GEOLOGY OF THE HUMPBACK MOUNTAIN AREA OF THE

BLUE RIDGE IN NELSON AND AUGUSTA COUNTIES, VIRGINIA

by

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INTRODUCTION

Purpose and Scope of Investigation

The main purposes of this investigation were: (1) to establish, by means of outcrop mapping, the stratigraphic succession of the Late Precambrian Catoctin Group along the western flank of the Blue Ridge in central Virginia; and (2) to determine the relationships among the underlying Middle Precambrian Virginia Blue Ridge Complex, the Catoctin Group and the overlying Early Cambrian Chilhowee Group. Secondary objectives were to establish: (1) the structural-metamorphic sequence of events in this region; (2) the various sources and/or source directions of the Catoctin Group rocks; (3) the mineralogy of the rocks; and (4) the metamorphic grade.

The mapped area was chosen because of the large aereal extent of the Catoctin Group as mapped by Bloomer and Werner (1955, <u>their</u> Plate 1). Reconnaissance indicated that a thicker than normal section of the Catoctin rocks was present and that the upper part of the Catoctin Group was probably in depositional, rather than fault, contact with the overlying Chilhowee Group.

Previous Work in Catoctin Belt in Northern Blue Ridge

Geologic work on the Blue Ridge of Pennsylvania, Maryland, and Virginia generally was begun prior to 1850.

Notable among this very early work was that done by Rogers (1884) in Virginia. Campbell (1882) also contributed to the geologic knowledge of the Blue Ridge portion of Virginia. Most other articles published prior to 1900, as well as these two, are reviewed in seven articles: Keith (1892, 1893, 1894); Geiger and Keith (1891); Williams (1894); Walcott (1896); and Bascom (1896). Keith's reports deal with regional discussions of the northern end of the Blue Ridge; Williams reported on the old volcanic rocks of eastern North America; Walcott discussed the Cambrian rocks of Pennsylvania; and Bascom presented a detailed petrographic analysis of the Catoctin volcanics.

Table I is a list of the maps and accompanying texts, published since the late 1800's, which are concerned with the Catoctin belt. Notable among these articles are those done by King (1950a), Reed (1955) and Nickelsen (1956), all of whom mapped along the western flank of the Blue Ridge. The distribution of the Catoctin Group is shown on Plate 1 and is a compilation of all the maps listed in Table I.

As can be seen from Plate 1, the Catoctin Group occurs along two distinct belts in Virginia. These two belts merge northward into a single belt which extends across Maryland into southern Pennsylvania. The more

TABLE I. LIST OF THE PUBLISHED GEOLOGIC MAPS IN THE NORTHERN PART OF THE BLUE RIDGE AND ADJACENT PIEDMONT WITH A SCALE OF 1:125,000 OR LESS: LOCATIONS ON PLATE 1.

Index Number	Scale	Author	Date of Publication
A-1	1:125,000	Keith	1894
A-2	1:125,000	Stose, Miser, Katz and Hewett	1919
A-3	1:125,000	Stose	1932
A-4	1:125,000	Furcron	1935
A-5	1:125,000	Bloomer and Werner	1955
B-1	1:76,032	Diggs	1955
B-2	1:63,360	Gravatt	1920
B-3	1:63,360	Knechtel	1943
B-4	1:62,500	Furcron	1939
B-5	1:62,500	Stose and Jonas	1939
B-6	1:62,500	Espenshade	1954
B-7	1:62,500	Reed	1955
B-8	1:62,500	Whitaker	1955a
B-9	1:62,500	Brown	1958
B-10	1:62,500	Bick	1960
B-11	1:62,500	Brent	1960
B-12	1:62,500	Nelson	1962
B-13	1:62,500	Allen	1963
B-14	1:62,500	Redden	1963
B-15	1:62,500	Werner	1966
B-16	1:62,500	Allen	1967
C-1	1:42,450	Scotsford	1951
C-2	1:31,680	Diggs	1955
C-3	1:31,680	Parker	1968
C-4	1:31,250	King	1950a
D-1	1:25,000	Nickelsen	1956
D-2	1:24,000	Toewe	1966
D-3	1:24,000 (outcrop map)	Freedman	1967
D-4	1:24,000	Fauth	1968
D-5	1:24,000	Root	1968
D-6	1:24,000	Spencer	1968
D-7	1:19,500	Bascom	1896
D-8	1:1,440	Foose	1945

easterly belt is intimately involved in stratigraphic-structure problems of the western Piedmont. Controversies have centered on the existence of both the Martic overthrust and the James River synclinorium and have been discussed by Jonas (1927, 1929, 1932), Mackin (1935), Cloos and Heitanen (1941), Furcron (1935, 1939, 1969), Espenshade (1954), Brown (1951a, b, 1953, 1958, 1970), and Redden (1963). Structural analysis by Cloos (1937, 1941, 1946, 1947, 1950, 1951, 1958), Root (1970) and others listed on Table 1 pertains principally to the Blue Ridge of Maryland and Pennsylvania. Both the eastern and northern belts are separated from the study area by large unmapped areas. Hence extrapolation westward and southward does not seem justified at this time.

Early regional studies dealing with the Blue Ridge in Virginia by Furcron (1934, 1940) on the rocks of the Shenandoah National Park and Butts (1933, 1940) on the Paleozoic rocks of the Valley and Ridge were followed by summations, mainly of the Precambrian-Cambrian boundary, by Furcron and Woodward (1936), Jonas and Stose (1939), Stose and Stose (1949) and King (1949). Subsequently, Gooch (1954, 1958) did a regional study on the Mechum River metasediments. This unit, which was named by him, appears to consist of two rock types, one of which is equivalent to the lower part of the Catoctin Group

(Furcron, 1969, p. 79-80). Reed (1964) discussed the chemistry of the Catoctin rocks. Other recent regional studies have been made by Colton (1970), Reed (1970), Espenshade (1970), Brown (1970), Tilton, Doe and Hopson (1970), King (1970b). In addition, the tectonic setting of the entire Blue Ridge with respect to the Appalachian orogenic belt has been reviewed by King (1950b, 1964, 1970a) and Eardley (1951, 1962).

Studies which overlap the present study area are: Stose, Miser, Katz and Hewett (1919) and Knechtel (1943) on mineral deposits; Bloomer and Bloomer (1947) and Bloomer (1950) on the Catoctin Formation; Bloomer and Werner (1955) on the Blue Ridge of central Virginia; and Schwab (1970) on the Antietam Formation. Griffin (1971) published a structural analysis of the Blue Ridge of central Virginia, and Werner (1966) mapped the adjacent quadrangle to the west of the study area. Dietrich (1962, p. 16-21) presents a road log of the Blue Ridge near the mapped area.

Recent unpublished these and dissertations, dealing with portions of the Catoctin belt, are by: R.M. Cordova (1955, University of Virginia); O.P. Bricker and V.M. Seiders (1958, Franklin and Marshall College); R.A. Landy (1961, The Pennsylvania State University); R.L. Perkings (1967, West Virginia University); H.J. Goett (1969, The

George Washington University); and J. Wickam (1969, The Johns Hopkins University). Current research is being conducted by J.L. Fauth (State University of New York at Cortland); D. Elliott (The Johns Hopkins University); Chico Onash (Franklin and Marshall College); and the Virginia Division of Mineral Resources.

Location, Access and Physiography

The mapped area consists of the Sherando 7½-minute quadrangle and the western portion of the adjacent Greenfield 7½-minute quadrangle located in north central Virginia (Plates 1 and 2). Plate 2 is indexed with a grid of letters and number coordinates for convenience in locating features cited. The boundaries, embracing 190 square km, are 38°00' to 37°52'30" north latitude and longitude 78°52' to 78°49'30" and approximately parallels State Highway 151.

The principal access roads within the area are the Blue Ridge Parkway from Mile 3 (J-1) to Mile 15 (A-13), and State HIghways 151 (O-1 to H-15), 664 (D-1 to F-15), and 610 (A-1 to I-1). The small communities of Sherando (D-3), Sherando Camp (C-6), Beach Grove (F-15), Wintergreen (I-14), Nellysford (L-13), Lodebar (M-12, Greenfield (N-9) and Avon (O-1) are located within the area. Present access to the mountainous part of the area is mostly by foot because of damage to jeep trails caused by erosion from the August, 1969, Hurricane Camille.

The area includes three physiographic subdivisions: (1) the Blue Ridge Mountains (part of the Blue Ridge physiographic province); (2) Rockfish Valley (an inter-montane valley); and (3) the Sherando Valley (part of the Valley and Ridge physiographic province) to the northeast of the mountains. Within the mapped area, the principal peaks generally range in elevation from 750 m to 1200 m. The elevation of the valley to the east is mostly less than 300 m. On the northwest side of the Blue Ridge, the valley floor is mostly 450 m or higher.

Principal physiographic features within the area are Elk Mountain (J-1), Dobie Mountain (J-4), Johnson Mountain (M-4), Humpback Mountain (I-7), Crawford Knob (F-11), Piney Mountain (C-13), Devils Knob (D-10), and Torry Mountain (B-7).

Field and Laboratory Work

Seventy-seven days were spent in the field between November, 1969, and February, 1971. The mapping was done on both 1:12,000 photo-enlargements of the 1:24,000 scale topographic maps and air photos with an approximate scale of 1 cm = 100 m. The base map was reduced to a scale of 1:24,000 after all data were plotted. On both the map

(Plate 2) and the cross sections (Plate 3), the zip-a-tone patterns are used to emphasize foliation and bedding trends. Thus, what may appear as angular unconformities, folds or faults are due to the linearity of the zip-a-tone patterns and should be disregarded unless such features are indicated by other symbols or discussed in the text. Most of the orientation data were collected with a Brunton pocket transit; however, because of relict magnetism in some of the greenstones, many of the strikes and trends were determined from the air photos.

Most contacts between different rock types were determined by using exposed ledges, float or slide rock, and topographic expression. In a few instances, however, vegetation was also of some aid. The following apparent plant-rock relationships were noted: (1) the general absence of mountain laurel and pines on the greenstone; (2) the general absence of rhododendron and black birch on granite-gneisses and sandstone-shale sequences; and (3) the general absence of mountain laurel and the presence of pines and rhododendron on the metasedimentary rocks interlayered with greenstones.

Two hundred sixty-four thin sections were examined and modes w ere estimated for all thin sections. The estimated modes were obtained by randomy picking one to three points on the slide with the low power objective and at each point, using a higher power objective, estimating how much of the visible area each mineral occupied. The estimated modes of the Catoctin Group greenstones and pyroclastics were then used to determine changes in mineralogy as a function of burial depth. Eight point counts were made on randomly selected thin sections of greenstones and pyroclastics to be used as standards for the estimated modes. The point counts also were used to calculate approximate chemical compositions of the greenstone and pyroclastic samples for comparison with actual chemical analyses done on the greenstones (Reed, 1964). Two hundred points per thin section were counted by making 10 traverses equally spaced across the slide and counting 20 points per traverse. The 200 points per slide were considered to be representative point counts by comparing point counts of 100, 200, 300, 400, 500 and 600 (Table II).

From Table II, it can be seen that the percentage of any mineral present in theppoint count of 200 grains does not differ by more than a few percent from the percentage of the same mineral present in the point counts of 300, 400, 500 and 600 grains.

Relative grain sizes were determined primarily in the field using a marked hammer. Grain sizes in the collected samples were measured using a small ruled scale. However, no systematic or statistical determination of

TABLE II. COMPARISON OF MODES OF SAMPLE NUMBER 11 AS DETERMINED BY POINT COUNTS OF 100, 200, 300, 400, 500 AND 600 POINTS.

Number of Points	Fine-grained Epidote	Epidote	Penninite	Actinolite	Magnetite and Hematite	Plagioclase
100	20%	5%	5%	15%	14%	41%
200	22%	8.5%	5%	18.5%	10%	36%
300	20%	6%	6%	14%	15%	39%
400	20%	5.5%	6%	14.5%	15%	39%
500	21%	7%	5.5%	16%	13%	37%
600	21%	6.5%	5.5%	16%	13%	38%

grain size in either hand specimens or thin sections was attempted.

Forty-eight x-ray diffraction patterns were made using a Norelco instrument with CuKα nickel-filtered radiation. The scans were made at a rate of 2 degrees 2θ per minute. The patterns principally are whole rock; however, a few are samples consisting of one or two minerals only.

The appendix is a tabulation of the locations, formations and rock types of the samples collected. The samples from which thin sections, x-ray patterns and point counts were made also are indicated in the appendix. Plate 5 is an index map of the sample localities. All samples, thin sections, thin section descriptions, x-ray patterns and a copy of the appendix plus a copy of Plate 5 are on deposit in the Department of Geological Sciences, Virginia Polytechnic Institute and State University.

STRATIGRAPHY AND PETROLOGY

Regional Stratigraphy and Petrology

The Blue Ridge geologic belt northeast of Roanoke, Virginia, consists of a breached anticlinorium in the center of which are exposed the granites and gneisses of the Virginia Blue Ridge Complex named by Brown (1958, p. 7). The study area lies about 10 to 15 km west of the median axis of the anticlinorium (Plate 1). According to Bloomer and Werner (1955, *their* Plate 1), the Lovingston Gneiss occupies the center of the anticlinorium and is flanked by the Marshall Gneiss and the granites and gneisses of the Pedlar Formation. These formations are partially metamorphosed sediments (Bloomer and Werner, 1955, p. 581), and age dates range from 770 my (Rodgers, 1952, p. 421) to 1150 my (Tilton and others, 1960, p. 4175). Erosion exposed all these rocks prior to Late Precambrian time.

On the western flank of the anticlinorium, the Catoctin Group rests unconformably on the Virginia Blue Ridge Complex. The Catoctin Group as herein defined generally consists of the basal sedimentary Swift Run Formation, the interlayered volcanics and sediments of the Catoctin Formation and the volcanics and arkosic sediments of the Loudoun Formation (Plate 2, Plate 4).

Overlying the Catoctin Group is the northeastern equivalent of the Chilhowee Group, the uppermost part of which is of definite Early Cambrian age. Equivalents of the Chilhowee Group from bottom to top consist of the Weverton Formation, the Harpers Shale, and the Antietam Formation. In southwest Virginia and adjacent Tennessee, the true Chilhowee Group consists from bottom to top of the Unicoi Formation, the Hampton Formation and the Erwin Formation. In the northern Blue Ridge, the Chilhowee Group constitutes the western foothills of the Blue Ridge and is overlain by lower Paleozoic carbonates, which are found to the west in the Sherando Valley. Metamorphism of the Chilhowee and Catoctin groups increases southeastward across the northeast structural grain of the Blue Ridge (Espenshade, 1970, p. 209) and stratigraphically downward. The latest metamorphism and the principal deformation of the Blue Ridge anticlinorium occurred during the Paleozoic and probably is post-Ordovician (Cloos, 1957, p. 838).

Virginia Blue Ridge Complex

Pedlar Formation

Name and age: The Pedlar Formation, which comprises the bulk of rock within the Blue Ridge anticlinal core north of Roanoke, was named by Bloomer and Werner (1955,

p. 582) for the upper Part of the Pedlar River, Amherst County, Virginia, where rock is typically exposed. Bloomer and Werner (1955, p. 582) consider the Pedlar Formation as consisting generally of an assemblage of undifferentiable "granitic-dioritic" rocks. Table III summarizes the available age determinations made on the Pedlar Formation. These ages are discussed later.

Distribution: Within the study area, the Pedlar Formation has been divided on the basis of texture into two mappable rock types: (1) Pedlar Granite; and (2) Pedlar Gneiss. The Peldar Granite is exposed in the cores of a series of east-west-trending anticlines (L,M-4,5; K,L-6,7; L-8,9; G-13,14). These antiforms are mantled by about 150 to 450 km of Pedlar Gneiss, which in turn grades outward into the Marshall Gneiss.

Pedlar Granite: The Pedlar Granite consists of a variety of rocks which mineralogically could be classified as granite, granodiorite, diorite and unakite. The Pedlar Granite is distinguished from the Pedlar Gneiss by lack of apparent mineral alignment (foliation) in the former. In hand specimens, these granites typically are coarse-grained (3 to 10 mm) and composed mainly of quartz and feldspar. In the unakite types, epidote, in addition to quartz and pink feldspar, is a primary

TABLE III. AGE DETERMINATIONS OF THE PEDLAR FORMATION.

Formation	Rock Type	Location	Material Dated	Method	Age	Reference
Pedlar Granite	pegmatite in hypersthene granodiorite	Amherst Co., Virginia	allanite	U-Th-Pb	800 my	Marble (1935, p. 351)
				recal- culated	860 my	Rodgers (1952, p. 421)
Pedlar Gneiss	hypersthene granodiorite gneiss	Shenandoah National Park, Virginia in area mapped by Reed (1955)	zircon	U-Th-Pb	1170 to 1110 my	Tilton and others (1960, p. 4175)
			biotite	Rb-Sr	880 my	Tilton and others (1960, p. 4176)
			biotite	K-A	800 my	Tilton and others (1960, p. 4176)
Pedlar Granite	unakite (collected by Drs. W.D. Lowry and R.V. Dietrich)	State Highway 56 in area mapped by Werner (1966)	whole rock	K-A	941 my	Letter to Dr. W.D Lowry (dated 10/5/61) from Clifford Lilly of Gulf Research Laboratory

constituent.

