a lithologic unit. An ophiolite sequence is composed of (from bottom to top): (1) an ultramafic sequence composed of harzburgite, lherzolite and dunite that may show a tectonic fabric and be serpentinized; (2) a gabbro sequence that is generally less deformed than the underlying ultramafic rocks; the gabbros may contain cumulate bodies of peridotite and pyroxenite; (3) a mafic sheeted-dike complex; (4) a mafic volcanic sequence, commonly containing pillow basalts; (5) an overlying sedimentary section composed of ribbon cherts, shale interbeds, minor amounts of limestone, podiform chromite bodies associated with dunite and sodic felsic intrusive and extrusive rocks. Faulted contacts between mappable units are common and produce incomplete, dismembered and metamorphosed ophiolite sequences. Ophiolites are considered to be derived from oceanic crust and upper mantle rocks, but the term is descriptive and not genetic. The stratigraphic sequence of metaigneous rocks overlain by metasedimentary rocks that is repeated by imbricate faults previously described in the Lynchburg Formation closely resembles the ophiolites and their sedimentary cover rocks, as defined by the Penrose Conference on ophiolites.

Geochemical Evidence

To further test the hypothesis that the ultramafic-metagabbro-metabasalt sequences in the Lynchburg are ophiolites, three metagabbro samples and four metabasalt samples were analyzed for major and trace elements. Major elements analyses were made by Oliver M. Fordham, Jr., Virginia Division of Mineral Resources. Analyses of the trace elements Nb, Ar, Y, Sr, Rb, and Ba were made by David Whittington, Department of Geology, Florida State University. The results of these analyses are presented in Table 1.

Pearce and Cann (1971, 1973) indicate that basaltic eruptions in different geological environments have different chemical compositions. The rocks that can be differentiated by this method are: ocean-floor basalts, low-potassium tholeiites, calc-alkalin basalts, shoshonites, ocean-island basalts and continental basalts. The chemistry of ophiolitic basalts is not significantly different from ocean-floor basalts (Pearce and Cann, 1971), which supports the widely held theory that ophiolites are obducted oceanic crust and overlying ocean-floor basalts that were originally deposited along spreading centers of oceanic basins.

A difficulty arises in comparing metamorphosed ophiolites with ocean-floor basalts because elements migrate differentially during metamorphism. Pearce and Cann (1973) found that the elements titanium, zirconium and yttrium have low mobility and therefore are relatively insensitive to low grade metamorphism. For this reason, these elements were chosen by them for determining magma types in metamorphic rocks that do not exceed upper greenschist or lower amphibolite facies.

A comparison of samples analyzed (Figure 15) shows that they follow the trend of ocean-floor basalts and samples 5 and 7 lie in this field. Samples 1 and 3 (and to a lesser extent sample 2) are much lower in titanium than samples plotted by Pearce and Cann (1971). Data in Figure 16, a triangular discrimination diagram for Ti

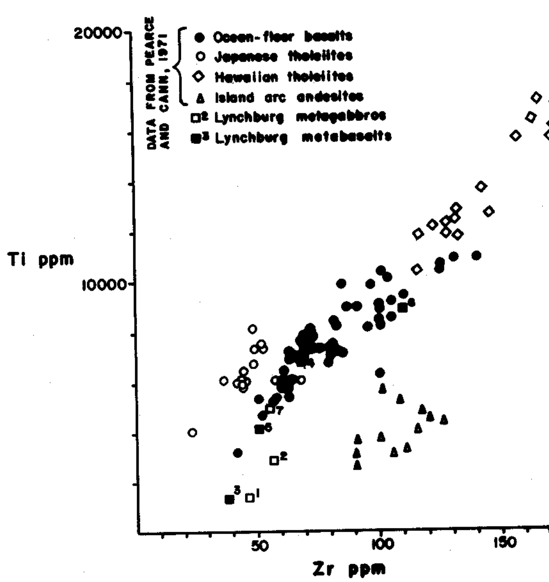

Figure 15. Ti-Zr plot for parental magma types.

Table 1. Chemical Analyses of Mafic Rocks from Alligator Back Formation

	1	2	3	4	5	6	7
	Metagabbro	Metagabbro	Metabasalt	Metabasalt	Basal Metabasalt	Metabasalt	Metagabbro
SiO_2	46.70	42.36	50.25	46.54	46.95	49.27	50.30
Al_2O_3	29.32	5.68	13.83	16.15	13.59	15.19	13.47
Fe_2O_3	5.15	13.79	11.64	13.66	15.88	11.46	14.70
MgO	8.21± .082	27.33± .642	8.94	8.21± .082	8.91± .175	6.84± .07	7.15± .115
CaO	16.79	4.19	11.39	8.05	10.00	10.59	9.42
Na_2O	1.38± .023	0.09± .002	1.99	3.12± .060	2.60± .033	3.71	2.65± .048
K_2O	0.06	0.01	0.17	1.04	0.20	0.22	0.27
TiO_2	0.26	0.49	0.26	1.21	0.68	1.53	0.83
MnO	0.10	0.20	0.19	0.31	0.22	0.19	0.23
P_2O_5	0.00	0.02	0.02	0.06	0.01	0.15	0.07
Cr	0.07	0.26	0.04	0.01	0.01	0.03	0.01
Co	0.00	0.01	0.00	0.01	0.01	0.00	0.00
Ni	0.01	0.12	0.01	0.01	0.01	0.01	0.01
Ba	0.02	0.01	0.01	0.04	0.01	0.01	0.02
Sr	0.025	0.002	0.007	0.018	0.003	0.034	0.006
Rb	0.02	0.02	0.02	0.03	0.02	0.02	0.02
PPM	-	-	-	-	-	-	-
Nb	16.9± 3.7	7.4± 5.5	3.5± 2.0	12.7± 0.7	5.8± 5.8	13.0± 1.6	2.3± 2.3
Zr	48.2± 2.9	55.5± 3.5	38.6± 2.3	69.7± 0.2	50.1± 5.4	110.5± 2.0	54.5± 3.4
Y	20.3± 1.6	25.0± 3.3	22.6± 3.1	25.6± 2.5	26.4± 5.2	29.7± 4.1	31.2± 1.0
Sr	198.1± 6.7	100.9± 1.3	230.7± 3.1	152.0± 3.7	11.8± 3.4	235.0± 2.0	104.1± 1.2
Rb	8.7± 3.9	10.0± 0.7	7.0± 2.8	29.9± 3.5	4.1± 2.6	6.3± 1.1	9.6± 1.7
Ba	81.5± 1.1	72.8± 0.5	63.2± 0.2	297.7± 2.8	104.7± 4.0	217.2± 12.7	145.6± 9.4
Ti	1558.70	2937.55	1558.70	7253.95	4076.60	9172.35	4975.85
LOF	1.88	5.95	1.33	1.64	1.04	.86	0.97
Na_2O/K_2O	23	9	11.71	3	13	16.87	9.81
Total Fe as FeO	4.63	12.41	10.47	12.29	14.29	10.31	13.23

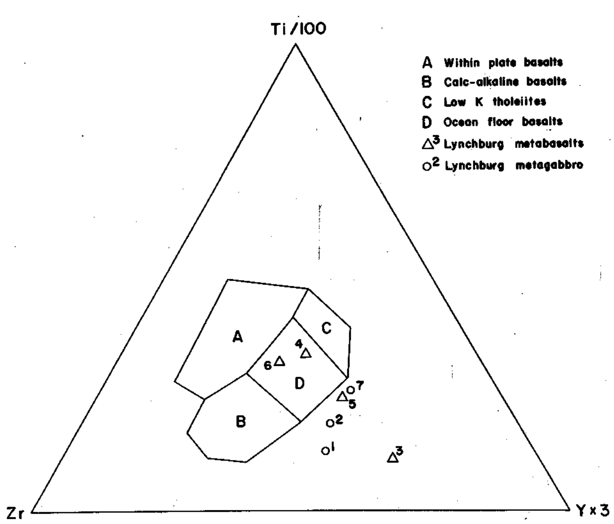

Figure 16. Discrimination diagram for Ti, Zr and Y.