Table IV summarizes the estimated modes of the typical granite samples examined in thin section. In thin section, the texture is typically hypautomorphic-granular or, to a lesser extent, xenomorphic-granular. Quartz and feldspar frequently are graphically intergrown. In addition, hypersthene is at least partially altered to one of three lower metamorphic assemblages: (1) garnet; (2) biotite; or (3) epidote and chlorite. Brown hornblende, however, exhibits only slight alteration near the edges. The feldspars are generally partially altered to sericite or sericite and epidote, and in some samples, biotite or chlorite also appear to replace feldspar. Typically the feldspars are perthitic, but in the unakite samples (77, 114, 224) the pink feldspars are maximum microcline as determined by x-ray. These pink feldspars have a perthite-like texture due to uneven extinction.

X-ray patterns of sample 209 indicate that ilmenite is the only opaque oxide present and and that quartz and low albite also are present as major constituents of the segregated light fraction. The x-ray pattern of the garnet present in sample 209 matches that calculated for a garnet with composition $(Mg_{1.6}Fe_{1.2}Ca_{0.2})\ Al_2(SiO_4)_3$ given by Borg and Smith (1969).

Most of the granites examined in thin section show

TABLE IV. ESTIMATED MODAL COMPOSITION OF TYPICAL PEDLAR GRANITE.

Sample Number	Quartz	Perthite ± K-feldspar	Plagioclase	Hypersthene	Hornblende	Biotite	Chlorite	Sericite	Epidote	Opaque Oxides	Garnet	Apatite	Zircon
4	25	50	5	10	--	--	2	3	5	X	--	--	--
39	30	20	--	5	15	7	X	10	X	5	2	X	--
52	10	30	--	15	7	2	X	30	2	4	--	X	--
56	25	40	X	--	20	--	5	5	5	X	X	--	--
57	5	40	5	5	10	--	10	10	5	7	X	--	X
58	20	40	--	15	--	3	--	10	10	X	X	--	--
75A	15	X	15	10	3	X	X	40	13	2	--	2	--
77A	20	50	20	--	--	X	--	X	5	4	--	--	X
77B	35	10	--	--	--	--	--	--	55	X	--	--	--
114A	40	25	--	--	--	--	--	X	35	--	--	--	--
114C	40	20	--	--	--	--	--	15	25	--	--	--	--
121	30	55	--	--	--	--	5	2	--	X	5	--	X
125A	15	55	--	20	--	5	--	5	X	X	--	--	--
206	20	55	X	X	2	X	X	15	5	X	--	--	--
208	25	65	--	--	--	X	2	5	3	X	--	--	X
209	10	25	5	25	X	--	2	10	--	5	15	--	X
211B	26	60	--	X	--	--	5	5	2	3	X	--	--
224	15	15	--	15	--	--	--	--	50	5	--	X	--

X - 1 percent or less

some degree of granulation and shearing, but the granitic texture of most is well preserved. Table V summarizes the estimated modal composition of the sheared granites examined in thin section. These samples typically are located near faults or intrusions into the granite and exhibit various degrees of granulation and shearing. In the more extreme examples due to shearing abrasion, the quartz grains are rounded with undulatory extinction or are microcrystalline. Feldspars are also rounded or bent and exhibit undulatory extinction. The microscopic shear zones are filled with sericite, fine-grained epidote $\pm$ biotite, $\pm$ chlorite, $\pm$ calcite. Hypersthene, hornblende and feldspar typically are highly altered or are not present in large quantities.

Pedlar Gneiss: The Pedlar gneisses in hand specimen are mineralogically, but not texturally, similar to the Pedlar granites. The gneisses typically are medium-grained (2 to 5 mm) quartz-feldspar rich rocks with one or more foliations. Table VI summarizes the estimated modal composition of the Pedlar Gneiss samples.

In thin section, the Pedlar Gneiss is similar to the Pedlar Granite in that both have quartz and perthite as the major constituents (Tables IV, V, VI). Also, both have some quartz and feldspar graphically intergrown, and any hypersthene, if present, shows signs of alteration

TABLE V. ESTIMATED MODAL COMPOSITION OF SHEARED PEDLAR GRANITE.

Sample Number	Quartz	Perthite + K-feldspar	Plagioclase	Hypersthene	Hornblende	Biotite	Chlorite	Sericite	Epidote	Opaque Oxides	Garnet	Apatite	Zircon	Calcite
17	35	45	--	--	--	--	--	15	--	5	--	--	--	--
18	30	50	--	--	--	--	--	20	--	--	--	--	--	--
113A	25	55	5	--	--	3	6	5	--	X	X	X	--	--
113B	35	10	5	--	--	5	7	25	10	X	--	--	--	--
123A	25	40	--	10	--	X	10	5	2	5	--	--	--	--
123B	25	40	--	--	25	X	--	X	X	--	--	--	--	10
123C	15	30	--	10	--	15	5	5	5	5	--	--	--	10
124A	15	50	--	--	--	5	--	X	20	5	X	X	--	5
124B	40	20	--	--	15	5	2	7	X	5	--	X	--	5
128	20	45	5	--	X	X	10	15	5	X	--	--	--	--
138	10	25	--	2	10	10	5	5	25	5	--	X	X	--
142	20	35	--	15	2	X	X	20	3	2	--	X	--	--

X - 1 percent or less

TABLE VI. ESTIMATED MODAL COMPOSITION OF PEDLAR GNEISS

Sample Number	Quartz	Perthite + K-feldspar	Plagioclase	Hypersthene	Biotite	Chlorite	Sericite	Epidote	Opaque Oxides	Garnet	Zircon	Calcite
115A	30	50	--	15	--	X	--	X	2	--	--	--
115B	25	35	--	15	--	5	--	5	3	2	X	--
115C	40	45	--	5	7	--	X	X	2	--	X	--
125B	25	50	--	5	10	5	--	--	2	X	X	--
134	30	40	10	--	--	5	3	5	--	X	X	--
135	10	25	40	--	--	10	10	--	X	--	X	--
141A	35	35	--	--	X	--	25	X	X	--	--	4
141B	50	35	--	--	--	--	10	X	5	--	X	--
141C	37	43	X	--	--	X	12	5	3	--	--	--
141D	20	10	X	--	X	2	60	2	2	--	X	--
141E	20	25	--	--	--	X	40	5	X	--	X	--
164A	25	30	25	--	--	10	X	X	X	--	X	--
164B	15	60	X	--	7	8	5	X	X	--	X	--
164C	25	35	20	--	5	5	3	2	2	X	X	--
164D	25	45	X	--	8	5	5	X	3	X	--	--
199	27	45	--	--	--	--	23	X	X	X	X	--
201A	20	60	X	--	--	X	20	--	X	--	--	--
201B	30	45	--	--	--	--	20	--	X	2	X	--
201C	20	50	--	--	X	--	30	--	X	--	--	--
201D	20	60	5	--	X	--	10	--	X	--	X	--
202	30	35	X	--	--	--	35	--	X	--	X	--
203	30	10	45	--	3	5	X	X	2	2	X	--
204	30	15	20	--	3	2	--	20	4	--	--	--
212A	20	50	20	--	--	X	X	--	X	X	X	--
212B	25	50	10	--	--	2	10	X	X	--	--	--
215A	15	55	15	--	--	4	X	--	6	X	X	--
215B	15	--	--	--	--	X	80	X	2	X	--	--
217	15	10	X	--	2	5	65	X	3	X	X	--
226 (dike)	35	--	5	--	5	15	7	10	15	2	--	--

X - 1 percent or less

to lower grade metamorphic minerals.

The Pedlar Gneiss differs mineralogically from the Pedlar Granite in that the former has little or no hornblende, apatite or calcite; and hypersthene is present only in a few samples collected from a thin gneissic layer (E-12) in the granite (Tables IV, V, VI). Also, microcrystalline quartz, sericite, plagioclase and zircon generally are more abundant in the gneisses, and epidote typically is less abundant. The gneisses are similar to the sheared granites in that both exhibit granulation and shearing. Typical retrograde alterations in the gneiss appear to be: (1) plagioclase and/or perthite to sericite and chlorite $\pm$ garnet; and (2) hypersthene completely altered to chlorite and epidote. In sample 164A, some of the feldspars appear to have overgrowths inasmuch as inclusions are present in the centers but not in the borders of grains. An x-ray pattern was made of sample 204 because of its fine grained granulated and sheared texture. Quartz and low albite are the dominant minerals present.

Sample 226 was collected from a typical dike cutting the Pedlar Gneiss. The dikes range from a few centimeters to about 1 m in width and can be traced along strike for up to 100 m where exposures are good.

<u>Pedlar metagabbro</u>: Three bodies believed to be metamorphosed gabbro were mapped in the Johnson Mountain anticline area (L-5; N-4; N,O-5). The bodies at the latter two localities both lie near the contact between the Pedlar and Marshall formations, whereas the other body appears to lie adjacent to a layer of Marshall Gneiss which is interlayered with Pedlar Gneiss and Pedlar Granite.

These rocks weather to large spheroidal pock-marked boulders 0.3 to 2 m in diameter. In hand specimen, the rocks are black and coarse-grained (up to 5 mm). Three thin sections of samples collected from the L-5 body were examined.

In thin section, the texture is xenomorphic-granular with most of the grains appearing more or less rounded. The mineralogy of the samples is variable even though they were collected near one another. In one section, collected near the border of the body, the estimated mode of the rock consists of about 25 percent orthopyroxene, 5 percent each clinopyroxene, quartz and magnetite, and 50 percent plagioclase. The plagioclase ranges in composition from andesine (An_{37}) to oligoclase (An_{26}). In addition, minor amounts of less than 2 percent of zircon, apatite (?), and chlorite were noted. All the edges of the mafic minerals exhibit some degree of alteration.

The other two samples from the interior of the body have estimated modes of about 50 percent orthopyroxene, 25 percent amphibole, 5 percent clinopyroxene, 2 percent magnetite and 10 percent altered brownish plagioclase. The relict plagioclase appears to have been altered to epidote and muscovite (?) which comprise about 5 percent of the rock. In one sample, about 5 percent dark green spinel is present.

In general, the mafic rock assemblages appear to be compatible with the granulite facies observed in the Pedlar and Marshall formations. Likewise, the mafic bodies show some signs of being partially altered to lower grade minerals. Hence, the age of the metagabbros is considered to be either coincident with the granulaite facies metamorphic event, or prior to this metamorphic event. The metagabbros may represent altered basic sills intruded into the Pedlar-Marshall rocks.

The differences in composition between the margin and the interior of the metagabbro body suggest that an inward silica migration may have occurred. This silica migration is indicated by the presence of quartz and relatively unaltered andesine-oligoclase in the margin sample and the presence of green spinel and relict (calcic?) plagioclase in the interior samples.

Marshall Gneiss

Name, age and distribution: The Marshall Gneiss is named for its type area near Marshall, Fauquier County, Virginia, and was first mapped by Jonas (1928, map). Within the map area, the Marshall is interpreted as more or less conformably overlying the Pedlar Formation and to be unconformably (?) overlain by the Lovingston Gneiss. The Marshall Gneiss is poorly exposed in a belt along the eastern and southern boundaries (L-1 to O-1 to L-13 to E-15). In the Johnson Mountain anticline, a layer of Marshall Gneiss is interlayered with both Pedlar Granite and Pedlar Gneiss (K-5, M-4, M-5).

Lithology: Within the Marshall there are what appear to be one or more quartzite layers, medium-grained (1 to 3 mm) granulitic gneisses, and the typical gray-green medium-grained quartz-feldspar-mica rich gneiss. The quartzite layers are shown on the map (H-15 to J-14, L-11, N-9), but the other two rock types were not mapped as separate units. The Marshall normally exhibits one well-developed feldspar and/or mica foliation and also commonly has one or more secondary crinkle foliations and/or shear foliations superimposed on the primary mineral foliation. The Marshall Formation is distingiished from the Pedlar

Gneiss by the formers better developed principal foliation, the common presence of two or more foliations, and the presence of quartzite layers.

In hand specimens, both quartz and feldspar are easily discernible, and accessory mica is generally present. Table VII summarizes the mineralogy of the thin sections examined. In general, the Marshall Formation is similar to the Pedlar Gneiss in both mineralogy and metamorphic history. Both gneisses appear to have been metamorphosed initially to the granulite facies and subsequently to have been retrograded to as low as the greenschist facies.

Lovingston Gneiss

Name, age and distribution: Jonas (1928, map) originally named the Lovingston Gneiss for exposures near Lovingston, Nelson County, Virginia. The Lovingston is herein interpreted as being younger than the Marshall Gneiss and older than the Catoctin Group of Late Precambrian age. According to Bloomer (1950, p. 758) and Bloomer and Werner (1955, p. 582), the Lovingston consists of two subunits; a gneiss and a granite. The Lovingston Gneiss and Lovingston Granite crop out in an area 40 by 13 km in the core of the Blue Ridge anticlinorium. Bloomer and Werner (1955, p. 582) also state that the gneissic portion of the formation

TABLE VII. ESTIMATED MODAL COMPOSITION OF MARSHALL GNEISS.

Sample Number	Quartz	Perthite + K-feldspar	Plagioclase	Hypersthene	Hornblende	Biotite	Chlorite	Sericite	Epidote	Opaque Oxides	Garnet	Apatite	Zircon
137A	60	23	--	--	--	13	--	X	3	X	--	--	--
143	35	--	15	--	2	--	--	40	--	5	--	X	X
144	25	60	--	--	--	X	10	--	--	X	X	--	X
145	30	50	X	10	--	2	--	2	X	--	--	--	X
146	23	50	15	--	--	X	--	10	--	2	X	X	X
148A	45	20	--	--	--	X	--	30	--	2	--	--	X
149	12	70	--	--	--	X	--	15	X	X	--	--	X
184	18	--	50	20	--	7	--	X	--	5	--	--	X
193A	10	70	15	--	--	--	2	X	--	X	--	--	X
193B	10	70	15	--	--	X	2	X	--	--	--	--	X
194	15	50	10	X	--	X	2	X	15	3	--	--	X
196	25	70	--	--	--	X	--	X	--	X	2	--	X
197	15	10	15	--	--	5	--	20	--	2	30	--	X
207	10	15	--	20	--	7	--	40	--	3	--	--	--
210	25	18	--	--	--	--	5	35	10	5	2	--	X
211A	5	10	--	--	--	--	2	70	2	5	5	--	X
218	30	10	--	--	--	--	--	55	2	2	--	--	X
219A	7	50	10	--	--	3	20	--	5	5	--	--	--
219B	75	--	--	--	--	X	3	X	15	--	--	--	X

X - 1 percent or less

surrounds the granitic portion. Within the map area, outcrops of the Lovingston Gneiss are limited to a northeast-trending belt in the southeastern corner (I-15 to N-11), and Lovingston Granite apparently is not present. Although none of the granitic part of the formation is found within the area, the Lovingston Granite does crop out within 1.5 km east of the area along State Highway 6 (N-11). The thickness of the Lovingston Gneiss, within the map area, appears to exceed 500 m.

Lithology: Within the area, exposures of the gneiss are gray and show a well developed mica foliation. The estimated mode of the rock consists of about 70 percent feldspar, 20 percent mica and 10 percent blue quartz. The feldspar augen range up to 20 m in length, whereas the mica occurs in a braided fashion around the feldspar and quartz. The quartz crystals are about 5 mm in diameter. An x-ray pattern of sample 478 indicates that the mica is muscovite and the feldspar is low albite. No thin sections of the Lovingston were examined.

Origin of Virginia Blue Ridge Complex

The origin of the Virginia Blue Ridge complex is discussed below in the following sequence: (1) probable metasedimentary features of Pedlar and Marshall formations; (2) significance of the 1070 to 1110 my zircon age (Table III) with respect to the granulite facies metamorphic event; (3) relative age of the retrograde metamorphic event; (4) origin of Lovingston Gneiss.

(1) The Pedlar Formation and the overlying (?) Marshall Gneiss appear to be partially metasedimentary in origin as indicated by the following features:

a) The interlayered nature of the Pedlar granites with Pedlar gneisses and, in one case, with Marshall type gneiss. This layering occurs both on the map scale (K,L,M-3,4,5,6; H-13) and on the outcrop scale (along Little Stony Creek; I-9).

b) The presence of probable crossbedding in the Pedlar gneiss along Little Stony Creek (I-9).

c) The presence of a thin band of gneiss (samples 115A, B, C) within the Pedlar Granite (E-12), which is about 60 cm thick and at least 4 to 6 m long. This band consists of eight alternating layers of light and dark gneiss ranging from 2 to 18 cm in thickness. All eight layers persist along the entire exposure. The light layers range from 7 to 18 cm in thickness and consist of light tan, fine-grained (1 to 2 mm) quartz-feldspar gneiss. The dark layers range from 2 to 5 cm in thickness and consist of greenish gray to black fine-grained (1 mm or less) gneiss with large biotite crystals

(1 to 3 mm).

(2) The significance of the 1170 to 1110 my age determinations (Table III) is dependent upon whether the Pedlar is considered as metasedimentary or metaigneous. The granulite facies event can be dated as approximately 1100 my if the Pedlar Granite is interpreted as a meta-igneous rock intruded during this event. Alternatively, the granulite facies event must be considered 1100 my or younger if the Pedlar zircons are interpreted as detrital in origin. Inasmuch as the age determination was made on zircons of the Pedlar Gneiss and the evidence presented above suggests the Pedlar and Marshall gneisses are probably metasedimentary, the zircon age determination is interpreted as the maximum age of the granulite facies event. This age, however, probably is more than 1100 my, because the sample probably experienced some lead loss during later metamorphism. Hence, the age probably is discordent. Future concordia-type age dating of what are believed to be metasedimentary layers and adjacent massive granites may necessitate alteration of this interpretation.

The remaining dates of 800 to 941 my (Table II) all seem to indicate the effect of a post-granulite-facies, lower grade metamorphic event, during which unakite bodies were formed, probably by hydrothermal activity (Reed, 1955, p. 876). This low grade metamorphic event may only represent the last stage of the metamorphism, which culminated in the granulite facies metamorphic event. The lower grade event appears to have reached different grades in different samples. The three principal retrograde stages appear to be: (1) brown hornblende, garnet and sericite; (2) brown hornblende, brown biotite and sericite; and (3) chlorite, epidote, quartz and sericite $\pm$ calcite; or epidote, microcline and quartz.

The age of the retrograde metamorphic event is younger than the granulite facies event but older than Catoctin volcanism (700+ my) on the basis of age determinations (Table III) discussed below. The minimum age appears to be approximately 950 my, and the actual age probably is much older. This retrograde metamorphic event appears to be older than the low grade Paleozoic metamorphism for the following reasons:

a) The presence of pink feldspar clasts in the Loudoun metasediments. These pink feldspars, which are maximum microcline, are similar to the pink feldspars present in the unakite rocks of the Pedlar Granite. Inasmuch as pink feldspar is limited in occurance to the low grade epidote-microcline-quartz rocks in the Pedlar, it appears that the pink feldspars are themselves a product of that low grade metamorphism. Hence, the formation of unakite must have preceded Loudoun sedimentation because the only known source of the pink feldspar is the unakite rocks of the Pedlar Granite.

b) The presence of unakite cobbles in the lower flows of the Catoctin Formation suggests that the formation of the unakite preceded Catoctin volcanism.

c) The presence of unakite and other retrograded rocks of the Pedlar Formation beneath unmetamorphosed Chilhowee Group rocks (Werner, 1966, p. 21, 24, 25) indicates that the retrograded event preceded not only Chilhowee sedimentation but also Paleozoic metamorphism.

d) Inasmuch as little or no Paleozoic metamorphism appears to have affected the unakite described in (c), the 941 my age determination of this unakite body (Table III) may be accepted as the minimum age of the retrograde event. This event may be somewhat older if significant argon diffusion has occurred. Likewise, the 860 my age determination

(Table III) indicates a minimal Precambrian age of epidote formation in the Pedlar. This date is probably discordent due to later lead loss. The biotite dated by Tilton and others (1960) is described by them (p. 4178) as "...fresh and undistorted" and is interpreted to be metamorphic in origin. Hence, this age date of 880 my (Table III) must also represent a minimal Precambrian age of the retrograde event.

(4) The Lovingston is interpreted as being older than the Catoctin volcanism but younger than the Middle Precambrian folding of the Pedlar and Marshall formations inasmuch as only northeast trends are apparent in the Lovingston, whereas the northeast trends are superimposed on older east-west trends in the Marshall and Pedlar formations. The Lovingston may be: (1) a metasedimentary sequence unconformably deposited on the folded Pedlar-Marshall rocks; (2) an intrusive body in which the Lovingston Gneiss is a reaction cloak of the Lovingston Granite; or (3) a metasedimentary sequence intruded by an igneous body.

Catoctin Group

The Catoctin Group as herein defined refers to the entire sequence which contains volcanic rocks that grade either upward or downward into what is called the Catoctin Greenstone or Catoctin Formation. This assemblage was first mapped as the Catoctin Group (Formation) by Reed (1955, p. 878-879). The distribution of the Catoctin Group is shown on Plate 1. Thus the Catoctin Group consists of the Swift Run Formation, the Catoctin Greenstone or Formation and the Loudoun Formation, but does not include either the Lynchburg Group (Furcron, 1969, p. 71, 83, 84) or the Unicoi of southwestern Virginia and adjacent Tennessee.

The Unicoi in southwestern Virginia should not be considered part of the Catoctin Group because it at least in part unconformably overlies the Mount Rogers volcanics (Stose and Stose (1957, p. 57, 98), which may be a Catoctin correlative.

The Swift Run Formation is included in the Catoctin Group because detailed mapping in the study area (Plate 1 and Plate 2) as well as in the Luray (Reed, 1955) and Harpers Ferry (Nickelsen, 1956) areas has shown that metasediments and metatuffs called Swift Run Formation, which rest on the basement, alternate upward with greenstones and more metasediments and metatuffs which are indistinguishable from the lower ones. Thus no definite lithologic break in the sequence exists. It could be argued that the basal metaconglomerate typically represents sedimentation prior to the Catoctin flows and that it is unique. On Crawford Knob (H,I-9) and Elk Mountain (L-3), however, the Swift Run metasediment has been traced from where it overlies basement to where it is underlain by greenstone which rests on Pedlar Gneiss. This relation indicates that Catoctin volcanism, in some places at least, preceded Swift Run sedimentation. Finally, the basal part of the Swift Run is considered to consist, at least in part, of basement weathered essentially *in situ* (Bloomer and Werner, 1955, p. 587; Reed, 1955, p. 880-881). Inasmuch as weathering is a surface chemical process, it follows that the age of such material is the time it ceased to be exposed to surface weathering or, in this case, the age of the flow or sediment that covered it. Hence, considering the paleotopographic relief (Plate 2 and Plate 3) of up to 300 m on which this colluvium developed, it follows that the age of the colluvium must be time transgressive and equivalent to the rock column which overlapped this topographic surface. Thus the formation of the basal metaconglomerate or metabreccia is equivalent of up to 300 m of greenstone, metatuffs and metasediments.

From the foregoing discussion, it is apparent that the Swift Run Formation is logically a part of the Catoctin volcanic sequence. The name Swift Run Formation, therefore, should be retained for use in those regions where, on the basis of detailed mapping, a thick section of mappable tuffaceous and/or arkosic metasediments exists between the basement complex and the first thick series of greenstone flows.

The Loudoun Formation is included in the Catoctin Group because the Loudoun greenstones are indistinguishable from those of the underlying Catoctin Greenstone or Formation. Also, siltstone phyllites and what are interpreted as pyroclastic phyllites are present in both formations. These two types of Loudoun phyllites are essentially identical to their Catoctin Formation counterparts in both composition and texture. The Loudoun Formation was distinguished from the underlying Catoctin Formation on the basis of a thick section of metasedimentary rocks with a few interlayered greenstones and the presence of medium-to-coarse-grained meta-arkose beds.

Swift Run Formation

Name and age: Stose and Stose (1944, p. 410) named the Swift Run Formation for the sequence of tuffaceous and arkosic sediments exposed along U.S. Highway 33 and along the Skyline Drive just east and north, respectively, of Swift Run Gap, Greene and Rockingham Counties, Virginia. Jonas and Stose (1939, p. 582-583) had previously described the rocks at this locality. The Swift Run unconformably overlies the Virginia Blue Ridge Complex and is overlapped by the Catoctin Formation.

Distribution: Within the study area, the Swift Run is exposed in a narrow north-trending belt (L-3 to L-10) in the Spruce Creek syncline (E-14 to H-12), in a U-shaped belt in the Stony Creek area (F-11 to G-9 to J-8 to J-12), and other small exposures (D-11; G-11; J,K-9). The Swift Run definitely pinches out in the Potato Patch Mountain area (F-11), and at two other localities (L-3, L-11), it also appears to pinch out.

Lithology: Within the map area, the Swift Run Formation consists primarily of metaconglomerate and metasandstone. Other lithologies are greenstones, rocks interpreted as pyroclastic phyllites, siltstone-mudstone phyllites, and metasaprolite derived from weathered rock of the Virginia Blue Ridge Complex. The thickness of the Swift Run is variable, and in several places (D-11; H-12,13; I-9; I-13; and J-5 to L-3), it probably exceeds 300 m. Figure 1 is a diagram showing the source indicators (crossbedding, thickness and relative grain size) for the Swift Run.

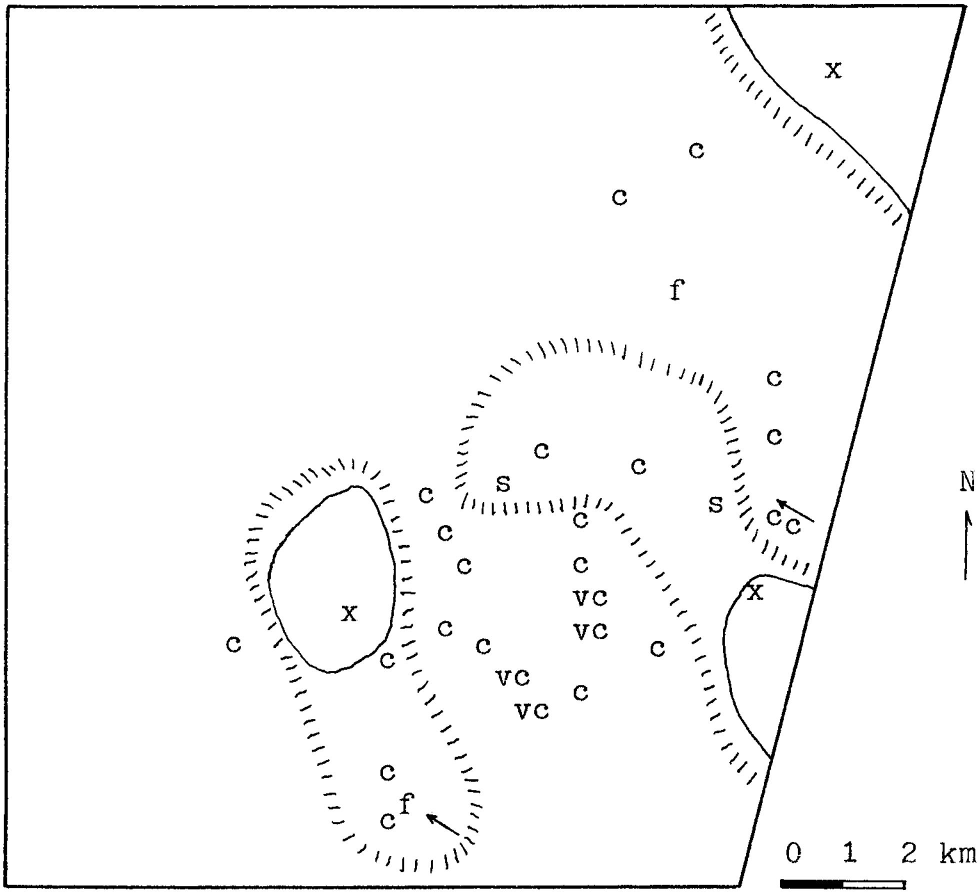

Figure 1. Distribution of Swift Run

- hill
x - Swift Run absent
- Swift Run pinch out
f - fine-grained detritus
c - coarse-grained detritus
vc - very coarse detritus
s - saprolite
- direction of transport-crossbedding

The Swift Run was not subdivided into smaller units except at a few localities (H,I-9; L-3). Basal greenstone flows were mapped at two places (H,I-9; L-3). A conglomerate layer was mapped at one locality (H-12), and the middle portion of the Swift Run consists of a mappable 3 to 6 m fine-grained siltstone phyllite with numerous ptygmatic quartz layers in the Spruce Creek syncline (G-13; F-14). At the latter locality, the upper and lower parts of the Swift Run are conglomeratic metasandstone.

The coarsest metaconglomerate within this formation was found near the middle of the formation (I-11), where disk-shaped purple-blue quartz cobbles up to about 40 cm in diameter occur. Most of the ellipsoidal cobbles at this location and another (H-12) are flattened parallel to the foliation plane. Many of the darker (blue purple) cobbles typically have white pressure-shadow overgrowths developed at either one or both ends. One overgrowth, typically is much better developed than the other.

Swift Run metasaprolite, which grades downward into Pedlar Gneiss, was found at just two localities: (1) on the trail just west of the Stony Creek gorge (G-8), and (2) in the stream bed of Pauls Creek (K-9). Both of these occurences are on the slopes of what are interpreted to be paleohills adjacent to the main basin of Swift Run deposition because at both localities the overlying lower Catoctin Greenstone or Formation pinches out against the Swift Run metasaprolite (Figures 1, 2, and 3). Thus, the metasaprolite appears to be a product of weathering during Catoctin Formation time rather than the initial surface of Swift Run deposition. The thickest and stratigraphically lowest portions of the Swift Run are detrital deposits resting directly on the Virginia Blue Ridge Complex, not on metasaprolite derived from the complex. Bloomer and Werner (1955, p. 587), Reed (1955, p. 877, 880-881), and Werner (1966, p. 11) believe that almost all of the basal Swift Run Formation is basement saprolite rather than detrital deposits.

Thick pyroclastic phyllites were noted at three localities (G,F-9,10; K-6). Similar pyroclastic phyllites also are present as very thin layers and lenses within the metasandstone-conglomerate. The reasons for interpreting these phyllites as pyroclastic in origin are discussed later.

<u>Greenstone</u>: Two greenstone flows which rest directly on Pedlar Formation and are overlain by thick conglomeratic metasandstone sequences are present at the base of the Swift Run Formation (H,I-9; L-3). In hand

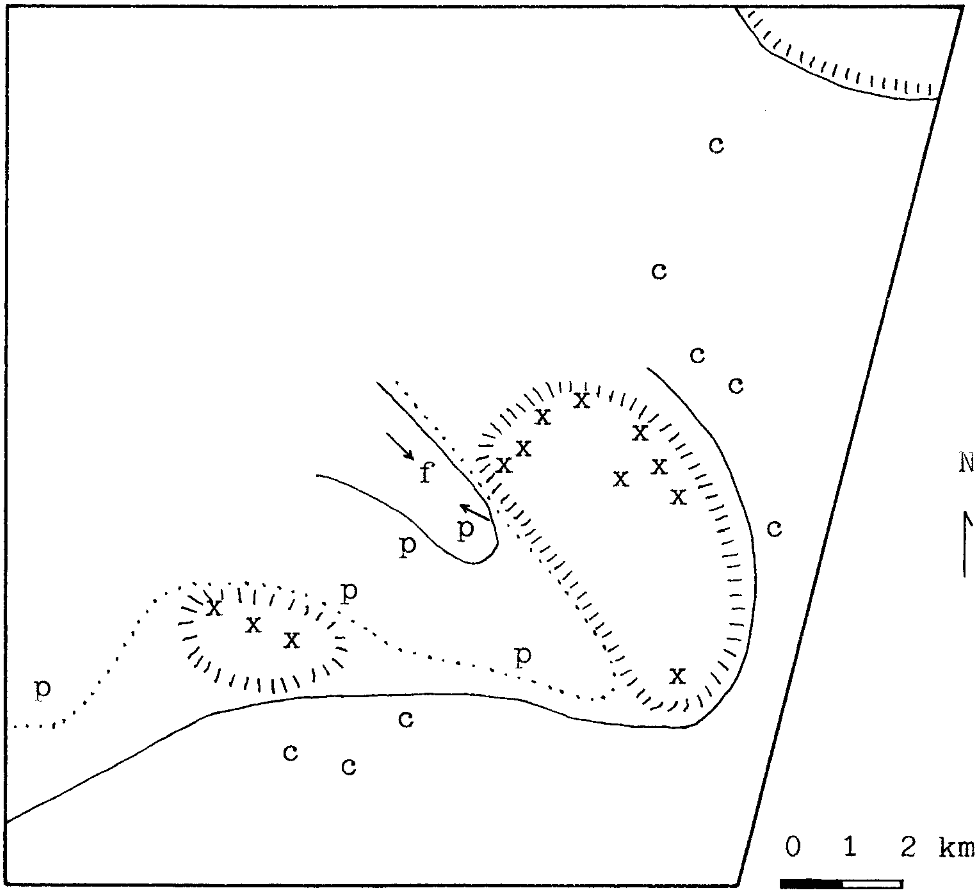

Figure 2. Distribution of Unit 1

- hill
x - Unit 1 absent
- sediment pinch out
..... - pyroclastic pinch out
f - fine-grained detritus
c - coarse-grained detritus
p - pyroclastics
- direction of transport-crossbedding

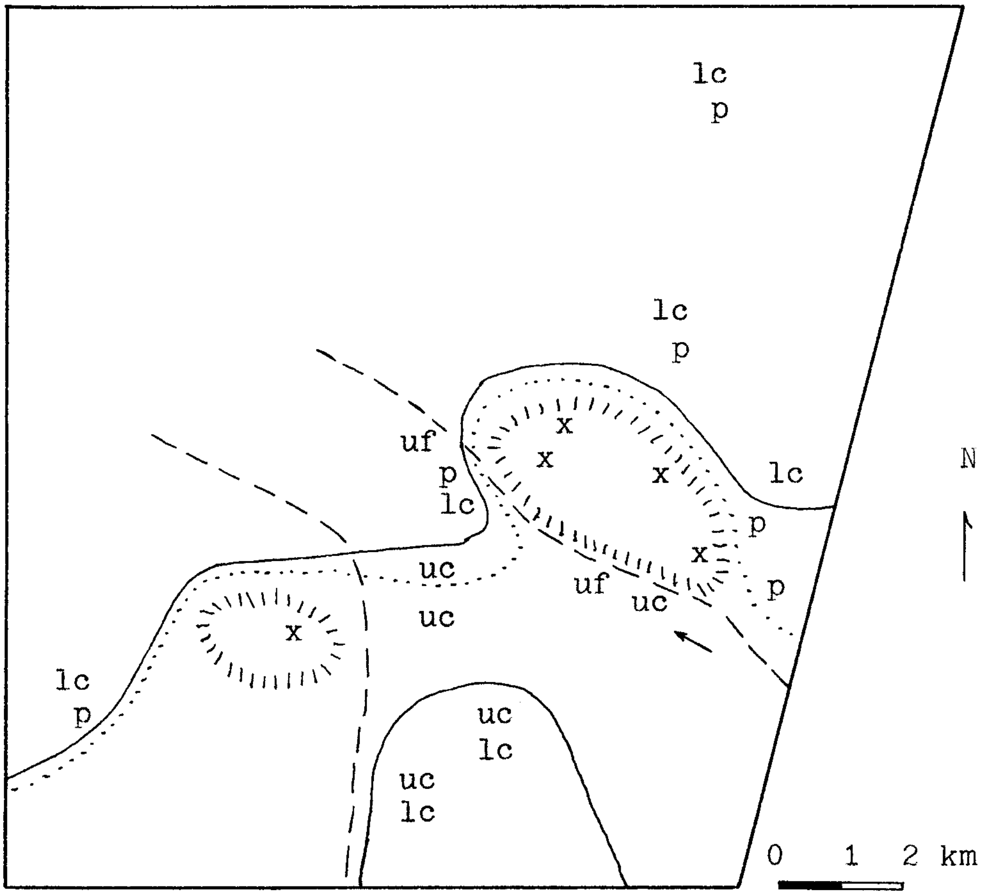

Figure 3. Distribution of Unit 2

ııııı - hill
x - Unit 2 absent
——— - lower (l) sediment pinch out
— — — - upper (u) sediment pinch out
..... - pyroclastic pinch out
c - coarse-grained detritus
f - fine-grained detritus
p - pyroclastics
↖ - direction of transport

specimens, these greenstones are essentially identical to the type 2 greenstone of the overlying Catoctin Formation described later in the text. The former are dark gray-green, dense rocks exhibiting well developed foliation or cleavage. At both localities the greenstones probably represent more than one flow, but exposures are insufficient to be certain.