- Zr - Y, show that two of the samples fall within the field of ocean-floor basalts; five samples plot to the lower right of this field because of high yttrium and low titanium content, but these samples are much closer in composition to ocean floor basalts than to other basalt fields of the diagram. This data is not conclusive but does indicate an affinity to ocean floor basalts. To be more definitive, a larger number of analyzed samples would be required. In addition, the effect of lower amphibolite grade metamorphism on some of these samples must also be determined.

In addition to the chemical analyses of the mafic rocks, 45 alluvial sediment samples were collected by McCollum during his field mapping of the Rocky Mount quadrangle. These samples were collected from streams issuing from Grassy Hill, Brier Mountain and the ultramafic-metagabbro-metabasalt sequences at the southern edge of the City of Rocky Mount. The samples were analyzed by O. M. Fordham, Jr., Virginia Division of Mineral Resources, for the elements Cr, Ni, Co, Cu, Pb, Li, Mn, Fe, Sr, Ag and Cd. Fourteen of the samples analyzed contained anomalously high values in one or more of the elements selected for analysis (Table 2). The elements Cr and Ni normally occur in the ultramafic rocks of an ophiolite suite (Coleman, 1977). The elements Cu, Pb, Zn, and Ag, reported from near Rocky Mount (Fontaine, 1883) and at the north end of Grassy Hill, just north of Rocky Mount (McCollum, in preparation), commonly occur in association with ophiolites. These elements originate in massive sulfide deposits that were originally deposited with the basalts along a mid-oceanic ridge (Coleman, 1977). In addition, there are massive sulfide bodies in the Lynchburg Group of the Great Gossan Lead district along strike to the southwest (Stose and Stose, 1957). Thus, the stratigraphic, structural and chemical evidence supports the theory that the layers of ultramafic-metagabbro-metabasalt rocks in the upper formation of the Lynchburg Group are partial sequences of ophiolites.

Table 2. Geochemistry of Stream Sediments from the Rocky Mount Quadrangle

by Oliver M. Fordham, Jr.

Forty five 80-230 mesh stream sediment samples from the Rocky Mount quadrangle were analyzed by atomic absorption spectrophotometry using a selective, hot, hydrochloric acid method of attack. Compared with many other stream sediment samples taken from the Virginia Piedmont and analyzed in like manner, there are a number of anomalous samples in the Rocky Mount quadrangle. The table presented below shows values for samples which have concentrations above the threshold (i.e. mean plus two standard deviations) for stream sediment samples collected from the Virginia Piedmont. Any samples showing Ag above 1 ppm or Cd above 0.5 ppm are anomalous. No anomalous samples were detected for the elements - Li, Mn, Fe, or Sr.

Repository Number	Element Concentration (ppm)							
	Cr	Ni	Co	Cu	Pb	Zn	Ag	Cd
R-7385	-	146	39	19	-	75	-	-
R-7386	-	118	28	-	19	32	-	1.6
R-7387	-	75	20	-	-	-	-	-
R-7388	-	-	-	15	23	-	-	-
R-7389	-	50	20	14	55	92	67	2.6
R-7390	-	44	-	15	30	46	60	-
R-7391	-	-	26	-	18	-	-	-
R-7392	-	-	-	13	32	-	-	-
R-7393	-	231	31	14	-	-	-	-
R-7394	-	218	41	-	-	-	-	-
R-7395	-	97	18	-	-	-	-	-
R-7396	-	47	-	16	20	-	-	-
R-7397	-	-	-	-	19	-	-	-
R-7398	-	62	-	-	-	-	-	-
Piedmont Values								
Mean		16	8	6	6	9	27	-
Deviation	14	5	4	6	4	13	-	-
Std. Threshold	44	18	13	17	16	54	1.0	0.5

TIME OF OBDUCTION AND EMPLACEMENT OF THE LYNCHBURG OPHIOLITES?

Fullagar and Dietrich (1976) determined rubidium-strontium whole-rock ages for 46 samples of metasedimentary rocks from the Lynchburg. The dates from these rocks fall into two age ranges: 520-583 m.y. and 350-420 m.y. The 520-583 m.y. age is interpreted by Fullagar and Dietrich (1976) as a time of dewatering and metamorphism of the Lynchburg sediments, whereas the younger age is equated to a later mid-Paleozoic metamorphic event.

Northeast along strike in the central and northern Virginia Piedmont and Blue Ridge the obducted Lynchburg is overlain by continental basalts belonging to the Catoctin Formation (Reed and Morgan, 1971; Bland, 1978). This would suggest that the Lynchburg ophiolites and metasedimentary cover rocks were emplaced prior to deposition of the Catoctin. These Catoctin lavas were probably extruded during the initial stages of the opening of the Iapetus Ocean. A discordant lead-uranium age of 820 m.y. was obtained from zircons in rhyolite interlayered with Catoctin Formation rocks in southern Pennsylvania (Rankin and others, 1969). This date is probably too old (Williams and Stevens, 1974; Lukert, 1981; Mose, 1981). The Robertson River Granite of central Virginia, which possibly intrudes the Lynchburg (Mose, 1981) and is itself cut by Catoctin-like metabasalt dikes, was dated at 700 m.y. by use of the lead-uranium method from zircons (Lukert, 1981) and at 550 and 580 m.y. by use of the rubidium-strontium whole-rock method (Mose, 1981). These radiometric age dates would bracket a time between 700 m.y. and 550 m.y. when the Lynchburg was already in place on Grenville basement in central Virginia.

In summary, the available evidence suggests that the Lynchburg ophiolites and metasedimentary rocks were deposited in late Precambrian time, prior to the 520-583 m.y. period of dewatering and metamorphism of the unit proposed by Fullagar and Dietrich (1976). Lynchburg sediments and underlying pieces of ocean floor were obducted onto the continental margin (which is now exposed on the southeastern limb of the Blue Ridge anticlinorium) during a period of plate closing. This orogenic event previously unreported in the rocks of the Blue Ridge Province occurred in late Precambrian or earliest Cambrian time in an interval before the outpouring of the Catoctin continental basalts which were generated in the beginning stage of the breaking apart of continents to form the Iapetus Ocean. Although not previously recognized in the Blue Ridge Province, evidence of a late Precambrian-early Cambrian orogenic event is documented in rocks from both the northern and south-central Virginia Piedmont (Glover and Sinha, 1973; Drake 1983; Drake, this volume). The presence of the Lynchburg almost continuously along the eastern limb of the Blue Ridge anticlinorium from northern Virginia to Georgia indicates the magnitude of this late Precambrian-early Cambrian event.