Pyroclastic phyllite: King (1950a, p. 16) first interpreted the purple to red slates as being pyroclastic in origin, whereas Reed (1955, p. 893) believed that the uppermost purple to reddish brown volcanic slate (Loudoun) was a metasaprolite derived from volcanic rocks. Reed's description (1955, p. 892-893) of this purple volcanic slate is sufficiently different from any of the pyroclastics of the study area to cast doubt both on its presence in this area and on a metasaprolite origin for all of these phyllites. Likewise, Reed (1955, p. 879) does state that he believes sericite phyllites present in the Luray area "...may be derived from volcanic ash layers". Within the study area, all of the purplish gray, reddish gray and greenish gray thinly-cleaved sericite phyllites of the Catoctin Group are mineralogically and texturally similar and are interpreted as pyroclastic in origin. This interpretation is based on the following comparisons between these phyllites and the numerous metasedimentary phyllites of the same grain size with which they are interlayered.

(1) The pyroclastics have a mineralogy consisting of muscovite ($\pm$ penninite), magnetite ($\pm$ hematite), and quartz $\pm$ epidote (Tables VIII and IX). Both metasediments and greenstones have more variable mineralogy.

(2) The pyroclastics lack bedding. All of the metasediments exhibit bedding both microscopically and in the outcrop. The only textural feature, besides cleavage, present in the pyroclastics are large discolored irregular blotches observable in hand specimens sliced perpendicular to the cleavage.

(3) The pyroclastics have no heavy minerals (in thin section) such as zircon, garnet or tourmaline, and magnetite (and/or hematite) grains are evenly dispersed throughout the rock. The metasediments all have some heavy minerals which typically are concentrated in thin layers along with magnetite (and/or hematite). Thus the bedding is defined by layers rich in heavy minerals alternating with layers poor in heavy minerals.

(4) The pyroclastics lack clastic grains of quartz and/or feldspar. All the metasediments have some recognizable clastic grains which are concentrated principally

TABLE VIII. ESTIMATED MODAL COMPOSITION OF CATOCTIN GROUP PYROCLASTIC PHYLLITES.

Sample Number	Formation	Muscovite	Penninite	Quartz	Epidote	Magnetite	Hematite	Fine-grained Opaques
91	Swift Run	60	--	--	--	10	10	20
287A	Catoctin Unit 1	30	10	35	--	15	--	10
287B	Catoctin Unit 1	50	15	20	--	10	--	5
287C	Catoctin Unit 1	35	15	35	--	10	--	5
287D	Catoctin Unit 1	50	--	25	--	5	X	20
287E	Catoctin Unit 1	35	15	30	--	10	--	10
283E	Catoctin Unit 2	55	--	20	5	5	--	15
74A	Catoctin Unit 3	20	30	15	15	2	2	15
74B	Catoctin Unit 3	60	10	5	2	10	--	10
74C	Catoctin Unit 3	10	60	5	15	X	--	10
130C	Catoctin Unit 3	60	--	3	5	10	10	10
80A	Catoctin Unit 5	50	--	--	--	25	10	15
80B	Catoctin Unit 5	50	--	--	--	25	10	15
80C	Catoctin Unit 5	50	--	--	--	25	10	15
237	Loudoun	35	--	40	5	15	5	X
315A	Loudoun	65	--	20	X	--	--	15
315C	Loudoun	40	--	30	X	30	--	X

X - 1 percent or less

TABLE IX. X-RAY MINERALOGY OF CATOCTIN GROUP PYROCLASTIC PHYLLITES.

Sample Number	Formation	$2M_1$ Muscovite	Penninite	Quartz	Epidote	Magnetite	Hematite
91 blebs	Swift Run	X	-	-	X	X	X
91	Swift Run	X	-	-	-	X	X
287A	Catoctin Unit 1	X	X	X	-	X	X
287D	Catoctin Unit 1	X	X	X	-	X	X
222	Catoctin Unit 2	X	-	X	-	X	-
283E	Catoctin Unit 2	X	-	X	-	-	X
74A greenish part	Catoctin Unit 3	X	X	X	-	-	X
74A speckled part	Catoctin Unit 3	X	X	X	X	X	X
74B	Catoctin Unit 3	X	X	X	-	X	X
130C	Catoctin Unit 3	X	-	-	-	X	X
80A	Catoctin Unit 5	X	-	-	-	X	X
237	Loudoun	X	-	X	-	X	X
315A	Loudoun	X	X	X	-	-	-
315C	Loudoun	X	X	X	-	X	-

in the layers that are poor in heavy minerals.

(5) The pyroclastics have muscovite and/or quartz blebs which are uniformly distributed throughout the rocks and are elongated parallel to the cleavage. None of the metasediments have blebs. These blebs may represent vugs in the original ash deposits.

In hand specimens, the Swift Run pyroclastic phyllite (sample 91) is a fine-grained, medium gray, thinly foliated rock with a silvery sheen on the foliation surface. Flattened disk-shaped blebs 5 to 10 mm in diameter and 1 to 3 mm thick comprise about 20 percent of the rock. They are composed of sheets of soft white micaceous material and are flattened parallel to the plane of foliation. From thin section examination and an x-ray pattern, the fine-grained opaque blebs were determined to consist principally of muscovite with epidote, magnetite and hematite. In other Catoctin and Loudoun pyroclastics, the blebs are quartz and muscovite.

The remainder of the rock in thin section was estimated to consist of about 10 percent each of magnetite and hematite intergrown with 60 percent sericite. The sericite exhibits a very strong preferred orientation.

The mode of sample 91 determined by a point count of 200 grains was: 62 percent sericite, 4.5 percent quartz, 14.5 percent magnetite plus hematite and 19 percent fine-grained opaque material. The differences between the point count mode and the estimated mode (Table VIII) are mainly in the magnetite plus hematite and the quartz contents. The higher estimated hematite content was determined with a low power objective and reflected light, whereas the point count was made using a high power objective and the condenser. Under the latter conditions, the hematite appears to be primarily staining intergrown with the fine-grained opaques.

The small quartz content (4.5 percent) noted in the point count was similarly included in the estimated sericite content. Because of the very fine grain size of both the sericite and quartz, they are not distinguishable with the lower power objective.

Coarse to fine-grained phyllite: The most common rock type is the coarse-grained (3 to 5 mm), greenish-gray to greenish-brown, quartz-rich conglomeratic metasandstone with blue-purple quartz clasts up to 3 cm. These beds are either massive or torrentially crossbedded and range up to about 6 m in thickness. The conglomerates, in addition to the rounded cobbles of blue-purple quartz, have many well-rounded white quartz cobbles and a few scattered fragments of granite and gneiss and many angular phyllite fragments. The size of most quartz cobbles and rock fragments ranges from 25 to 150 mm.

The siltstone phyllites differ from the metasandstones in grain size and amount of obvious detrital quartz present. Bedding in the siltstone phyllites is 25 to 75 mm in thickness.

In thin section, the metasandstones consist of 25 to 50 percent detrital quartz grains which were rounded or subrounded but which have been metamorphically altered and hence appear more angular. Ten percent or less of the rock is composed of feldspars and/or feldspar-rich rock clasts. Perthite, microcline, plagioclase and orthoclase all have been identified optically on the basis of twinning in different samples. The feldspars and rock fragments, like the quartz grains, appear to have been rounded ellipses which subsequently have been altered and now appear more angular. By contrast, the metasiltstones appear to contain little if any feldspar and rock fragments and 15 percent or less quartz.

Both rock types typically have 10 to 20 percent opaques with magnetite predominating over fine-grained opaques and hematite. These rock types are characterized by 40 to 80 percent matrix which is composed of 60 to 100 percent *in situ* metamorphic sericite plus microcrystalline quartz $\pm$ up to 30 percent epidote $\pm$ 5 percent or less penninite ± 5 percent or less biotite (?) ± 1 percent or less garnet or zircon.

Where the metasaprolite is present, the Swift Run Formation contains cobbles of quartz, feldspar, unakite and Pedlar Gneiss. Cobbles of the latter increase in size and angularity until they are not distinguishable from the underlying Pedlar Gneiss.

Metamorphic alterations: Replacements observed in thin sections include the following:

(1) perthite replaced by epidote; and sericite

(2) plagioclase replaced by epidote; sericite; and penninite

(3) orthoclase replaced by sericite

(4) quartz replaced by muscovite flakes; and sericite plus microcrystalline quartz

(5) epidote replaced by penninite

(6) magnetite replaced by cloudy fine-grained epidote (?); and biotite

These alterations suggest that the metasediments were altering to a greenschist assemblage of epidote, sericite, quartz and penninite.

Catoctin Formation

Name and previous work: The Catoctin Formation as used herein refers to the interlayered greenstones, pyroclastics, and metasediments between the underlying Swift Run Formation and the overlying Loudoun Formation. This tentative assignment may cause some confusion through the use of Catoctin as both a group name and a formation name within that group. However, until a better formation name is found, this usage should be continued inasmuch as Catoctin does express the primarily volcanic nature of both stratigraphic units. Catoctin Greenstone may be an acceptable substitute for Catoctin Formation. However, greenstone is misleading in that the formation is an interlayered sequence of both volcanic and clastic units.

The Catoctin Formation, which is named for Catoctin Mountain, Loudoun County, Virginia, and Frederick County, Maryland, was named originally by Geiger and Keith (1891, *their* Plate 4) and described later by Keith (1893). Prior to these two works, Rogers (1884) and Williams (1892) had interpreted the Catoctin rocks as igneous (intrusive) and volcanic, respectively. More detailed information on the Catoctin in or near the present study area may be found in Bloomer and Bloomer (1947) and Bloomer (1950), who describe the Catoctin in central Virginia. Reed (1955), King (1950a), Brent (1960), Nelson (1962) and Allen (1963) all mapped portions of the Catoctin belt north of the area. Werner (1966) mapped the Catoctin belt in the Vesuvius quadrangle adjacent to the west, and Bloomer and Werner (1955) mapped both the Vesuvius quadrangle and the study area.

Regional distribution: The Catoctin is exposed most extensively in central and northern Virginia (Plate 1), where it consists of two belts about 32 km apart. The westerly belt more or less follows the prominent crest of the Blue Ridge, whereas the easterly belt is found in the foothills of the Blue Ridge. These two belts join in Frederick County, Maryland (Plate 1), and the single belt continues northward through Maryland as far as the South Mountain area of Pennsylvania. In addition, the Accomac volcanics of eastern York County, Pennsylvania, (Plate 1) probably represent the easternmost exposure of the Catoctin volcanic belt.

Regional lithologic variations: In general, the Catoctin Formation in Virginia consists of thick sequences of greenstones (metabasalt to meta-andesite lava flows) interbedded with a few metasedimentary and pyroclastic phyllites. In Pennsylvania and Maryland, however, the upper part of the Catoctin is rhyolitic in composition. Reed (1955) in the Luray area, and Nickelsen (1956) in the Harper's Ferry region mapped the distribution of metasedimentary and pyroclastic phyllites within the Catoctin Formation. Within the study area, seven phyllites (Plate 2 and Plate 3) have been differentiated, but their relationships to those mapped by previous workers is unknown. No correlations were attempted because some of the metasedimentary units pinch out within the study area, and it seems unlikely that any of these units persist as far north as the areas mapped by Reed (1955) and Nickelsen (1956).

Distribution and thickness within the mapped area: The Catoctin Formation is exposed extensively in the central part of the area (J-1 to A-12 to K-12) in a belt that widens from 1.6 km at either end to 4.8 to 6.4 km in the middle. The Catoctin appears to have a maximum estimated thickness of about 2200 m (Plate 2); however, the top has been eroded throughout much of the area southeast of the crest of the Blue Ridge.

Age: The Catoctin overlies the Swift Run Formation where the latter is present. However, because the Swift Run is time-transgressive, the Catoctin Formation overlaps the Swift Run (Plate 4). Where the Swift Run Formation is absent, the Catoctin Formation rests directly and unconformably on the Virginia Blue Ridge Complex. The Catoctin Formation within the area appears to be conformably overlain by the Loudoun Formation (Plate 4).

This upper contact is gradational by interleaving and the base of the Loudoun is placed at the lowest occurrence of conglomeratic metaarkose beds.

Rankin and others (1969, p. 743) have dated the rhyolitic part of the upper Catoctin Formation in Pennsylvaniaand report a discordant Pb^{207}/Pb^{206} age of 700 my. Using the data for this sample plus data from two samples of the Mount Rogers volcanics and two samples of Grandfather Mountain volcanics, Rankin and others (1969) define a chord suggesting an original age of 820 my for all three volcanic groups. These volcanic groups have been correlated on the similarity of lithology and because they conformably and/or unconformably are overlain by the Early Cambrian Chilhowee Group. This correlation is tenuous, and the writer is not inclined to accept it without better evidence. Therefore, the discordant 700 my date is accepted as the minimal age of the upper part of the Catoctin Formation rather than the 820 my age. In either case, the Catoctin Formation is definitely Late Precambrian.

Lithology: In the following descriptions of the seven principal metasedimentary and pyroclastic phyllite units, the section in the Stony Creek Gorge area (G-9 to E-8) is considered as the type section for most of the units and is the basis for their numerical designation

(Plates 2 and 3). The coarser grained metasandstones resemble the Swift Run metasandstones, and the fine-grained metasiltstones resemble the Loudoun metasilstones. Similar pyroclastic phyllites are found in both the Swift Run and Loudoun formations.

Unit 1: At the type section (G-9), Unit 1 has a measured thickness of about 10 m. The lower 2 m are a dense, speckled pyroclastic phyllite similar to Unit 3. The pyroclastic phyllite is overlain by 8 m of both massive and crossbedded conglomeratic metasandstones. The conglomeratic sandstone beds are from 5 cm to about 1 m thick. Grain size of detrital material is less than 5 mm. Quartz is the main detrital material. Crossbedding in different beds (G-9) indicates transport from both southeasterly and northwesterly directions, suggesting two sources of the detritus.

South (G-10; F-11) and west (A-12) of the type section, only Unit 1 metasiltstone and pyroclastic phyllites were found. This observation suggests that the sandstone pinches out very rapidly, and in some places Unit 1 is entirely absent (D-11). Farther south in the Spruce Creek syncline (F-14 to H-12), a coarse-grained (3 to 5 mm) massively bedded conglomeratic sandstone is present. The thickness at this location probably does not exceed 2 m. Unit 1 pinches out east of the type section and is not present in the area between Stony

Creek fault (G-9 to A-13) and Pauls Creek (L-9) (Figure 2). It is present farther northeast in a narrow belt (L-10 to K-1). Unit 1 in this easterly belt consists principally of coarse to fine-grained (1 to 5 mm) conglomeratic quartz metasandstone. The measured thickness of this massive bed is 1.3 m (J-5).

In thin section, the metasandstones and metasiltstones differ in the amount of coarse clastic quartz, which constitutes as much as 70 percent of the rock. The quartz is primarily subrounded to rounded. A few rock fragments (microcrystalline quartz) and grains of perthite, microcline and plagioclase are present. The heavy minerals and the opaque iron oxides (up to 10 percent) are subrounded detrital zircons and garnets. The latter two comprise about 1 percent of the rock. The matrix of the rock consists of a quartz-sericite (70 percent of matrix) groundmass intergrown with about 20 percent fibrous actinolite and 10 percent fine-grained epidote (?). The matrix has optically preferred orientation. Optical and x-ray data on the Unit 1 pyroclastic phyllite is summarized in Tables VIII and IX.

<u>Unit 2</u>: At the typical section, the base of Unit 2 consists of about 8 m of coarse-grained (3 to 5 mm) metasandstone, which contains pebbles as large as 2.5 cm in length. Most of the detrital grains are quartz, and most

of the massive beds are 50 to 100 cm thick. Above the basal metasandstone is about 15 m of pyroclastic phyllite. This phyllite, like that in Unit 3, is fine grained with a speckled surface and a silky sheen. The pyroclastic phyllite is overlain by about 30 m of fine-grained conglomeratic metasandstone similar to that in Unit 1.