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GEOLOGY OF THE MOBLEY MOUNTAIN GRANITE PINEY RIVER QUADRANGLE, VIRGINIA BLUE RIDGE

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CONTENTS

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ILLUSTRATIONS

TABLES

ABSTRACT

The Mobley Mountain granite of the Blue Ridge of central Virginia occupies an area of about 2 by 4 km. It is a two-mica, alkaline granite, which is medium grained and slightly foliated, with associated aplitic and pegmatitic dikes, and intrudes the Grenville age Turkey Mountain pluton. It is subsolvus and contains principally quartz (36%), microcline (25%), and albite (17%); biotite ranges from 7-17% and muscovite up to 2%. Accessory minerals include epidote, florencite, zircon, sphene, apatite, fluorite, and garnet.

The pluton is a relatively deep mesozonal type, and may have a primitive source in the uppermost mantle. Its chemical characteristics are those of an alkali granite, with an initial Sr isotopic ratio of 0.7045, and a mean K/Rb ratio of about 430. A gravity model suggests a diapiric body reaching 7.2 km in depth; emplacement may have been at about 18 km. The Rb/Sr age of 652 ±32 m.y. corresponds well with the tin-bearing granite at Irish Creek, which has an $^{39}Ar/^{40}Ar$ plateau age of about 635 m.y.

The Mobley Mountain granite may have been emplaced in a rifting continental block, with minimal crustal involvement, by differentiation of upper mantle derived basalt at depths below that of plagioclase stability. It is one of a line of latest Precambrian plutons of similar chemical affinities and origin that extends from southeastern New York to North Carolina.

INTRODUCTION

This paper presents the field relations, petrology, and Rb-Sr whole-rock age of the Mobley Mountain granite, located in the Roseland district of central Virginia about 25 km north of Lynchburg, in the southwestern part of the Piney River 7.5-minute quadrangle (Figure 1). The granite crops out over an area of about 2 by 4 km, underlying Mobley Mountain with elevations of over 400 m and relief up to about 150 m. The rock is medium grained, and shows slight foliation which is more pronounced towards its margins. The border relations of the Mobley Mountain granite are characterized by small-scale anatectic mobilization within the mylonitic biotite gneiss which it intrudes, and a 2 to 3 km halo of abundant related aplitic and pegmatitic dikes. Microcline perthite, plagioclase (An_{1-5}), quartz, biotite, and clinozoisite-epidote are the major minerals. Accessories include muscovite, florencite (a rare-earth phosphosilicate), sphene, apatite, fluorite, and in places, almandine-grossularite garnet. Just to the south of Mobley Mountain, perrierite, $(Ce, Ca, Th)_2(Ti, Fe)_2Si_2O_{11}$, is found together with other rare earth minerals in a deeply weathered pegmatite (Geitgey, 1967).

The Mobley Mountain granite possesses the characteristics of relatively deep mesozonal plutons (Buddington, 1959): contact aureole, chilled margins, and miarolitic structures lacking; minor assimilation and anatexis restricted to the border zone; abundant aplites extending into the country rock; and rough concordance to the regional trend. The presence of magmatic epidote suggests pressures ($P_t=P_{H_2O}$) greater than 2 kb and up to 8 kb (Zen and Hammerstrom, 1982).

Recent models of the evolution of the Appalachian orogen include the formation of a proto-Atlantic or Iapetus Ocean by continental fragmentation during the late Precambrian (Bird and Dewey, 1970; Rankin, 1975). According to Rankin (1975), this lateral expansion was concurrent with the onset of emplacement of an anorogenic, bimodal plutonic-volcanic group whose silicic members have peralkaline affinity. Although studied in northwestern North Carolina (Rankin, 1975), this rock suite may be recognized in the Precambrian terrans of central Virginia (Figure 2). Along the southeastern flank of the Grenville and older Blue Ridge anticlinorium, the late Precambrian to early Paleozoic volcanic-volcanoclastic sequences are locally intruded by granitic plutons (Rodgers, 1972). In central Virginia and to the southwest, this sequence includes the Catoctin and Mount Rogers formations; the granitic plutons with peralkaline affinities, include the Robertson River, Beech, Crossnore, Striped Rock (Rankin, 1975), and Mobley Mountain.

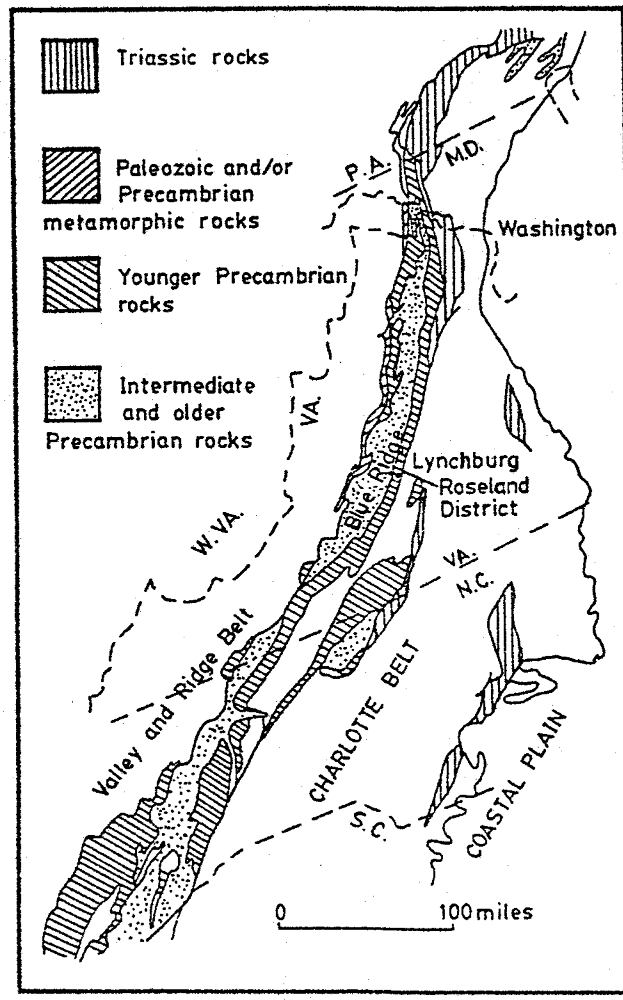

Figure 1. Generalized geological map of the Blue Ridge of Virginia and adjacent states (after Rankin and others, 1973).

GEOLOGIC SETTING

The Mobley Mountain granite lies on the southeastern flank of the basement core of the Blue Ridge anticlinorium, in the southwestern part of the Roseland district (Herz and Force, 1984), in central Virginia (Figure 2). In central Virginia, the Blue Ridge is physiographically a single mountain range 10 to 20 km across, that reaches elevations of about 1200 m (Rodgers, 1972). A variety of Grenville and pre-Grenville plutonic gneisses, ranging in composition from granitic to charnockitic, forms the basement core of the anticlinorium, unconformably overlain on the northwest and southwest limbs by younger Precambrian metasediments and metavolcanics (Espenshade, 1970).

Grenville and older basement rocks, including anorthosite, granulite, Roses Mill and Turkey Mountain ferrodiorite plutons, and the Pedlar charnockite underlie the Roseland district (Herz and Force, 1984). Leucocratic charnockites, mangerites, and layered granulites constitute the oldest lithologies, and crop out in the southeast half of the district. The Roseland anorthosite cuts these older rocks; this sequence, in turn was intruded by ferrodiorites and charnockites about 1,040 m.y. ago. The Grenville deformational event followed this intrusive cycle. A northeast-trending ductile deformation zone extends across the northwest part of the district, separating the Roses Mill and Turkey Mountain plutons in the southeast from the Pedlar massif to the northwest (Figure 2). The zone may have its origin in Grenville time, reactivated in Paleozoic deformations.

PETROLOGY OF THE COUNTRY ROCKS

Several lithologies of Grenville and pre-Grenville age described in the Roseland district (Herz and Force, 1984) crop out in the vicinity of Mobley Mountain. Only the Turkey Mountain pluton is intruded by the Mobley Mountain granite.

Feldspathic gneiss

Associated with granulites of the Roseland district are massive feldspathic gneisses containing blue quartz, mesoperthitic K-feldspar, plagioclase, ortho- and clinopyroxenes, biotite, and traces of ilmenite, apatite, and zircon. These rocks are among the oldest in this region and predate the Grenville orogeny, during which they underwent granulite facies metamorphism. Paleozoic greenschist metamorphism later retrograded much of the feldspathic gneiss.

Feldspathic gneiss crops out northwest, north, and southeast of the Mobley Mountain granite and exhibits a range of cataclastic deformation. The degree of schistosity generally increases towards the northwest as a ductile deformation zone is approached. Where least deformed, this lithology is nearly massive, possessing the textural features of a cataclasite (Higgins, 1971) with granulated quartz fragments about 0.5 mm in size. Otherwise, quartz is anhedral, polycrystalline, and ranges up to 5 mm in size. Slightly sericitic mesoperthitic K-feldspar up to 4 mm, plagioclase up to 2.5 mm with inclusions of white mica, and clinozoisite, biotite, opaques, and epidote-clinozoisite make up the remaining mineralogy.

Shaeffer Hollow Granite

The Shaeffer Hollow Granite is a coarse-grained leucocratic rock with tabular feldspar phenocrysts up to 5 cm in length and predates the Turkey Mountain pluton. The granite is composed of blue quartz, perthitic microcline, antiperthitic plagioclase, and biotite, with trace amounts of ilmenite, apatite, and zircon (Herz and Force, 1984).

A cataclastic version of this lithology occurs along the southeastern contact of the Mobley Mountain granite as a leucocratic mylonite quartz gneiss. The gneiss contains large (3 to 4 cm) augen of recrystallized quartz (0.1 to 3 mm), perthitic microcline (up to 2.5 cm), and kink-banded plagioclase (0.5 to 3 mm); biotite, associated with epidote, sphene, garnet, and opaques, defines a fluxion structure.

Turkey Mountain Pluton

The Turkey Mountain pluton crops out extensively in the southeastern portion of the Roseland district. Just to the northeast is the Roses Mill pluton, which is related to the Turkey Mountain pluton and about 960 m.y. in age. Both plutons are of Grenville age (Herz and Force, 1984) and both contain xenoliths of the older granulite gneiss and anorthosite (Force and Herz, 1979). The Turkey Mountain pluton occurs in two major phases: massive to well-layered quartz mangerite and charnockite, and well-foliated biotite mylonite gneiss. The mylonite gneiss may have been derived by retrogression from the original charnockite; pyroxene was replaced by biotite. Biotite mylonite gneiss forms the country rock into which the Mobley Mountain granite was intruded, and contains abundant Mobley-derived aplitic dikes.

Turkey Mountain charnockite in this area is hypidiomorphic granular. Quartz, perthitic microcline, plagioclase (andesine antiperthite), uralite, biotite, sericite, apatite, ilmenite, and sphene are the dominant minerals. Ortho- and clinopyroxenes are rare, and are rimmed by uralite, biotite, and chlorite. This, plus common highly sericitized plagioclase, is evidence for widespread retrogradation of the orginal igneous

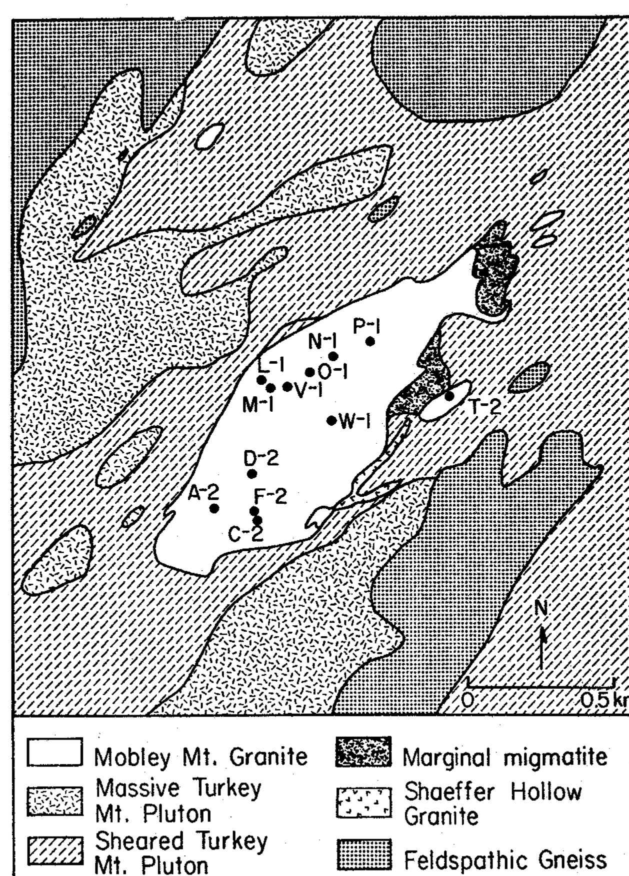

Figure 2. Geological and sample location map, southwestern part of the Roseland district, Virginia (after Brock, 1981; Herz and Force, 1984).

mineral assemblage. Quartz in the massive phase is anhedral, with undulatory extinction, and ranges from 0.1 to 5 mm in size. Plagioclase grains (An_{30-40}) are commonly fractured, anhedral, highly sericitic, twinned, and 1 to 3.5 mm. Perthitic microcline is anhedral and ranges in size from 1 to 5 mm. Pyroxenes are corroded and embayed, with maximum grain sizes of about 1.5 mm. Ilmenite, 0.3-1.5 mm in size, is mantled by sphene.

Biotite mylonite gneiss is the most common lithology of the Turkey Mountain pluton in this area. The rock possesses a blastomylonitic texture; polycrystalline quartz, kink banded, sericitized and saussuritized perthitic microcline, and plagioclase (An_{0-20}), biotite, epidote, sphene, apatite, and ilmenite are the dominant minerals. Augen of feldspar and granular quartz, encased in biotite oriented along the principal foliation plane, are characteristic.

PETROLOGY OF THE MOBLEY MOUNTAIN GRANITE

The Mobley Mountain granite is white to light yellow with subparallel black flecks. It is subsolvus, fine to medium grained (0.01 to 2 mm), with a hypidiomorphic granular to poorly foliated texture and is dominantly composed of quartz (36 percent), slightly perthitic microcline (25 percent), and albite (17 percent). Biotite is the principal mafic mineral (7 to 17 percent) and is accompanied by as much as 2 percent muscovite. Characteristic accessory minerals include epidote, sphene, apatite, florencite ($CeAl_3(PO_4)_2(OH)_6$), zircon, garnet, and fluorite (Brock, 1981).

Quartz

Quartz is anhedral and polycrystalline, forming aggregates of grains up to 2.8 mm in size at the center of the body, with individual grains ranging from 0.2 to 0.8 mm. Along the margins of the body, quartz aggregates reach only about 1.2 mm and are generally composed of single grains less than 0.1 mm. Grain boundaries are irregular, and strongly undulose extinction is universal. Quartz inclusions about 0.2 mm are seen in places within the perthitic microcline grains.