The distribution of Unit 2 is similar to that of Unit 1 (Figure 3). Unit 2 thins rapidly southward, until the measured thickness is about 3 m (G-10). At this locality, the lower 30 cm is pyroclastic phyllite, and the remainder of the unit is a massive, medium-grained (2 to 3 mm), quartz-rich conglomeratic metasandstone. In the northwestern quarter of G-11, Unit is represented also by about 3 m of the upper metasandstone. Unit 2 is absent in the Pond Hollow area (D-11), but farther west (A-12) the lower metasedimentary part is present along with the pyroclastic phyllite. The lower metasedimentary part consists of coarse-grained (3 to 10 mm), quartz-rich metaconglomerates and sandstones interbedded with very fine grained (less than 1 mm) siltstone-mudstone phyllites. The contact with the pyroclastic phyllite is gradational at this locality.

Unit 2 also is present in the Spruce Creek syncline, where it consists of both the upper and lower metasedimentary parts separated by a greenstone flow. The lower part consists of coarse-grained (3 to 5 mm), massively bedded metasandstones, which thin eastward from about 6 m to about 2 m. The upper part is lithologically similar but is thicker (3 to 15 m).

Unit 2 is absent in the area northeast of the Stony Creek fault (G-9 to I-12), but the lower metasedimentary part and the pyroclastic phyllite are present farther to the northeast (L-10 to K-1). The metasedimentary part is similar to that exposed in the Spruce Creek syncline, whereas the pyroclastic phyllite has numerous thin (1 to 5 mm) quartz layers and lenses interbedded with the pyroclastics. The pyroclastic part is about 2 m thick, whereas the metasandstone thickens northward from 0 (K-8) to about 3 m (J-5).

At two localities (A-12; L-3), a pyroclastic phyllite is present between Units 1 and 2. The phyllite is about 3 m thick at A-12. X-ray and optical data on the Unit 2 pyroclastics are included in Tables VIII and IX.

The metasedimentary parts of Unit 2, in thin section, are composed of 20 to 40 percent rounded to subangular quartz, 5 percent subangular perthite, plagioclase and microcrystalline rock fragments, 10 percent magnetite $\pm$ hematite, 5 to 10 percent epidote and minor amounts of chlorite, fibrous actinolite and zircon. The remaining 35 to 60 percent of the rock is mainly a sericite matrix exhibiting a well developed mineral alignment. The following replacements were observed:

plagioclase replaced by epidote plus quartz

plagioclase replaced by chlorite

perthite and quartz replaced by sericite

sericite and epidote replaced by fibrous actinolite

These replacements suggest initial alteration toward the greenschist assemblage of epidote, sericite, quartz and chlorite followed by actinolite replacement.

The pyroclastic part of Unit 2 consists of 20 to 35 percent microcrystalline quartz in blebs about 3 mm in length, 10 percent magnetite and 10 percent fine-grained epidote (?) in a matrix composed mainly of either sericite plus actinolite or sericite plus chlorite. The sericite generally comprises about two-thirds of the matrix.

Unit 3: This unit, throughout the area, consists of a dark-purple to greenish-gray, mottled, thinly foliated, very fine grained dense rock with white speckles and a silky sheen on the foliation surfaces. The unit is best exposed in the Crawford Knob area (I-12 to J-8 to K-9), where the thickness is 1 to 3 m. Unit 3 appears to pinch out eastward (K-8,9) (Figure 4). At the type section (G-8), it probably is 3 to 6 m thick. In A,B-12, Unit 3 has been correlated with the uppermost of three pyroclastic phyllites present above Unit 2; however, it could be represented by any of the three phyllites. These

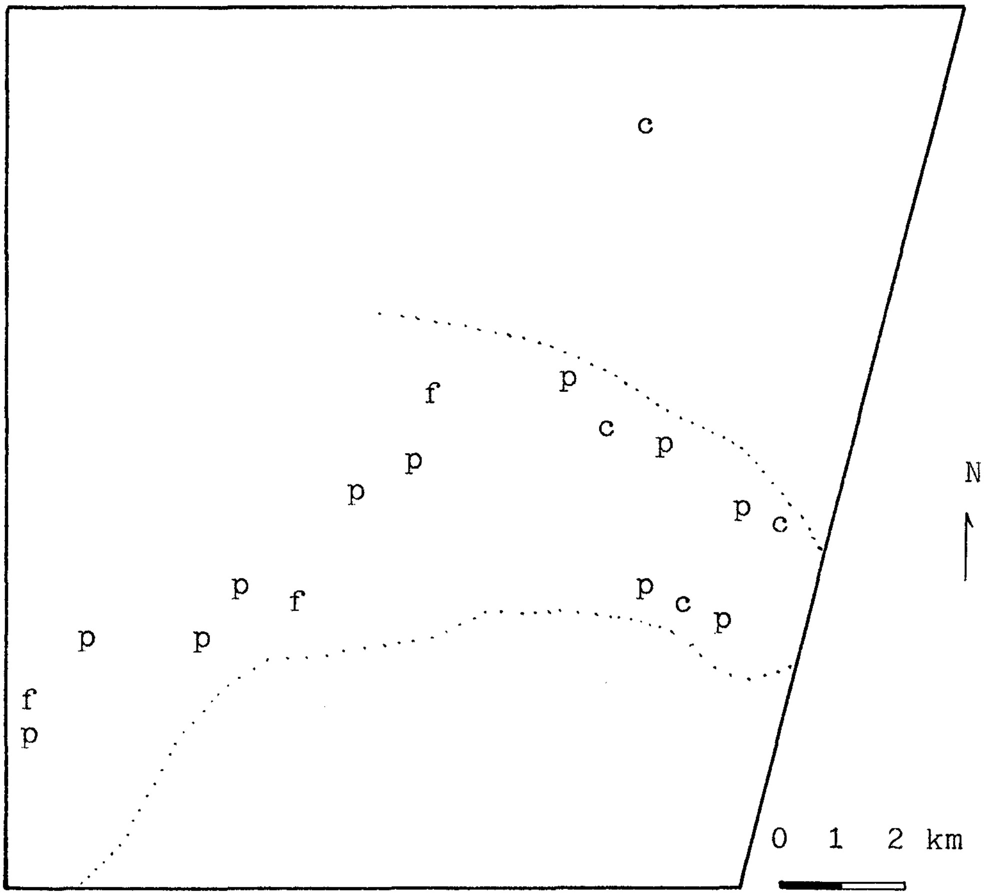

Figure 4. Distribution of Unit 3 and Unit 4

..... - pyroclastics pinch out
p - pyroclastics-Unit 3
f - fine-grained detritus-Unit 4
c - coarse-grained detritus-Unit 4

phyllites are 0.5 to 3 m thick. Tables VIII and IX summarize the x-ray and optical data on this unit.

Unit 4: Unit 4 is a poorly exposed conglomeratic metasandstone of unknown thickness. Its lithology is similar to those of Units 1 and 2, and it appears to be present throughout the area except where it has been removed by erosion. At the best exposure of Unit 4 (K-11), it consists of a measured thickness of 6 m, and it consists of both massive and torrentially crossbedded, coarse-grained (3 to 5 mm) conglomeratic metasandstones and conglomerates with quartz and feldspar fragments as large as 20 mm long and phyllite fragments up to 8 cm long. The beds are 30 to 100 cm thick. The crossbedding suggests transport from an easterly direction (Figure 4), and this observation agrees with the diminishing grain size to the north and west (Figure 4).

Unit 5: This unit is a pyroclastic phyllite easily recognized because of large "knots" of contorted quartz and phyllite present at every exposure visited. The knots are 2 to 8 cm in diameter and are probably caused by clasts of metasediemnts in the pyroclastics. The unit appears to be about 9 to 15 m thick in the Humpback Mountain region (F-7 to J-7) and thins southward to about 2 to 3 m on the south side of Crawford Knob (J-10 to K-11).

In thin section, Unit 5 appears to be composed of both sedimentary and pyroclastic material. The coarser sedimentary material, ranging in size from 1 to 5 mm, typically consists of about 30 percent quartz, 15 percent perthite feldspars along with some orthoclase and plagioclase, 15 percent magnetite, 2 percent epidote and 1 percent fine-grained epidote (?). The matrix consists of about 30 percent sericite. The feldspars typically are partially replaced by sericite. Optical and x-ray data on the pyroclastic portion of Unit 5 are included in Tables VIII and IX.

Unit 6: Unit 6 consists of about 2 to 3 m (F-8) of thinly laminated siltstone-mudstone phyllite. It is very similar to the siltstone-mudstone phyllites of the overlying Loudoun Formation. The laminae are 1 to 3 mm thick and consist of alternating dark gray, very fine-grained metamudstones and light gray, fine-grained (about 1 mm) metasiltstones. Unit 6 is found only between D-9 and J-1. It may be present in the folded section between the Reed Gap fault and the South Fork fault (A-12 to C-8), but Unit 6 probably pinches out to the northeast of this fault block.

In thin section, the metasiltstone layers consist of about 25 percent subangular to subrounded quartz with zircon inclusions along with 3 percent subangular magnetite, 2 percent epidote, and 1 percent each of subrounded to subangular zircon, fine-grained epidote (?), chlorite and msucovite. The remainder consists of a very fine grained quartz-sericite matrix exhibiting an optically preferred orientation. Many quartz grains have corroded rims, and only the outlines of replaced grains of quartz and feldspar (?) are present in the matrix.

The metamudstone layers consist of about 10 percent fine-grained opaques, 5 percent detrital quartz and 2 percent magnetite. The remainder of these layers consists of a quartz-sericite matrix similar to that in the metasiltstone.

Unit 7: Unit 7 is the uppermost metasedimentary unit of the Catoctin Formation and is found in a band from E-6 to J-1 and also in A-10,11. At the latter location, the unit consists of a conglomeratic metasandstone composed principally of quartz grains (up to 2.5 cm) but with some feldspar. It appears to be about 9 m thick. Unit 7 also crops out at J-3, where it consists of light gray, fine-grained (less than 1 mm) metasiltstone interbedded with greenish-gray to greenish-brown coarse-grained (3 to 5 mm) metasandstones. Bedding is 2 to 100 cm thick, and the coarser grained beds are massive. The thickness appears to be on the order of 6 to 9 m.

Greenstone lava flows: The Catoctin greenstones are divisible into two broad types, which are easily recognized in hand specimens and thin sections. These types are: (1) a light pastel green, fine-grained variety (epidosite) cut by quartz, epidote and/or actinolite veins; and (2) a dark gray-green, fine-grained variety with epidote, chlorite, quartz and/or calcite amygdules. A variation of the first type is a mottled type which consists of light green, fine-grained blotches or amygdules mixed with dark gray to reddish gray blotches with quartz amygdules, veins and blotches. Type Two greenstones are the most common. Type One is found typically as large bodies up to 3 m thick in well-exposed flows of Type Two. Type One also is found as large random outcrops and in a few instances as what appear to be individual random flows. Type One greenstones appear to be completely altered lavas becuase they contain no relict minerals, whereas Type Two contains relict plagioclase and/or pyroxene.

Greenstone contacts: Flow-on-flow contacts are poorly exposed and are not easily recognized, although on steep slopes they may be discerned by their stairstep or ledgy outcrop pattern of flows. Two flow-on-flow contacts were observed in the northeast quarter of D-11 and in the northwest quarter of H-6 immediately west of the Humpback Mountain tear-fault zone in the road cut on the Blue Ridge Parkway. At the D-11 locality, an 18 m thick Type One flow overlies a Type Two flow. At the H-6 location Type Two flows are in contact. The lower flow is characterized by large quartz-epidote amygdules up to 15 cm in length. The upper flow has none of these amygdules. The contact zone consisted of a ptygmatically folded light green layer about 15 cm thick. Immediately above the contact zone are several pods of Type One greenstone, which resemble breccias. The pods are about 2 m long and about 30 cm thick. The pods are capped by large hemispheric bodies of Type One greenstone which may be pillows. The bodies have radii of about 30 to 60 cm. The remainder of the upper flow appears to be typical Type Two.

Cobbles within flows: At one locality (J-8) near the base of the Catoctin, unakite cobbles up to 15 cm were found in the greenstone. These cobbles range from subrounded to subangular. The unakite consists principally of large quartz and pink feldspar grains (identified by x-ray as maximum microcline) partially replaced by finer grained epidote and microcrystalline quartz. Two sets of quartz veins cut through both the enclosing greenstone and the cobbles. Within the cobbles, the vein material as well as the unakite is partially replaced by sericite. The cobbles are surrounded by cherty contact zones about

2 mm wide. The unakite at the contact consists of very fine grained epidote and microcrystalline quartz in irregular pods and lenses penetrating inward from the contact zone to about 3 mm. The greenstone adjacent to the contact consists of very fine grained magnetite and epidote (?).

Greenstone flow mineralogy: The major mineralogic differences between Type One greenstone and Type Two greenstone is that the former typically has an estimated mode of 5 percent or less plagioclase, more than 20 percent quartz, 30 to 60 percent epidote, 10 percent iron oxides and no pyroxene, whereas Type Two consists of more than 25 percent plagioclase, 5 percent or less quartz, 5 to 30 percent epidote and 10 percent iron oxides (principally magnetite). The amounts of chlorite, fibrous actinolite, magnetite and pyroxene appear to depend on both stratigraphic depth and rock type. Most of the greenstones also contain about 10 to 15 percent fine-grained cloudy opaque clots with specks of highly birefringent grains within the masses which, on the basis of optical and x-ray data, are tentatively identified as fine-grained epidote (?). Calcite is present in one sample (130) from a ledge above Unit 3. In this sample, calcite comprises about 20 to 25 percent of the rock and is present in both amygdules and in the fine-grained Type

Two matrix.

Tables X, XI, XII, and XIII summarize the compositions of the two types of greenstones. Point counts were made from two Type One greenstone samples, 33B and 34. The mode of sample 33B, as determined by point counting with a high power objective, is 13 percent quartz, 79.5 percent epidote plus fine-grained epidote (?), 2.5 percent penninite, 4 percent magnetite plus hematite and 2.5 percent actinolite. The estimated mode (Table XI) as determined with a low power objective was about 10 percent low on the epidote plus fine-grained epidote and 7 percent and 3 percent high on the penninite and actinolite respectively.

The mode of sample 34, as determined by point counting, is 31.5 percent quartz, 38.5 percent epidote, 14 percent fine-grained epidote and 16 percent magnetite plus hematite. The principal difference between this mode and the estimated mode (Table XI) is that with the low power objective the opaques appear to be all magnetite (30 percent-Table XI), whereas with the high power objective the opaques were determined to be both magnetite (16 percent) and very fine-grained epidote (14 percent).

Four Type Two greenstone samples, 11, 49, 78A and 131B, were point counted. The modes determined by point counts are given below with the corresponding estimated

TABLE X. ESTIMATED MODAL COMPOSITION OF CATOCTIN FORMATION TYPE ONE GREENSTONES--ZONES A AND B.

Sample Number	Zone	Quartz	Epidote	Fine-grained Epidote (?)	Plagioclase	Penninite	Magnetite	Hematite	Actinolite
53A	A	30	--	70	--	--	--	--	--
114A-2	A	10	60	15	--	--	15	--	--
114A-4	A	15	60	10	--	--	10	2	--
114B	A	35	30	17	--	--	5	10	2
289	A	10	20	15	20	10	10	10	5
14-1	B	30	50	5	--	--	3	10	--
14-2	B	40	40	5	--	--	3	12	--
15	B	--	55	5	10	25	2	--	3
16-1	B	25	55	3	--	--	3	8	3
22	B	25	60	--	--	5	5	5	--
50A	B	30	50	--	--	--	10	5	--
50B	B	60	30	--	--	--	5	--	5
50C	B	60	15	--	--	--	15	--	10
51A-1	B	25	40	5	--	25	5	--	--
51A-2	B	50	40	8	--	--	2	--	--
51A-3	B	50	40	5	--	--	5	--	--
51B-1	B	50	40	--	--	--	10	--	--
51B-2	B	20	60	5	--	15	--	--	--
51B-3	B	25	45	2	--	20	2	5	--
51C	B	25	40	2	--	25	2	5	--
78B	B	50	50	--	--	--	--	--	X
79A	B	40	50	--	--	--	10	--	--
79B	B	15	65	10	--	--	5	--	5

X - 1 percent or less

TABLE XI. ESTIMATED MODAL COMPOSITION OF CATOCTIN FORMATION TYPE ONE GREENSTONES--ZONES C AND D.

Sample Number	Zone	Quartz	Epidote	Fine-grained Epidote (?)	Plagioclase	Penninite	Magnetite	Hematite	Actinolite
29	C	10	60	10	5	5	10	--	--
29A	C	25	40	10	5	--	10	5	2
29B	C	25	25	20	5	--	10	X	10
69	C	30	60	--	--	--	2	5	3
73A	C	15	55	--	10	--	5	5	5
73B	C	5	40	10	20	10	10	--	5
81	C	25	50	15	--	--	5	--	X
131C	C	35	30	10	--	--	X	20	--
28A	D	50	20	15	--	--	5	10	--
28B	D	60	15	10	--	--	10	5	--
33B	D	10	45	25	--	10	5	--	5
34	D	30	35	--	5	--	30	--	--
70	D	20	60	--	--	12	--	5	--
178	D	25	55	--	--	20	--	--	--

X - 1 percent or less

TABLE XII. ESTIMATED MODAL COMPOSITION OF CATOCTIN FORMATION TYPE TWO GREENSTONES--ZONES A AND B.