Plagioclase

Plagioclase forms unzoned, twinned crystals within the body up to 1.6 mm; near its boundaries generally less than 0.8 mm. Most grains are subhedral and saussuritized.

Microcline

Microcline is the coarsest grained mineral phase, reaching 7 mm at the center and 1.6 mm at the margin of the pluton. Crystals are anhedral to subhedral, with quadrille twinning and stringers of exsolved albite; many grains have irregular boundaries. Sericitized plagioclase, biotite, quartz, and white mica are common inclusions.

Biotite

Biotite is the dominant mafic phase, constituting up to 17 percent as coarse subhedral flakes, commonly with subparallel alignment. Pleochroism is from olive to yellowish green; inclusions of zircon with pleochroic halos and rutile needles are common. At the center of the body, biotite occurs as unaligned patchy masses (0.5 to 2.5 mm), associated with subhedral grains and partially altered to chlorite. Nearer the contact, biotite flakes up to 0.8 mm are well aligned as a result of the strong shear deformation and define the principal foliation plane of the granite. Sphene is also associated with biotite, forming anhedral inclusions in places.

Muscovite

Primary muscovite was found in all thin sections, up to 2.2 percent, but always in amounts subordinate to biotite. The muscovite commonly occurs as thin tabular flakes up to 0.5 mm, intergrown with biotite. A close association with epidote is also seen and small (0.01 mm) epidote inclusions are common. Secondary sericite, 0.01-0.05 mm in diameter, occurs abundantly as an alteration product of plagioclase and ranges up to 4.4 percent in the most altered samples.

Chlorite

Minor amounts (up to 1.2 percent) of chlorite occur as an alteration product, commonly as mantles on biotite. Pleochroism is weak, with color ranging from dark green to green.

Epidote

Epidote is an abundant accessory mineral (2.8 to 7.5 percent), generally occuring in patchy intergrowths intimately associated with biotite. Grains are subhedral, up to about 1.5 mm. Inclusions of quartz and biotite are common; less commonly, euhedral and orange florencite crystals core the epidote. Small inclusions of clinozoisite, less than 0.05 mm, occur as plagioclase saussuritization products.

Florencite

Florencite is a basic cerium earth phosphate with substitution possible by numerous elements, so that its ideal chemical formula, $CeAl_3(PO_4)_2(OH)_6$, is only approximate. Color ranges from dark red through pink and olive to dark green, with reddish shades denoting the presence of iron (Geitgey, 1967). A deeply weathered pegmatite at Burley Farm, 0.8 km southwest of the Mobley Mountain granite, was studied by Geitgey (1967). He noted that florencite, the most abundant rare earth phosphate in the rock, occurs as an alteration product of perrierite, a radioactive rare earth titanosilicate. Within the granite, florencite is crimson to orange and commonly cores epidote grains. The largest grains (up to 1.5 mm) are euhedral and appear to be zoned. In plane light, the mineral is turbid with slight pleochroism; under crossed polars, it is dark. It seems unlikely that florencite represents a primary igneous phase in the granite. The secondary relationships seen by Geitgey (1967) at Burley Farm suggest that florencite also formed by deuteric processes in the Mobley Mountain granite.

Garnet

Garnet is present up to 1 percent in most thin sections. Crystals are subhedral to prismatic, from 0.1 to 1 mm and with minute inclusions of quartz (0.01 mm). Grain boundaries are distinct and show no evidence of resorption.

Other Accessory Minerals

Other accessory minerals include apatite, sphene, fluorite, and zircon. Apatite forms small (0.1 mm) prismatic grains; sphene is subhedral to euhedral and up to 1 mm. Fluorite is anhedral, as much as 0.3 mm in size and is colorless, aside from rare purple spots. Zircon occurs as minute inclusions in biotite surrounded by pleochroic halos.

MOBLEY MOUNTAIN LEUCOCRATIC DIKES

Abundant leucocratic dikes cut the border zone of the Mobley Mountain granite and extend for 2 to 3 kilometers within the Turkey Mountain pluton. These dikes apparently developed late in the magmatic history of the granite and range in width from about 3 meters to less than 0.5 cm. Most, however, are less than 0.5 meters in width. They consist of variable amounts of quartz, microcline perthite, and albite. Accessories include biotite (commonly partially altered to chlorite), epidote, and muscovite, plus trace amounts of garnet, florencite, zircon, and sphene.

Within the unsheared portions of the Turkey Mountain pluton, leucocratic dikes are aplitic. Microcline perthite is typically 2 mm, and plagioclase and quartz range between 0.5 and 1.5 mm. Felsic dikes cutting sheared portions of the Turkey Mountain pluton are typically transposed to concordancy with the northeast-trending, southeast-dipping mylonitic fabric of the country rock. This transposition may be observed in the field as felsic dikes pass from unsheared into sheared country rocks, and has resulted in cataclastic textures and attendant reductions in grain size in the felsic dikes.

MARGINAL MIGMATITIC ZONE

Contacts of the Mobley Mountain granite are characterized by a zone of localized anatexis in the Turkey Mountain pluton up to about 300 meters thick. The best exposures of this marginal migmatitic zone are found along the eastern and northern boundaries of the granite, especially along the Buffalo River. Partial melting of Turkey Mountain xenoliths has produced abundant leucosome material that grades into rock visually indistinguishable from the Mobley Mountain granite. Xenoliths are often encased in several centimeters of biotite rich melanosome material. Additionally, a ferrohastingsite-bearing granitic rock occurs as patches in, and interlayered with, leucosome material on a scale ranging from several centimeters to several meters.

Marginal leucosome rock is hypidiomorphic granular to slightly foliated and is predominantly composed of quartz, microcline perthite, and plagioclase. Quartz is anhedral and ranges from 0.05 to 0.5 mm, and microcline perthite, also anhedral, is up to 2 mm. Plagioclase, 0.1 to 2 mm, is subhedral, and normally zoned from An_{10} to An_{20}. Accessories are biotite with zircon inclusions, and epidote. Marginal melanosome is biotite rich with subordinate amounts of quartz, microcline perthite, and albite. Epidote is a common accessory, and trace amounts of apatite, sphene, and ilmenite are seen.

The marginal ferrohastingsite-bearing granitic phase may have formed by reaction of Mobley Mountain magma with the Turkey Mountain pluton. Mineralogically it is quite similar to typical Mobley Mountain granite, except for the presence of greenish blue subhedral to euhedral ferrohastingsite, commonly rimmed by biotite. Texturally it possesses a faint mylonitic foliation. Quartz (0.05 to 1 mm), microcline perthite (up to 0.8 mm), and plagioclase An_{1-2} (up to .05 mm) are the dominant minerals. Accessories include epidote, florencite, garnet, and zircon.

STRUCTURE AND FIELD RELATIONS

The Mobley Mountain granite bears an intrusive relationship to the Turkey Mountain pluton (Figure 2). Both intrude the older Shaeffer Hollow granite and granulites which also form xenoliths in the Mobley Mountain and the Turkey Mountain.

The Mobley Mountain granite appears to core a northeast trending antiform, slightly overturned to the northwest. Evidence for the structural position of the pluton at the core of a fold structure is seen in the repetition of older rock types to the northwest and southeast. The Turkey Mountain pluton, with an intrusive relationship to the pre-Grenville rocks and the Roseland anorthosite, is structurally above and intruded by the Mobley Mountain granite. The entire fold structure about Mobley Mountain is thus interpreted as an antiform.

Although the Mobley Mountain granite lacks any planar element predating Paleozoic deformation, folding of the granite is suggested by several observations, including the development of a mylonitic foliation (presumably axial planar to folding) and a northwest-southeast attenuation of the pluton's outcrop geometry. The granite is most strongly foliated along its northwest margins. Near the contacts in this area, sheared Turkey Mountain pluton is intensely mylonitized, with ultramylonite zones up to about 0.5 meters wide. The Buffalo River shows a deflection in its course towards this shear foliation trend.

GEOCHEMISTRY OF THE MOBLEY MOUNTAIN GRANITE

The Mobley Mountain granite is relatively homogeneous in its major-element chemistry (Table 1). Comparison of the Mobley Mountain granite major-element composition with average granites of LeMaitre (1976) and Rogers and Greenberg (1981), reveals that the average Mobley granite resembles most closely alkali granites, especially for Al_2O_3, MgO, CaO, Na_2O, and K_2O (Table 2).

Table 1. Major element variation of the Mobley Mountain granite.

	HIGH	LOW	MEAN	STANDARD DEVIATION
SiO_2	71.63	64.03	67.76	2.22
TiO_2	0.67	0.16	0.47	0.10
Al_2O_3	14.26	12.99	13.66	0.40
Fe_2O_3*	7.28	3.10	5.54	1.01
MnO	0.09	0.03	0.07	0.02
MgO	0.39	0.13	0.33	0.08
CaO	1.60	0.59	1.10	0.19
Na_2O	4.51	2.90	3.32	0.30
K_2O	5.97	4.80	5.08	0.34
P_2O_5	0.35	0.04	0.20	0.05
L.O.I.	0.78	0.26	0.64	0.14

Key to Table: data given as weight percent; *all iron as Fe_2O_3, L.O.I. = loss on ignition; 14 samples analyzed (Brock, 1981).

The distinction between calc-alkalic and alkalic granites is important because the former is generally subduction related, accounting for the bulk of granitic rocks exposed in continental areas, whereas the latter is associated with crustal attenuation and found in rifting environments. Petrogenetic relationships between the constituents of calc-alkalic suites can be portrayed on Harker diagrams where plots of the major element oxides against silica trace the systematic changes in the composition of a magma undergoing fractionation (Figure 3).

Rogers and Greenberg (1981) have suggested several criteria to aid in discrimination of alkali granites. Chemically, most alkali granite suites are either metaluminous or peraluminous; total K_2O plus Na_2O is high (greater than 6 percent), whereas CaO and MgO contents are low, and SiO_2 is greater than 65 percent. Based on these criteria, the Mobley Mountain granite, with SiO_2 of 67.76 percent, total Na_2O and K_2O of 8.4 percent, and low MgO (0.33 percent) and CaO (1.10 percent), is an alkali granite.

Another test of calc-alkalic vs. alkali granite suites can be shown by a plot of SiO_2 content vs. the ratio of potassium, the most lithophilic major element, to magnesium, the most mafic element (Ghuma and Rogers, 1980). Most Mobley Mountain granite samples plot within the alkali granite fields; those which do not are slightly more enriched in K_2O (Figure 4).

Alkali granites intruded into Precambrian shield areas are commonly post-tectonic and appear to precede the development of a stable crust. This contrasts with calc-alkalic granites which are generally syntectonic (Buddington, 1959). Many alkalic granites with low initial strontium isotopic ratios are localized within orogenic belts geographically related to continental margins and are commonly emplaced during the onset of continental rifting. The relatively low initial strontium ratios (less than 0.710) suggests a tectonic setting with limited crustal involvement in the generation of these these magmas (Loiselle and Wones, 1979; Rogers and Greenberg, 1981). The Mobley Mountain granite, which is post-Grenville, thus post major tectonism, has an initial strontium ratio of 0.7045, (discussed in the next section); thus it is a low initial $^{87}Sr/^{86}Sr$ alkali granite. A mantle-derived melt which did not interact with rocks of the lower crust seems best to describe the Mobley Mountain. Anatexis of upper mantle to yield alkali basalt has been well established (Ringwood, 1975). Loiselle and Wones (1979) proposed that alkali granites with low initial strontium ratios emplaced in continental blocks undergoing fracturing may have evolved with minimal crustal involvement by the fractionation of an alkali basalt magma.

The uppermost mantle, to about 70 kilometers depth, is predominantly composed of spinel lherzolite (Ringwood, 1975). Computer modeling of a typical Fen lherzolite by XLFRAC mass balance partial-melting simulation (Stormer and Nicholls, 1977), followed by fractionation of mafic phases, produced a magma virtually identical to the Mobley Granite magma (Brock, 1981). This procedure, of course, only demonstrates the possibility of a genesis for the granite by direct crystallization from this primitive liquid. The fractionation of mafic phases from a lherzolite-derived basaltic magma, at depths below the stability range of plagioclase (greater than about 40 kilometers), is the favored origin for the Mobley Mountain granite (Brock, 1981).

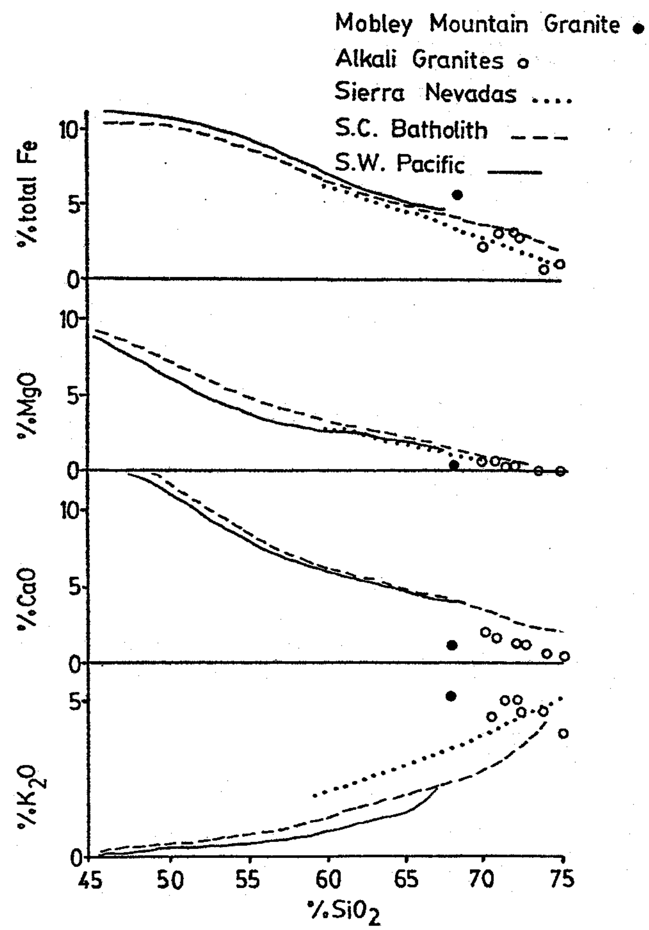

Figure 3. Harker diagram, Mobley Mountain granite compared to Calc-Alkaline and Alkaline granites. S.C. Batholith = Southern California; S.W. Pacific = New Britain (after Rodgers and Greenberg, 1981; Brock, 1981).