Sample Number	Zone	Quartz	Epidote	Fine-grained Epidote (?)	Plagioclase	Penninite	Magnetite	Hematite	Actinolite	Pyroxene	Calcite
53B	A	2	30	10	30	5	--	--	27	--	--
59	A	25	35	5	--	--	5	15	--	--	--
60	A	20	40	--	20	2	5	--	11	--	--
65	A	--	--	15	45	10	10	--	20	--	--
117	A	2	15	--	30	2	10	--	32	--	--
126	A	--	20	--	2	35	5	--	35	--	--
137B	A	--	10	15	40	2	10	5	17	--	--
214	A	20	20	10	--	3	--	--	45	--	--
227	A	--	10	45	--	5	15	--	25	--	--
228B	A	--	10	30	35	5	--	2	15	--	--
288	A	--	10	20	25	--	10	--	30	--	--
296	A	--	20	5	35	10	15	--	15	--	--
16-2	B	--	10	15	40	15	10	--	10	--	--
20-1	B	15	55	5	15	--	5	--	2	--	--
20-2	B	5	50	5	25	--	5	--	10	--	--
21	B	--	3	10	45	3	5	--	25	--	--
45	B	--	15	13	40	10	5	--	15	--	--
63	B	5	10	15	20	25	3	--	22	--	--
64	B	5	30	10	30	15	5	--	5	--	--
78A	B	--	15	15	30	5	5	2	21	--	--
124C	B	20	30	10	--	5	2	--	25	5	--
130A	B	--	3	25	40	5	20	--	10	--	--
139	B	--	5	15	40	10	5	--	22	--	--

TABLE XIII. ESTIMATED MODAL COMPOSITION OF CATOCTIN FORMATION GREENSTONES--ZONES C AND D AND DIKE.

Sample Number	Zone	Quartz	Epidote	Fine-grained Epidote (?)	Plagioclase	Penninite	Magnetite	Hematite	Actinolite	Pyroxene	Calcite
6	C	2	10	5	30	25	10	--	X	5	--
7A	C	--	10	15	30	15	10	--	5	--	--
7B	C	--	15	--	25	15	20	10	5	2	--
8	C	2	10	15	20	10	15	5	7	5	--
11	C	X	5	20	20	20	15	5	11	--	--
12	C	--	5	20	50	5	10	--	2	5	--
24	C	2	30	10	45	--	10	--	--	5	--
46	C	2	30	15	20	10	10	--	5	2	--
47	C	--	10	15	50	10	10	--	2	2	--
48	C	2	--	25	25	10	15	--	X	5	--
49	C	2	2	10	60	10	5	--	2	2	--
66	C	--	3	15	60	3	10	--	--	3	--
67	C	--	10	10	60	5	10	5	--	--	--
68	C	--	10	20	40	5	10	--	10	X	--
71	C	--	5	30	25	10	20	--	5	5	--
72	C	--	5	25	15	10	10	3	12	2	--
131A	C	5	5	15	30	20	5	--	--	--	20
131B	C	15	2	5	5	40	10	--	--	--	20
177	C	--	25	--	40	3	10	--	3	--	--
9	D	--	60	--	15	5	10	--	2	5	--
10	D	10	15	15	25	25	10	--	--	--	--
23	D	5	15	15	20	25	10	5	--	--	--
25	D	--	40	10	5	25	10	X	--	--	--
26	D	--	--	25	30	15	15	5	2	5	--
27	D	5	20	20	20	20	15	X	X	--	--
30	D	--	--	5	50	30	5	--	5	5	--
31	D	--	30	--	20	30	5	--	10	5	--
32	D	10	10	--	40	10	15	--	--	5	--
33	D	--	25	--	25	15	20	--	12	--	--
36	D	10	30	15	15	15	15	--	--	--	--
37	D	--	35	--	25	25	10	--	--	5	--
38	D	5	15	20	35	10	10	--	X	5	--
123A	dike	25	--	10	30	--	2	--	20	--	--
123B	dike	30	2	15	--	40	X	--	10	--	--
123C	dike	35	--	--	20	35	X	X	--	5	--

X - 1 percent or less

modes from Tables XII and XIII in parentheses. The mode of sample 11 is 8.5 percent epidote (5), 22 percent fine-grained epidote (20), 36 percent plagioclase (20), 5 percent penninite (20), 10 percent magnetite plus hematite (20) and 18.5 percent actinolite (11). The mode of sample 49 is 8 percent quartz (2), 7 percent epidote (2), 17.5 percent fine-grained epidote (10), 36 percent plagioclase (60), 8.5 percent penninite (10), 16 percent magnetite plus hematite (5), 4.5 percent actinolite (2), 3 percent pyroxene (2). The mode of sample 78A is 20.5 percent epidote (15), 12 percent fine-grained epidote (15), 32.5 percent plagioclase (30), 1.5 percent penninite (5), 13 percent magnetite plus hematite (7), 20.5 percent actinolite (21). The mode of sample 131B is 13 percent quartz (15), 6 percent epidote (2), 11.5 percent fine-grained epidote (5), 2 percent plagioclase (5), 35.5 percent penninite (40), 9 percent magnetite (10), and 24 percent calcite (20).

The estimated mode for samples 78A and 131B agree with the point count modes within 6.5 percent for all minerals.

The estimated mode for sample 49 is significantly different from the point count mode in the amount of plagioclase present. The higher estimated value resulted from underestimation of all the other minerals present which were subtracted from 100 to obtain the plagioclase content. The estimated and point count modes of sample 11 match poorly with respect to the plagioclase, actinolite, penninite, and magnetite plus hematite content. The latter two are estimated low whereas the first two are estimated high. Penninite is present principally as amygdules which are not evenly distributed, hence, even though up to 600 grains were counted, the number of grains may have been insufficient to accurately determine the penninite content. Another possible source of error in estimating the penninite content is the fact that actinolite and penninite are frequently intergrown and are difficult to distinguish with a low power objective. The error in plagioclase content resulted in the same manner as did the error in plagioclase content of sample 49. The magnetite plus hematite content was overestimated, probably because of operator bias.

Table XIV summarizes the dominant minerals present in x-ray patterns of representative greenstones of both types. Table XV is a summation of the data used to determine plagioclase composition in the greenstones. In general, relict plagioclases appear to hava a composition of oligoclase-andesine and are recognized from

TABLE XIV. X-RAY MINERALOGY OF CATOCTIN FORMATION GREENSTONES.

Sample Number	Zone	Type	Quartz	Epidote	Plagioclase	Penninite	Magnetite	Actinolite	Calcite
114A	A	1	X	X	-	X	X	X	-
53A	A	1	X	X	-	-	?	X	-
50A (light part)	B	1	X	X	-	-	-	-	-
50A (dark part)	B	1	X	X	-	-	-	-	-
73	C	1	X	X	-	-	-	X	-
33B	D	1	X	X	-	-	-	-	-
70	D	1	X	X	-	X	-	-	-
53A	A	2	X	X	-	X	X	X	-
126	A	2	X	-	X	X	X	X	-
214	A	2	X	X	X	X	X	X	-
228B	A	2	?	X	X	X	X	X	-
78A	B	2	X	X	X	X	X	X	-
130A	B	2	X	-	X	X	X	X	-
11	C	2	X	-	X	X	X	X	-
67	C	2	X	X	X	X	X	X	-
131A (amygdule)	C	2	X	-	-	-	-	-	X
131A	C	2	X	-	X	X	X	-	X
26	D	2	X	X	X	X	X	X	-
31	D	2	X	-	X	X	X	-	-
38	D	2	X	X	X	X	X	X	-
447 (amygdule)	dike	2	X	-	-	X	X	-	-
447	dike	2	X	X	X	X	X	X	-
123A	dike	2	X	X	X	X	X	X	-

TABLE XV. GREENSTONE PLAGIOCLASE COMPOSITION.

Sample Number	Zone	Optical determinations of relict plagioclase (principally Carlsbad-Albite twins)	X-ray determination of dominant plagioclase ((1$\bar{3}$1-131 °20 separation)
447	dike	--	1.1
126	A	--	1.0
228B	A	--	1.0
78A	B	An_{28}	0.9
130A	B	An_{28}	0.9
139	B	An_{31}	---
11	C	--	1.0 (?)
67	C	--	1.0
131A	C	--	1.1 (?)
131B	C	An_{33}	---
26	D	--	1.0
31	D	--	0.9
38	D	An_{32}	0.9

albite by their larger size and partially altered character. The dominant plagioclase is much finer grained and, from the x-ray data, corresponds to low albite. Bloomer and Werner (1955, p. 591) report the relict plagioclase composition andesine (An_{40}) and the neomineralic plagioclase albite (An_{10}) and oligoclase (An_{20}).

Table XVI summarizes the average mineralogical changes which have occurred with increasing depth of burial. These averages were obtained by first dividing the Catoctin Formation greenstones into four stratigraphic divisions. The lowest division (A) includes the greenstones between the Swift Run Formation and Unit 1. Division B encompasses the greenstones between Unit 1 and Unit 3, and division C includes the greenstones between Units 3 and 6. The greenstones between Unit 6 and the base of the Loudoun Formation belong to division D. The thin sections of the greenstones for each division were examined, and the estimated percentages then were averaged by adding the percentage of each mineral present and dividing the total by the total number of thin sections examined. From comparison of the estimated and point counted modes, it is believed that the errors are sufficiently random not to appreciably affect the results of averaging the estimated modes.

It is apparent from Table XVI that in the Type Two

TABLE XVI. COMPARISON OF AVERAGE ESTIMATED MINERALOGY (VOLUME PERCENT) OF TYPES ONE, TWO AND DIKE GREENSTONES.

Division	Number of Thin Sections	Epidote	Fine-grained Epidote (?)	Quartz	Plagioclase	Pyroxene	Penninite	Magnetite	Hematite	Actinolite
Type one										
A	5	34	25	20	4	0	2	8	4	1
B	18	46	3	34	0	0	6	5	3	1
C	8	45	9	21	6	0	2	6	4	3
D	6	38	8	33	1	0	7	8	3	1
Total	37	43	8	29	2	0	5	6	3	2
Type two										
A	12	18	13	5	22	0	6	7	2	23
B	11	20	12	7	30	0	8	6	0	15
C	19	10	14	2	34	2	12	11	1	4
D	13	23	10	3	25	3	19	11	1	2
Total	55	17	12	4	28	1	12	10	1	10
Dike	3	1	8	30	17	2	25	1	0	10

greenstones the percentages of fibrous actinolite increase with depth at the expense of pyroxene, chlorite and magnetite. However, no systematic changes with depth appear to occur in the Type One greenstones.

The following replacements were observed in thin section:

(1) plagioclase replaced by epidote and quartz

(2) plagioclase replaced by penninite

(3) epidote replaced by penninite

(4) pyroxene replaced by penninite

(5) plagioclase replaced by fibrous actinolite

(6) epidote replaced by fibrous actinolite

(7) pyroxene replaced by fibrous actinolite

(8) penninite replaced by fibrous actinolite

This series of metamorphic alterations suggests initial alteration of the lavas to the greenschist assemblage of epidote, quartz, penninite and albite and subsequent alteration toward the epidote, quartz, albite and actinolite assemblage.

Amygdules: Amygdules present in the Type One greenstones usually consist of: (1) quartz and epidote ; (2) quartz; or (3) epidote. Amygdules in the Type Two greenstones are more variable in composition and consist of the following types.

1a) penninite

b) penninite rim with quartz interior

c) penninite rim with calcite-quartz-epidote interior

2a) epidote

b) epidote and quartz

c) epidote rim with penninite interior

d) epidote rim with inner penninite-quartz rim with fibrous actinolite interior

3a) microcrystalline quartz rim with epidote and quartz interior

b) microcrystalline quartz rim with inner epidote rim with penninite-quartz-epidote $\pm$ fibrous actinolite interior

c) microcrystalline quartz rim with penninite-quartz $\pm$ fibrous actinolite interior

Greenstone dikes: Two large greenstone dikes were mapped (G-12, N-7). Both dikes appear to be 3 to 6 m wide, and neither was traceable for any distance beyond the outcrops. The G-12 location, however, appears to have been a major site of Catoctin intrusion inasmuch as numerous smaller dikes up to about 4 m criss-cross Little Stony Creek for a distance of about 0.6 km upstream from the large dike.

Greenstone dike mineralogy: In hand specimens, the dike rock is a dark greenish-gray, fine-grained rock resembling the Type Two greenstone flow rocks. The mineralogy of the three samples is given in Tables XIII, XIV, and XV.

The contact zone between the greenstone dikes and the granite country rocks consists of fibrous actinolite on the greenstone side and fine-grained epidote on the granite part. The actual contact zone is seldom more than 2 to 3 mm. Because of the extremely brecciated nature of the granite, however, the contacts appear wider.

The only replacements observed in thin section were:

(1) plagioclase replaced by epidote and quartz

(2) plagioclase replaced by penninite

(3) plagioclase replaced by fibrous actinolite

(4) pyroxene replaced by penninite

These alterations are compatible with the greenschist assemblages suggested previously toward which the lavas are partially altered.

Metamorphic alterations: In general, the following three stages of partial replacement are recognized in thin section:

(1) plagioclase replaced by epidote and quartz

(2) plagioclase, epidote and pyroxene replaced by penninite

(3) plagioclase, epidote, quartz, pyroxene and penninite replaced by fibrous actinolite

In addition, the following composite temporal sequence was obtained from direct examination of all thin sections:

(1) alteration of plagioclase to epidote and quartz with some epidote veins formed during stage 1.

(2) older quartz veins which cut both epidote crystals and epidote veins during stage 1.

(3) younger quartz veins which cut both epidote veins and older quartz veins during stage 1.

(4) replacement of plagioclase, epidote, vein quartz and pyroxene by penninite, during stage 2, and by actinolite, during stage 3.

In 13 thin sections, a secondary crinkle foliation is apparent. In several of these thin sections, it is obvious that the actinolite replacement is closely related to the formation of the secondary crinkle foliation. Virtually all quartz grains present in all the greenstone are highly undulatory indicating some tectonic activity after metamorphism.

Loudoun Formation

Name and age: The Loudoun Formation was described by Keith (1892, p. 365) and named by him (*in* Williams and Clark, 1893, p. 68) for rocks in Loudoun County, Virginia. Despite the ambiguities in nomenclature and in designation of type sections (Cloos, 1951, p. 29; King 1950a, p. 16), it is quite clear that Keith (1893, p. 324, 327-329) considered the Loudoun Formation to consist of the heterogeneous succession of sedimentary rocks which lies immediately beneath the Weverton Formation and, generally, on top of the Catoctin Formation. Likewise, it was the intention of Stose and Stose (1946, p. 31-34) in their revised, restricted definition of the Loudoun that the name be applied to the beds between the Catoctin and Weverton formations. King (1950a, p. 16-17) applied the name to rocks in this stratigraphic position in the Elkton area, and Reed (1955, p. 878-879) includes these Loudoun rocks in the Catoctin Formation. These rocks are similar to the heterogeneous assemblage within the study area; hence the name Loudoun will be used herein to indicate the uppermost part of the Catoctin Group, which is immediately overlain by the Weverton Formation.

<u>Distribution</u>: Within the study area, the Loudoun Formation crops out along a narrow faulted belt on the western slope of the Blue Ridge (A-11 to J-1). Based on the lithologies present, the belt can be divided into a northern and a southern portion. The change in lithology occurs in the vicinity of Toms Branch (H-3).

Lithology: North of Toms Branch (H-3 to J-1), the Loudoun Formation consists almost exclusively of siltstone phyllite with a few thin lenses and beds of pebble conglomerates and a pyroclastic phyllite at the top. The thickness of this northern portion is estimated to be 150 to 210 m.

The siltstone phyllite is mainly interbedded fine-grained siltstone and mudstone layers averaging about 2 to 5 mm in thickness. The largest grains in these layers are less than 1 mm in diameter. In thin section, the siltstone phyllite consists of about 23 percent fine-grained material and about 77 percent recognized detrital material. The matrix is mad up of about equal parts of sericite and microcrystalline quartz. The sericite exhibits an optically preferred orientation parallel to the plane of foliation indicating a metamorphic source.

The subangular and subrounded clasts (in total rock percentages) are made up of about 40 percent rock fragments, 25 percent orthoclase, 11 percent magnetite and fine-grained opaques and 1 percent heavy minerals. The rock fragments, of metamorphic origin, consist of interlocking grains of microcrystalline quartz and sericite with some coarse-grained muscovite, zircon, garnet (?) and orthoclase. These phyllite fragments range up to

5 mm in length.

The orthoclase is subangular to subrounded, and has slightly undulatory extinction. The grains are about 3 mm in diameter and contain trains of inclusions of garnets (?) and other very fine grained unidentified minerals. The orthoclase grains are partially replaced by sericite and quartz. The magnetite grains are subangular to subrounded and are concentrated in thin layers along with minor amounts of fine-grained opaques, tourmaline, zircon and garnet.

The pebble conglomerate layers are 25 to 150 mm thick and are composed mainly of quartz pebbles ranging up to 20 mm in length (average-5 mm). In thin section, these beds are composed of about 75 percent recognized detrital material and 25 percent fine-grained sericite and microcrystalline quartz. The sericite has a preferred orientation parallel to the foliation.