Table 2. Mobley Mountain granite compared to other granites (data in weight percent).

	Mobley	1	2	3	4	5
SiO_2	67.76	71.30	72.5	71.6	72.4	74.8
TiO_2	0.47	0.31	0.3	0.4	0.4	0.1
Al_2O_3	13.66	14.32	14.1	13.8	13.7	12.7
Fe_2O_3*	5.54	2.85	2.2	3.0	3.1	1.0
MgO	0.33	0.71	0.4	0.5	0.4	0.1
CaO	1.10	1.84	1.3	1.7	1.3	0.6
Na_2O	3.32	3.68	3.6	3.5	3.4	4.1
K_2O	5.08	4.07	4.8	5.2	5.1	4.5

Key to table: 1 = average granite (LeMaitre, 1976); 2 = average alkali granite (Rogers and Greenberg, 1981); 3 = six plutons of Mpageni and Sicumusa granite types (Rogers and Greenberg, 1981); 4 = Town Mountain granite, Llano uplift, central Texas (Rogers and Greenberg, 1981); 5 = Younger granites of eastern Egypt (Rogers and Greenberg, 1981); * = all Fe as Fe_2O_3

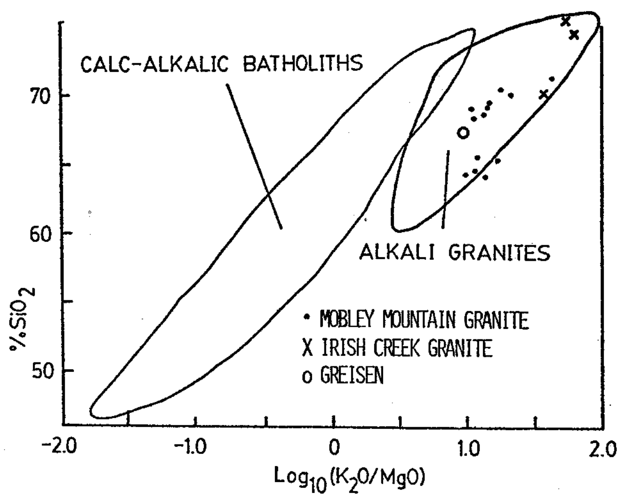

Figure 4. SiO_2 weight percent vs. log (K_2O/MgO), Mobley and Irish Creek granites compared to Calc-Alkalic and Alkalic granites (Rodgers and Greenberg, 1981; Brock, 1981).

Further evidence for a primitive source is seen in the mean K/Rb ratio of the Mobley Mountain granite (about 430). The normal range for K/Rb ratios for granites is 150,to 300 (Taylor, 1965). Low K/Rb values are a common consequence of extreme fractionation, as K is selectively removed relative to Rb by feldspar fractionation. The abnormally high K/Rb of the Mobley Mountain granite also suggest that following the fractionation of mafic phases, further fractionation, especially of feldspar, was minimal or nonexistent.

A gravity model also provides a strong argument for a deep source for the Mobley Mountain granite (Eppihimer, 1978). The model was based on the residual gravity anomaly measured above the granite and the density difference between biotite mylonite gneiss of the Turkey Mountain pluton and the granite. The model shows the Mobley Mountain granite as a 7.2 km-deep diapiric body. Assuming mesozonal emplacement of the Mobley Mountain granite, this model places the base of the granite near the lower crust-upper mantle interface (Ringwood, 1975).

Recent models of the evolution of the Appalachian Caledonian orogenic belt based on plate tectonics include an interval of continental rifting and the initial opening of the Iapetus Ocean during the late Precambrian (Windley, 1978). Within the central Appalachians, this lateral expansion was marked by the emplacement of nonorogenic, bimodal plutonic-volcanic rocks, whose silicic members consists of alkalic rocks of peralkaline affinity (Rankin, 1975). A Rb-Sr study was therefore conducted to see if the Mobley Mountain granite was a member of this late Precambrian group of alkalic rocks.

Rb-Sr AGE DETERMINATION

Twelve samples were collected from the Mobley Mountain granite for Rb-Sr study (Figure 2). Modal (Table 3) and major-element (Table 4) analyses were carried out as well as Rb and Sr isotopic analyses (Table 5). The Rb-Sr technique used in this study is based on the radioactive decay of ^{87}Rb to ^{87}Sr by beta-emission. The half-life for the decay is estimated to be about 4.89×10^{10} years; the decay constant is $1.42 \times 10^{-11} \times$ years $^{-1}$ (Steiger and Jager 1977). The samples of the granite were each about 10 kg. They were crushed and split to 10-g portions which were powdered; about 0.3 g of each sample was isotopically analyzed. Each analysis was done using ^{84}Sr and ^{87}Rb spikes, and ultrapure HF, $HClO_4$, and HCl.

The strontium analyses were made using a Nier-type, 6-inch radius, single filament mass spectrometer with a programmable automatic data-acquisition system at the Department of Terrestrial Magnetism of the Carnegie Institution in Washington, D.C. The Rb isotopic analyses were made using a 12-inch radius mass spec-

Table 3. Modal analyses of Mobley Mountain granite.*

Mineral	C-2	D-2	F-2	L-1	M-1	N-1	O-1	P-1	V-1	W-1
QUARTZ	37.5	35.3	33.9	37.5	37.4	30.9	33.8	33.0	37.2	36.3
MICROCLINE PERTHITE	24.7	20.1	32.1	23.4	23.9	33.2	25.6	28.2	24.6	19.7
PLAGIOCLASE	16.0	18.1	15.4	20.9	17.4	13.9	17.6	19.2	18.1	15.0
BIOTITE	12.1	15.5	10.7	8.5	9.9	13.0	11.7	12.0	10.0	13.4
CHLORITE	0.1				0.9	0.1	1.2			
EPIDOTE	4.5	7.5	3.6	3.6	4.3	6.8	4.9	3.5	5.4	12.4
MUSCOVITE	2.0	0.4	2.2	1.7	1.7	0.6	0.4	tr	1.5	1.5
SERICITE	1.7	1.8	1.6	2.1	2.7	1.0	4.4	2.2	2.2	1.5
GARNET	0.2	0.1		0.5	0.1			0.2	0.1	
FLUORITE	0.2	0.1	0.1					0.2		0.1
FLORENCITE	0.2	0.2		0.3	0.5	0.5	tr	tr	0.6	tr
APATITE	0.1		0.2	0.4	0.4	0.1				0.1
ZIRCON	0.1	0.1	tr	tr	tr	tr	tr	0.7	0.1	tr
SPHENE	0.5	0.9	0.1	0.4	0.6	0.2	0.4	0.7	0.1	0.4

* = total points counted in each=800; tr = trace

Table 4. Major element analysis of Mobley Mountain granite.*

	C-2	D-2	F-2	L-1	M-1	N-1	O-1	P-1	V-1	W-1
SiO_2	65.51	68.29	69.10	68.58	64.03	65.21	70.04	69.45	70.42	65.19
TiO_2	0.52	0.55	0.48	0.43	0.42	0.40	0.41	0.52	0.36	0.67
AL_2O_3	13.13	13.50	13.99	13.54	14.37	14.12	13.01	13.79	14.03	13.21
Fe_2O_3	5.42	7.50	5.09	5.05	4.78	4.71	4.82	6.06	4.78	7.28
MgO	0.38	0.39	0.33	0.30	0.36	0.28	0.26	0.32	0.26	0.52
CaO	0.89	1.10	0.91	1.60	1.07	1.06	1.17	0.96	1.01	1.64
Na_2O	2.90	3.13	3.37	4.51	3.30	3.31	2.85	2.85	3.29	3.42
K_2O	4.91	4.77	5.02	4.28	5.19	5.13	5.86	5.00	4.91	5.47
MnO	0.09	0.09	0.09	0.03	0.08	0.07	0.03	0.09	0.07	0.04
P_2O_5	0.20	0.27	0.20	0.15	0.21	0.16	0.14	0.19	0.15	0.35
L.O.I.	0.78	0.66	0.65	0.26	0.62	1.26	0.57	0.66	0.37	0.68
Sum	94.73	100.25	99.42	101.43	94.42	95.72	99.17	99.89	99.66	98.46