The clasts are mainly subangular to rounded rock fragments (35 percent), quartz (25 percent), and magnetite (15 percent). The rock fragments include both coarse and fine-grained metamorphic quartzites and quartz-sericite phyllite with or without euhedral garnets and/or zircon crystals. The quartz clasts appear to have been, at least partially, from the same source as the rock fragments. The magnetite occurs mainly as smaller subangular grains associated with minor amounts of fine-grained opaques and hematite.

Southward from Toms Branch (F-5,6), the main rock types in the 300 m section of Loudoun change from siltstone phyllites with conglomeratic layers to coarse arkosic metaconglomerates and siltstone phyllites interlayered with greenstones. A pyroclastic phyllite is present also at the top of the formation in the vicinity of Toms Branch. At the extreme southwestward edge of the map area (A-11), the Loudoun Formation is much thinner and is composed principally of alternating 2 to 3 m thick beds of finer grained, crossbedded and massive arkoses and siltstone phyllites. The maximum feldspar grain size within the arkoses, which comprise the lower part of the formation, decreases southwestward from about 20 mm (F-5,6) to 5 mm (A,B-10) to less than 3 mm (A-11) (Figure 5). Crossbedding (A-11) also suggests transport from the northeast toward the southwest (Figure 5). The arkoses exposed along F-5,6 are composed of 40 to 60 percent coarse-grained feldspars, 10 to 40 percent coarse-grained quartz, 10 percent coarse-grained rock fragments and 10 to 30 percent matrix material, accessory minerals and alteration products.

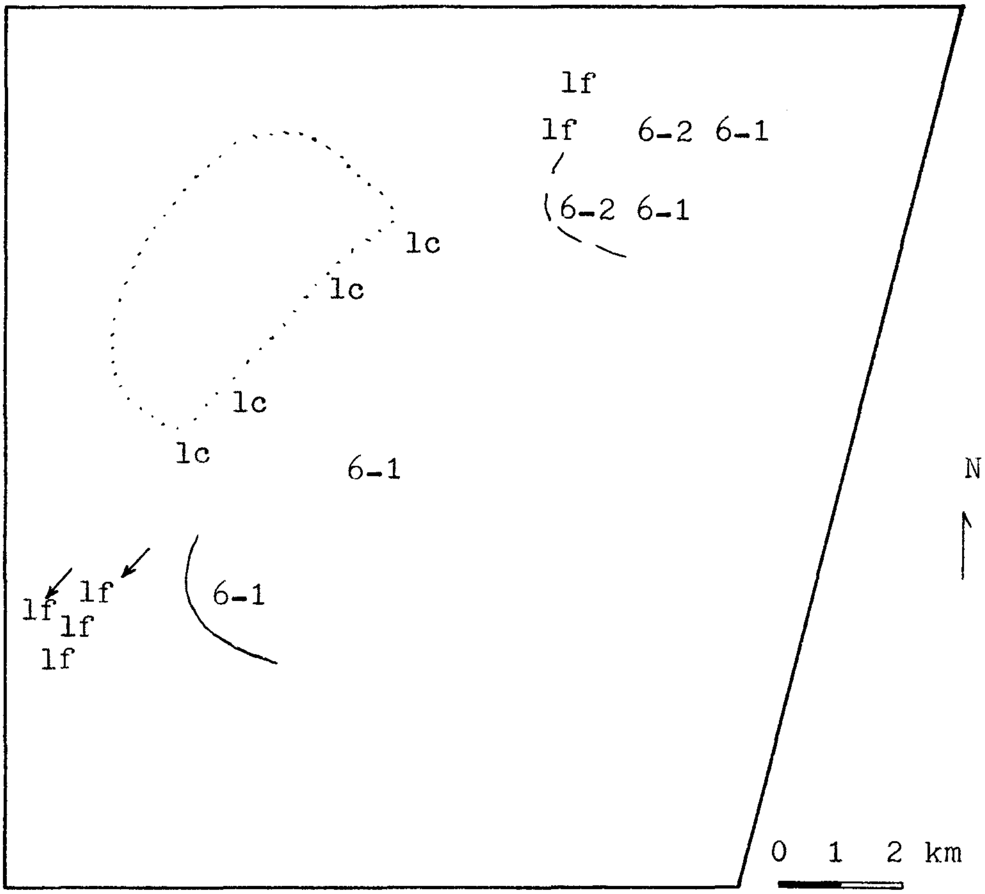

Figure 5. Distribution of Unit 6 and Loudoun

lc - Loudoun coarse-grained arkosic detritus
lf - Loudoun fine-grained arkosic detritus

.:.:. - probable localized source of Loudoun arkosic detritus
↖ - direction of transport of Loudoun detritus
6-1 - Unit 6 arkosic detritus
——— - Unit 6 pinches out southward
6-2 - arkosic detritus between Units 6 and 7
– – – - 6-2 pinches out southward

The large feldspar grains are about 3 to 5 mm in diameter and consist of orthoclase and a pink feldspar with a perthite-like texture. The latter was identified as maximum microcline by x-ray. These larger grains are subangular to rounded. Grid-twinned microcline and plagioclase grains are not as abundant (25 percent of all feldspars) and are smaller than the orthoclase and perthite-like microcline grains. Their size is about 1 mm in diameter, and their shapes are usually angular. Most of the fledspars have both zircon and quartz inclusions.

The quartz grains typically are rounded to subrounded, and their size range is about the same as the orthoclase and perthite-like microcline grains. The quartz is undulose and contains inclusions of zircon, rutile (?), and apatite (?), as well as finer grained unidentified material.

The rock fragments are primarily of metamorphic origin and consist mainly of quartzite, with or without garnet and epidote; quartz-sericite phyllite; and quartz-perthite-plagioclase rocks. The rock fragments are rounded to subrounded.

The remaining portion of the metaarkosic rocks consists principally of euhedral crystals of metamorphic epidote (40 to 70 percent of the matrix); microcrystalline quartz (10 to 20 percent); magnetite-hematite (10 percent) and fine-grained opaques (10 percent). Penninite is present in some of the rocks as a minor constituent.

The arkoses exposed in A-10 differ from the more northerly arkoses in that the grains are much finer, but with a very wide range (1 to 10 mm). These grains are very poorly sorted, and shapes range from angular to rounded, with the majority generally subangular. Color varies with composition and ranges from pinkish-purple to greenish-gray. The quartz content is about 40 percent, and the quartz is well rounded. Many of the quartz grains have unidentified inclusions. If the rock has little matrix, the grains are welded together with overgrowths. In general, "perthitic" and grid-twinned microcline are more abundant and have not been altered as much as plagioclase and orthoclase. Biotite, muscovite, zircon and sphene are all present in minor amounts. Quartz and epidote veins are present in thin section 277B. Rocks with a sericite and microcrystalline quartz matrix typically have a well developed mineral foliation.

The siltstone phyllites in the northern section (F-5,6) and in the southern section (A-10), where they were exposed along a new road cut, are lithologically similar and consist of purple-spotted, silvery-gray or greenish-gray, fine-grained (about 1 mm) laminated rocks. The laminae range from 1 to 5 mm in thickness and are well defined by layers, alternately rich then poor in heavy minerals. Most detrital grains are subangular to subrounded. Sorting is medium. Some rock fragments, principally quartzites, are present.

The rocks consist principally of 30 to 70 percent (average-about 50 percent) sericite and microcrystalline quartz matrix; 10 to 15 percent magnetite, hematite and fine-grained opaques; 10 to 20 percent larger (about 1 mm) quartz grains; 15 to 30 percent feldspars (orthoclase, plagioclase, perthite and grid-twinned microcline); less than 5 percent epidote and less than 1 percent heavy minerals (mainly tourmaline and zircon). In addition to the above, muscovite and/or yellowish-brown or golden-brown stained msucovite (?) may be present locally in amounts up to 15 percent.

Metamorphic alterations: The following mineral transitions were noted:

(1) orthoclase replaced by sericite, microcrystalline quartz, epidote and penninite

(2) plagioclase replaced by sericite, microcrystalline quartz and epidote

(3) "perthitic" microcline replaced by sericite, microcrystalline quartz and epidote

(4) magnetite replaced by hematite

(5) epidote replaced by actinolite

These partial alterations appear compatible with the greenschist assemblage present in the underlying Catoctin Formation greenstones, pyroclastics and metasediments. In addition to the preceding transitions, albite (?) appears to have formed, along with sericite, in situ in thin section 281M.

Sources of Loudoun metasediments: As stated previously, feldspar grain size decreases southwestward, and crossbedding (A-11) suggests transport from the northeast toward the southwest (Figure 5). Hence, because the coarse-grained feldspars do not occur northeast of Toms Branch, a local source must have existed near Toms Branch (F-5,6) (Figure 5). Moreover, this local source must have been to the west (Figure 5) inasmuch as 2100 m of Catoctin volcanics are present to the east, and no Catoctin greenstone rock fragments were found in the metaarkose.

The siltstone-phyllite detritus of both the Loudoun and upper Catoctin formations probably also originated from a source west of the Blue Ridge, and like the metaarkoses, this detritus probably was transported southwestward more or less parallel to the Blue Ridge (Figure 5). The principal reason for this interpretation is that the siltstone phyllites appear to be the fine-grained equivalents of the metaarkoses and, hence, have originated in a similar manner but have been transported farther.

Pyroclastic phyllite unit: The pyroclastic phyllite (J-1) consists of 20 percent quartz-blebs and 80 percent matrix. The quartz-blebs are 5 to 10 mm in length and, in thin section, are composed of about 75 percent microcrystalline quartz and 25 percent sericite. The blebs are elongated parallel to a well developed foliation. Composition and mineralogy of this pyroclastic phyllite are summarized in Tables VIII and IX. No accessory heavy minerals were noted.

The pyroclastic phyllite (E-6) at the top of the Loudoun (F-5,6) is similar to the one previously described. The blebs comprise about 20 percent of the rock and are made up of microcrystalline quartz with minor fine-grained opaques. One bleb consists of a crystalline quartz center surrounded by euhedral crystals of epidote. No accessory heavy minerals were identified.

Greenstones: The Loudoun greenstones are thinner, more weathered and not as well exposed as the underlying Catoctin greenstones. Columnar jointing was noted in the saprolite of one Loudoun (or upper Catoctin) greenstone lava flow (A-10). Study of one thin section of the light green variety indicates that the fine-grained rock consists primarily of 65 percent epidote, 15 percent quartz, 15 percent magnetite, hematite and fine-grained epidote (?) and 5 percent actinolite. The epidote occurs in euhedral crystals, whereas the quartz typically fills in the space between crystals. Actinolite has grown at the expense of both the epidote and quartz. The opaque minerals are more or less randomly distributed. Veins of quartz (only slightly undulose) cut across the rock. These veins are about 1 to 2 mm wide. This rock is similar to the Type One (epidosite) greenstone of the Catoctin Formation.

Chemical composition: Table XVII is a list of calculated chemical compositions of Catoctin Group pyroclastics and greenstones. The chemical compositions were calculated for the eight samples for which point count modes were available. Reed's average of four chemical analyses (Reed, 1964, p. C71) is included for comparison. Samples 78A, 11 and 49, which are Type Two greenstones, all have calculated compositions that are similar to Reed's average composition for the Type Two greenstones.

Sample 131B is a calcite-rich Type Two greenstone. The calculated composition shows a relative decrease (compared with typical Type Two greenstones) in SiO_2, Al_2O_3, Fe_2O_3, FeO and Na_2O and a relative increase in MgO, CaO, H_2O and CO_2. Samples 33B and 34, typical

Type One greenstones, show an increase in CaO at the expense of MgO and Na_2O.

From these calculated compositions it is apparent that either:

(1) Type Two greenstones are similar to the original bulk chemistry of the lavas, and Type One and calcite-rich Type Two greenstones represent localized alterations; or

(2) The original bulk composition was similar to some mixture of Types One and Two greenstones, and both types represent substantial alteration and segregation of the original components.

In the first case, chemical analyses such as that done by Reed (1964) may be useful in determining original bulk chemistry. However, if the second case is true, then only analyses of the two combined rock types would yield compositions similar to the original bulk chemistry. In either case, the origin of Type One greenstones must first be determined before chemical analyses are used to determine original bulk chemistry.

Two calculated chemical compositions of pyroclastic phyllites are included in Table XVII. These compositions (for samples 91 and 237) show that the pyroclastics have SiO_2, Fe_2O_3, and FeO contents similar to the greenstones but are depleted in MgO, CaO, and Na_2O. They also have

TABLE XVII. CALCULATED CHEMICAL COMPOSITIONS OF PYROCLASTIC PHYLLITES, TYPE ONE GREENSTONES, AND TYPE TWO GREENSTONES.

Sample Number	Rock Type	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	H_2O	CO_2
91	pyroclastic	45%	20%	8%	8%	---	---	---	8%	12%	---
237	pyroclastic	48%	17%	10%	9%	---	2%	---	6%	10%	---
33B	type 1 greenstone	54%	14%	8%	2%	1%	14%	---	---	7%	---
34	type 1 greenstone	58%	9%	12%	8%	---	9%	---	---	4%	---
131B	calcite rich type 2 greenstone	35%	8%	7%	6%	12%	15%	2%	---	6%	12%
78A	type 2 greenstone	51%	10%	9%	10%	4%	8%	4%	---	4%	---
11	type 2 greenstone	53%	10%	8%	8%	5%	7%	5%	---	4%	---
49	type 2 greenstone	54%	10%	10%	10%	4%	5%	4%	---	3%	---
average of 4 chemical analyses (Reed, 1964, p. C-71)	type 2 greenstone	47%	14%	6%	8%	6%	7%	4%	1%	3%	1%

K_2O and H_2O contents substantially higher than the greenstones.

Origin of the Catoctin Group

The following list represents the relative sequence of events relevant to the sedimentation, volcanism and metamorphism of the Catoctin Group. This list is primarily a summary and interpretation of the data presented previously.

(1) The Catoctin Group appears to have been formed along what may be the downdropped side of a northeast-trending *en echelon* normal fault system as evidenced by the rapid pinching out of the group westward (Plate 1). This appears to more or less coincide with the initial establishment of the Appalachian trend superimposed on the Middle Precambrian east-west structural trend. The entire Catoctin Group except for Swift Run metasaprolite is believed to be marine in origin because of the complete lack of greenstone detritus in the interlayered sediments.

(2) Initiation in Late Precambrian time of Swift Run sedimentation and volcanism. The Swift Run sediments appear, during this stage, to have been deposited in localized basins as indicated by rapid changes in thickness and pinch outs (Figure 1). This sedimentation was accompanied by movement along the Stony Creek fault,

which continued into Catoctin Formation time as indicated by the pinch out of lower Catoctin units (G-9) (Figures 1, 2, 3). Swift Run sediments are characterized by coarse clastics consisting of cobble and boulder size quartz and large chunks and fragments of phyllites (shales ?) in a sandstone matrix. The direction of transport as inferred from crossbedding, grain size and thickness probably was from the southeast quadrant (Figure 1).

Swift Run volcanism consisted of both lava flow and pyroclastic activity essentially identical with that which followed throughout Catoctin Group time. The lava flows probably erupted from fissures and dike swarms (G,H-12) (Plate 3). Hence, the lava flows are essentially *in situ*.

(3) The directions of transport of the clastic portions of Units 1, 2, 4 and 7 appear to have been from the southeast quadrant and hence represent a continuation of Swift Run sedimentation (Figures 2, 3, 4). The directions of transport were determined primarily on the basis of grain size, distribution, thickness and crossbedding (Figures 2, 3, 4). In general, though, clastic fragments do not appear as large as those present in the Swift Run.

Crossbedding, distribution and thickness also indicate a secondary source of clastic detritus in Unit 1

(Figure 2) and possibly Unit 2 (Figure 3). The direction of transport in this case appears to have been from the northwest quadrant (Figures 2,3). This secondary source corresponds with that suggested by Werner (1966, p. 19). These sediments are found west of the northern end of Stony Creek fault (G-9 to F-12). Stony Creek fault appears to have been active at least through deposition of Unit 2. Swift Run metasaprolite was developed on the flanks of adjacent paleohills (Figure 1) concurrent with fault activity. These paleohills with up to 300 m relief were overlapped by Swift Run sediments equivalent in age to the lower part of the Catoctin Formation (Figures 1, 2, 3, 4).

Pyroclastic activity occurred throughout Catoctin Formation time as evidenced by pyroclastics in Units 1, 2, 3 and 5 (Figures 2, 3, 4). Based on distribution and thickness, the direction of pyroclastic transport was from the northwest quadrant for all units (Figures 2, 3, 4).

Lava flow activity from fissure-type eruptions is inferred. Some of the vents cut through the Pedlar Formation as indicated by the unakite cobbles picked up by the lava flows of the lower Catoctin Formation (J-8).

(4) From late Catoctin Formation time through

Loudoun Formation time, the dominant direction of sediment transport appears to have been from the northeast parallel to the Appalachian structural trend (Figure 5). This change in direction of transport coincides with a change in composition of the sediments to arkosic detritus. Likewise, the source terrain for this detritus appears to have been nearby localized areas west of the Catoctin basin (Figure 5).

Flow activity continued throughout the Loudoun Formation period and differed from the previous activity only in that the flows appear thinner, probably due to separation by clastic detritus, whereas the Catoctin Formation flows are stacked one on top of another. Loudoun Formation time was ended by pyroclastic activity in this area.

(5) Paleozoic metamorphism appears to have partially altered the lava flows (Type Two) toward the following greenschist assemblages:

a) quartz, low albite, epidote, penninite and magnetite

b) quartz, low albite, penninite, magnetite and calcite

c) quartz, low albite, epidote and actinolite.