*all Fe as Fe_2O_3; L.O.I. = loss on ignition, measured after 1 hour at 1000°C; analysis by XFR (Philips XRG-3000 x-ray generator, PW 1410 x-ray spectometer, Cr target tube, machine settings 40KV and 30MA); standards = USGS GSP-1, USGS G-2, and SDC-1; for normative minerals and trace element analysis, see Brock, 1981.

trometer at Florida State University. All the strontium isotopic compositions were calculated from analyses of sample plus spike mixture. The $^{85}Rb/^{87}Rb$ ratio was taken to be 2.593 (Steiger and Jager, 1977).

The Rb-Sr age and initital $^{87}Sr/^{86}Sr$ ratio on the isochron diagram were calculated using the regression treatment described by York (1966). The one-standard-deviation experimental error on $^{87}Rb/^{86}Sr$ was calculated to be 2 percent, and the one-standard-deviation experimental error in $^{87}Sr/^{86}Sr$ was calculated to be 0.05 percent. These error estimates were derived from an examination of duplicate analyses done over the past seven years, and these estimates include sample splitting errors. The estimate of one-standard-deviation experimental errors which do not include an error increment related to sample splitting are 1 percent for $^{87}Rb/^{86}Sr$ and 0.02 percent for $^{87}Sr/^{86}Sr$. The errors assigned to the reported age and initial $^{87}Sr/^{86}Sr$ ratio on the isochron diagram are given at the 68 percent confidence level (1 sigma). Visual examination of the fit of the Rb-Sr isotopic data to an isochron line is provided by Figure 5.

As shown in Figure 5, the Mobley Mountain granite appears to have crystallized at 652 ± 22 m.y. ago. Other Rb-Sr studies have identified several granitic plutons of approximately this age that intruded Grenville rocks in New York (Long, 1969; Mose and Hayes, 1975; Mose and others, 1982), Virginia (Eckelmann and Mose, 1981; Mose 1981; Smith and others, 1981), and North Carolina (Odom and Fullagar, 1971; 1982). Studies presently in progress indicate that the Blue Ridge and similar terranes in the northern Appalachians contain many more granitic plutons which were emplaced in latest Precambrian time, about 700 to 550 m.y. ago.

RELATIONS OF THE MOBLEY MOUNTAIN GRANITE TO THE IRISH CREEK GRANITES

A granite with an associated tin deposit crops out at Irish Creek about 23 kilometers to the northwest of Mobley Mountain, intruding the westernmost Blue Ridge basement terrain. The $^{40}Ar/^{39}Ar$ age spectra on 2 syngenetic muscovite samples from Irish Creek greisen have yielded well defined plateau ages of 637 and 634 ± 11 m.y. (Hudson and Dallmeyer, 1982). This fluorite-bearing 2-mica granite appears to be spatially, mineralogically, and genetically related to greisenization of the Irish Creek tin district. These ages are interpreted to approximate the time of emplacement of the granite at Irish Creek (Hudson and Dallmeyer, 1982). The petrological, geochemical, and geochronological similarities of the granites of Mobley Mountain and Irish Creek

Table 5. Rb and Sr isotopic analyses of Mobley Mountain granite.

Sample Number	$^{87}Rb/^{87}Sr$ Atomic Ratio	$^{87}Sr/^{86}Sr$ Atomic Ratio	^{86}Sr ppm	^{87}Rb ppm
C-2	1.797	0.7210	16.488	29.978
D-2	1.548	0.7183	18.085	28.319
E-2	1.083	0.7142	18.484	20.250
F-2	1.742	0.7200	16.858	29.709
L-1	0.552	0.7094	40.777	22.755
M-1	1.658	0.7200	16.228	27.215
N-1	1.776	0.7219	16.127	28.969
O-1	1.843	0.7224	15.648	29.166
P-1	1.716	0.7211	17.650	30.638
S-1	3.216	0.7332	8.451	27.494
V-1	1.945	0.7230	15.378	30.260
W-1	0.942	0.7135	22.304	21.256

Analyses E-2 and S-1 are not shown in Tables 3 and 4 but are located in Figure 2. E-2 is from same outcrop as D-2.

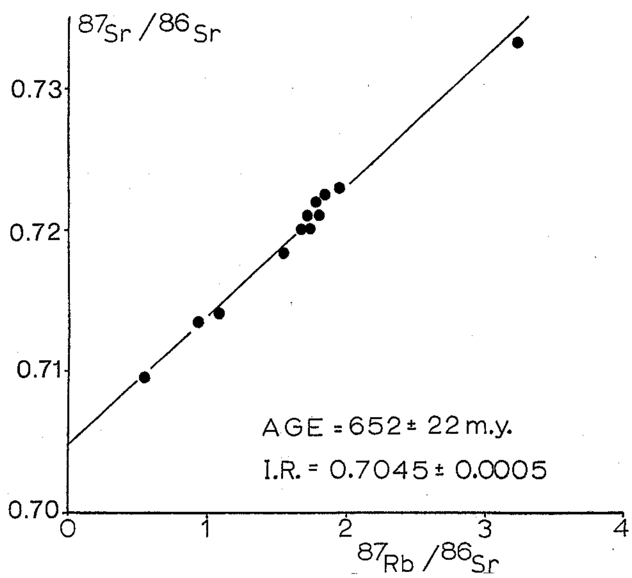

Figure 5. Rb/Sr isochron diagram, Mobley Mountain granite.

have led Herz and others (1981) to suggest that the latter may represent a tin-rich cupola area of the Mobley Mountain granite transported to the northwest by Paleozoic thrusting.

The widespread occurrence of tin mineralization associated with anorogenic, fluorite-bearing, alkaline granitic plutonism has been discussed by Sillitoe (1974), who proposed an origin for these stanniferous complexes by fractionation of mantle plumes. Such a genesis for the Irish Creek tin greisenization is consistent with the upper mantle origin of the Mobley Mountain granite proposed in a preceding section.

CONCLUSIONS

The Mobley Mountain granite intruded Grenville rocks of the Blue Ridge basement complex 652 ± 22 m.y. ago. The migmatitic contacts and petrology suggest a deep mesozonal emplacement and the absence of a chilled margin suggests a low temperature gradient between the granite and country rock during emplacement. Lower structural levels are presently exposed along the centerline of the pluton due to antiformal folding concurrent with Paleozonic deformation.

The low initial strontium isotopic ratio (0.7045), 7.2-kilometer deep diapiric gravity model, Rb-Sr age, and major-element chemistry of the Mobley Mountain granite suggest that is is an alkali granite emplaced in a rifting continental block. It evolved with minimal crustal involvement by the differentation of upper mantle-derived alkali basalt at depths below that of plagioclase stability.

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