Type Two greenstones have relict oligoclase-andesine (An_{28}-An_{33}) laths, and in the upper flows (zones C and D), relict pyroxenes are present. Type One greenstones are represented by the assemblage:

d) epidote, quartz.

Primary hematite also may characterize this assemblage.

In the greenstones, epidote and quartz apear to have formed at the expense of plagioclase, followed by formation of penninite from epidote, plagioclase and pyroxene, followed by formation of actinolite from epidote, plagioclase, pyroxene and penninite. In the Type Two greenstone, actinolite content increases with depth and is associated with a secondary crinkle foliation. The pyroclastic phyllites are characterized by the metamorphic assemblage of:

e) muscovite ($2M_1$) and/or penninite, quartz, magnetite and/or hematite ± epidote.

No relict minerals were noted in the pyroclastics.

The Catoctin Group metasediments are characterized by the formation of sericite, microcrystalline quartz, epidote, penninite and actinolite. The feldspars were altered to sericite, microcrystalline quartz, epidote and/or penninite, whereas the actinolite was formed at a later stage, principally at the expense of epidote and penninite.

(6) The question of the original nature of the Catoctin Group greenstones has been discussed by Bloomer

and Bloomer (1947, p. 99-100) who consider the lavas as "...an altered plateau andesite...". Bloomer (1950, p. 773) who states "...the Catoctin resembles spilite", and Bloomer and Werner (1955, p. 592) who consider the Catoctin as spilite of original basaltic or andesitic omposition. Reed (1955, p. 893; 1964, p. C72) considers the Catoctin Group volcanics as tholeiitic plateau basalts (rather than basalt-andesite) chemically altered to spilite by low-grade regional metamorphism. Reed's conclusions are based primarily on the argument that the lavas were extruded subaerially (Reed, 1955, p. 894; 1964, p. C72) and that the lavas have no relict olivine and do have relict pigeonite and relict basaltic texutre (Reed, 1955, p. 893). It is interesting to note that although Reed was able to identify the relict texture he was unable to identify any relict plagioclase (Reed, 1955, p. 893). Reed's evidence (1955, p. 894) for subaerial extrusion is:"...presence of columnar jointing, absence of pillow structure, aereal extent of individual flows, and relatively thin breccia zones between flows...". The writer does not consider any of the above as conclusive evidence of subaerial extrusion. In contrast, the arguments for submarine eruption of the Catoctin volcanics are (a) that in the area mapped by Bloomer and Werner (1955, p. 590) and in the study area, interlayered sediments lack any greenstone detritus, and (b) that within the study area, no recognized soils are developed on the tops of flows.

The original composition of the volcanics appears to be basaltic-andesite as indicated by the mineralogy of the relict grains present. Bloomer and Bloomer (1947, p. 98) and Bloomer (1950, p. 772-773) report relict andesine ("...50 percent of the primary mineral composition..."), biotite and pyroxene. Bloomer and Werner (1955, p. 592) report relict andesine (An_{40}), pyroxene and biotite (?). Reed (1955, p. 888) reports only relict pyroxene. Within the study area, the Type Two greenstones contain abundant relict oligoclase-andesine (An_{28}-An_{33}) and a few relict pyroxenes. Thus the writer interprets the greenstones as having originated from submarine andesite-basalt lavas which were altered by regional metamorphism to greenstones.

Chilhowee Group

Regional Distribution and Nomenclature

The Chilhowee Group was named originally by Safford (1856, p. 149, 152-153), and its type locality is Chilhowee Mountain, Sevier and Blount counties, Tennessee. The Chilhowee Group and its southwestern and northwestern equivalents have been mapped in a belt along the western flank of the Blue Ridge (or its southwestern equivalent) from Alabama to southern Pennsylvania. In Pennsylvania, they wrap around the nose of the Blue Ridge-Catoctin Mountain anticlinorium and can then be traced southward back through Maryland into Virginia. This thick clastic sequence, which is overlain by Lower Paleozoic carbonates, rests unconformably on the older Precambrian crystalline basement of the Blue Ridge in a few places such as near Roanoke, Virginia. Typically, however, it overlies the younger Precambrian Ocoee, Mount Rogers or Catoctin sequences.

In the type area in northeastern Tennessee, the Chilhowee Group consists, in ascending order, of the Unicoi, Hampton and Erwin formations. The term "Chilhowee Group" has been extended to include rocks in southwestern Virginia where the type formations have been mapped as far north as Roanoke, Virginia (Woodward, 1932, p. 28-30; Butts, 1933, p. 2-3, 1940, p. 26-40). Butts (1933, p. 2-3, 1940, p. 26-40) extends the term "Chilhowee Group" to include a sequence of rocks found in the Blue Ridge of northern Virginia, Maryland and southern Pennsylvania, which are lithologically similar to the type formation. This sequence, originally mapped as Formation I by W.B. Rogers (1884, p. 167-168, <u>his</u> Plate 1) is the Loudoun, Weverton, Harpers and Antietam formations. The writer believes that the Loudoun should be included in the Catoctin Group rather than in the Chilhowee Group because, in addition to the reasons previously mentioned, no fossils have been found in the Loudoun or Catoctin metasediments, whereas Scolithus has been reported in the Weverton, Harpers and lower Antietam along with Scolithus and diagnostic Early Cambrian trilobites and brachiopods from the upper Antietam. Hence, in this report, the Cambrian-Precambrian boundary is placed between the Weverton of the Chilhowee Group and the Loudoun Formation of the underlying Catoctin Group. This boundary was chosen by Reed (1955, his Plate 1), who originally defined the Catoctin Group.

The names Weverton, Harpers and Antietam are used in this report because of the similarity of the mapped units with those mapped by King (1950a) in the Luray area. Based on the present study as well as that of Werner (1966, Plate 1, p. 19-25), however, it appears that a major thrust fault zone (Back Creek-North Fork-South Fork faults) in these two areas separates the more easterly Weverton and Harpers successions from the more westerly Unicoi and Hampton successions.

Weverton Formation

Name and age: The type section of the Weverton Formation is in the Potomac River gorge near Weverton, Washington County, Maryland. Keith first described (1892, p. 365) and later named (1893, p. 329-333) the formation. The southwestern equivalent, the Unicoi Formation, was named by Campbell (1899, p. 3). Members of the Weverton have been differentiated from Pennsylvania to central Virginia; however, none have been formally named. The Weverton overlies the Loudoun Formation and conformably underlies the Harpers Formation. Within the study area, the lower contact appears conformable.

Fossils: No fossils were found within the area; however, *Scolithus* tubes have been reported in the Weverton or its equivalent by two writers: Stose and Jonas (1939, p. 38) and King (1950a, p. 19). The latter states that *Scolithus* is present in the Elkton area in two places: (1) west slope of Devils Tanyard; and (2) southeast slope of Grindstone Mountain.

Distribution and thickness: Exposures of the Weverton Formation are found in three belts within the map area: (1) the normal northwest limb of the Sherando syncline (D-4 to E,G-1); (2) the overturned southeast limb of the Sherando syncline (H,I-1 to D-5,6); and (3) the southeast region of the Back Creek-North Fork faults

(A-10,11 to D-7). The Weverton in these three belts has estimated thicknesses of 1800 m, 900 m and 600 m (Plate 3) respectively.

Lithology: The Sherando syncline belts of Weverton are divisible into three members which are essentially the same as those described by King (1950a, p. 18-19) in the Elkton area.

The lowest member is 240 to 480 m thick (Plate 3) on the southeast limb and probably exceeds 600 m in thickness on the northwest limb of the Sherando syncline. This member consists principally of quartzite and quartzose sandstones with a few interlayered quartz siltstones and quartz and lithic pebble conglomerates. Sandstone and quartzite beds have a maximum thickness of about 2 m and average about 15 to 30 cm. Foreset bedding occurs (H-2,3) in these beds.

The quartz grains are typically about 2 to 3 mm in diameter. The pebbles in the conglomerates are 5 to 20 mm in diameter and are typically subrounded grains of quartz, although some subrounded rock fragments up to 25 mm in length occur in the conglomeratic sandstones. The conglomeratic sandstones are poorly sorted, and beds are typically less than 150 cm thick. The contact between the basal part of the Weverton Formation and the underlying Loudoun Formation usually is concealed but can be located easily within a few meters because of the sharp contrast in lithologies. This lower contact does not appear to be gradational anywhere within the map area. It can be observed in A-11; E,F-6; and H,I-1,2. The middle member of the Weverton Formation consists of about 60 to 120 m of shale-phyllite of the overlying Harpers Formation. Both the upper and lower contacts are gradational, consisting of siltstone between the mudstone and the overlying or underlying sandstones. The middle member was not mapped in the southwestern belt near the North Fork fault.

The upper member of the Weverton is 360 to 510 m thick. The upper member consists of dark greenish brown siltstones and interbedded ferruginous sandstones and quartzites. The siltstones typically are fine grained with grains less than 2 mm in diameter. Larger (3 to 5 mm) floating grains of quartz, however, are abundant within the siltstones. The sandstones and quartzites are coarse grained (2 to 3 mm diameters) and cemented with hematite. The upper contact with the Harpers Formation is vertically gradational and is discussed with that formation.

Six thin sections, which were made from one sample of conglomerate, two samples of siltstones, and three samples of sandstone, were examined. The conglomerate consists mainly of subangular, very poorly sorted grains and fragments of quartz, quartzite (pressure quartz) and quartz sericite chlorite schist. These clastic grains constitute 70 percent of the rock. Thirty percent of the matrix is hematite, at least some of which is secondary because it spreads out from veins. The remainder of the matrix consists of 20 percent sericite (fine-grained) and 50 percent epidote (coarse-grained aggregates). The sericite and epidote matrix exhibits a strong preferred orientation and resembles the Harpers shale-phyllite in appearance and composition.

Subrounded magnetite grains about 2 mm in diameter constitute about 5 percent of the rock. Both euhedral and well rounded zircon crystals are present as minor constituents. Fine-grained opaques and a blue copper are present in minor amounts, but no feldspars were recognized.

The siltstones consist of about 40 percent matrix similar to the Harpers shale-phyllite in composition. Forty percent of the rock consists of fine-grained, well sorted, subangular quartz grains, and 10 percent consists of larger floating quartz grains with a diameter of about 5 mm. The remaining 10 percent is made up of segregated detrital layers of heavy minerals, principally ilmenite, magnetite, hematite, fine-grained opaques, apatite, zircon and tourmaline. The tourmalines are generally subangular, but the zircons are both euhedral and subrounded. The matrix exhibits little preferred orientation, and no feldspars were observed.

The three remaining thin sections revealed that the sandstones consist of 50 to 60 percent poorly sorted, subangular to subrounded, elongated quartz grains and a few rock fragments (principally quartzites), but no feldspars. The matrix consists almost exclusively of hematite and magnetite which have replaced the original matrix, which resembles that in other parts of the Weverton and Harpers. Only a few subrounded zircons, unaltered by hematite, are present as remnants of the original matrix. Hematite-magnetite veins and stringers cut through the clastic grains in the rock along with non-undulatory quartz veins about 2 mm wide. No preferred orientations were noted except for the vein quartz. The quartz veins, like those in the Harpers, are parallel to the foliation.

Harpers Formation

Name and age: The type section of the Harpers Formation is located in the Potomac and Shenandoah River gorges near Harpers Ferry, Jefferson County, West Virginia. This section, first described by Keith (1892, p. 365) was named later the Harpers Ferry shale by him (*in* Williams and Clark, 1893, p. 68). The name subsequently was shortened to Harpers shale (Keith, 1893, p. 333-335). Within the Harpers Formation, the Montalto quartzite member (Stose, 1906, p. 207) has been differentiated in Pennsylvania, whereas the Snowden sandstone member of the Hampton Formation (Bloomer and Werner, 1955, p. 595-597) has been mapped in parts of central Virginia. The southwestern equivalent of the Harpers is the Hampton Formation, first named by Campbell (1899, p. 3).

The Harpers Formation conformably overlies the Weverton Formation and also conformably underlies the Antietam Formation.

Fossils: *Scolithus linearis* have been found in both the Montalto and Snowden members of the Harpers and Hampton formations respectively. The *Scolithus*-bearing quartzite member of the Harpers does not appear to be present within the map area.

Distribution and thickness: Within the study area, the Harpers Formation crops out in two belts (H-1 to D-6; B-9 to B-10). The northern belt is in the central part of the Sherando syncline, and the southern belt lies along the axis of another smaller syncline located between the North Fork and South Fork faults.

Werner (1966, p. 23) and Bloomer and Werner (1955,

p. 596) estimate the thickness of the Hampton to be 90 to 300 m in the quadrangle adjacent to the west. They describe the contact with the underlying Unicoi as gradational and place the contact at the top of the stratigraphically highest pebble conglomerate bed in the Unicoi. Werner (1966, his Plate 1) shows the Hampton to be 1700 m thick. Both estimates, however, are for rocks located northwest of the Back Creek-North Fork fault, a major thrust fault. Inasmuch as the Harpers appears to be part of a more southeasterly sequence, the thickness of the northwesterly Hampton succession may not be valid for the Harpers.

The upper contact is not exposed within the map area, but about 400 m of Harpers is present. The lower contact is exposed at three locations (A-10; E-5; F-3). The contact between the Harpers and the underlying Weverton in the Sherando syncline (F-3 and E-5) is gradational and consists of about 6 m of siltstone. The contact exposed to the southwest (A-10) is sharp, with shale lying directly on the greenish sandstone.

Lithology: The Harpers Formation in the Sherando syncline consists principally of interbedded claystone and siltstone which have been mildly metamorphosed to phyllites with a well developed northeast-trending cleavage. The Harpers is very similar to that described by King (1950a, p. 19-20) in the Elkton, Virginia area, where it consists mainly of siltstone with shale near the base. Bedding is 3 to 15 mm thick, although a few thin siltstone beds up to 15 cm thick are present locally. Fresh samples are bluish-gray, whereas weathered specimens are usually greenish gray or tan. Quartz veins 1 to 2 mm are present and are parallel to the foliation.

A microscopic examination of one thin section showed that the Harpers phyllite is a very fine grained laminated rock in which cut-and-fill structures are visible. The laminae are about 2 mm thick and consist of alternating bands of two mixtures: (1) coarser grained layers of 70 percent quartz and 30 percent fine-grained minerals --principally sericite of <u>in situ</u> metamorphic origin; and (2) finer grained layers of 50 percent quartz and 50 percent fine-grained minerals (30 percent sericite). The fine-grained unidentified matrix, which may be iron-stained sericite, appears brown under white light and reddish-brown with crossed nicols. An x-ray pattern of sample 301 indicates a composition of muscovite ($2M_1$), quartz and penninite. The matrix material and the sericite exhibit an optically preferred orientation parallel to the foliation. The quartz grains do not appear to have a well developed orientation and are mostly subrounded. The vein quartz, however, does show preferred orientation with extinction parallel to the cross hairs. The vein quartz also exhibits only slight undulatory extinction.

Antietam Formation

Name and age: The type locality of the Antietam quartzite was first reported by Keith (1892, p. 365), and later was designated by him (Keith, 1894, p. 3). The name Antietam was used first by Keith (*in* Williams and Clark, 1893, p. 63) for these rocks located in Washington County, Maryland on a tributary of Antietam Creek. The Erwin Formation named by Keith (1903, p. 5) is the southwestern equivalent of the Antietam Formation.

Fossils: The oldest fossils (Early Cambrian) in the Appalachians have been found in the upper part of the Antietam Formation.

The fossils occur near the transitional beds at the top of the Antietam. The fossiliferous rocks usually have been described as thin bedded, friable, shaly and iron-stained. Although no fossils were found by the writer, rocks very similar to the fossil-bearing strata do occur at the western edge of the area mapped (A-6).

Scolithus tubes occur abundantly in some layers of the Antietam, both within the map area and on a regional basis. Scolithus is a straight tube perpendicular to bedding. Within the map area, they are 15 to 60 cm in length and have cross sections with 3 to 5 mm diameters. Schwab (1970, p. 356) reports tubes up to 90 cm in length in central Virginia. The Antietam conformably overlies the Harpers Formation and conformably underlies the Tomstown Dolomite.

Distribution and thickness: Within the map area, outcrops of the Antietam Formation are limited to the northwest corner (A-4 to A-9 to D-7). Neither the lower nor the upper contact was observed; hence, the thickness could not be measured. Werner (1966, p. 25) estimates the thickness to be about 200 m in the quadrangle adjacent to the western boundary of the area. However, Werner (1966, *his* Plate 1) shows the thickness as about 580 m.

Lithology: The Antietam within the area is typically a light-gray, tan pink or rust-colored, weathered quartzite or quartzose sandstone. On a fresh surface, the rock is light-gray. Bedding is 30 to 150 cm thick, and within the beds the rocks appear massive. Grain size is less than 2 mm in diameter. Variations from the typical quartzite include conglomeratic layers on top of Torry Ridge (A-7) with quartz pebble ellipsoids with 5 by 5 by 15 mm axes. Interbedded mudstone and siltstone are present (A-6) in layers up to 15 cm thick.

The following description is based on the study of two thin sections of typical quartzite. The rock is about 85 to 90 percent clastic grains and about 10 to 15 percent cement. The cement is principally quartz (90 to 95 percent) with 5 to 10 percent hematite concentrated mainly in and near <u>Scolithus</u> tubes. The clastic fragments are about 95 percent quartz and quartzite fragments, and are subrounded to subangular. Most of the quartz has undulatory extinction, and many inclusions. The grains are welded together with quartz overgrowths or exhibit sutured contacts. The remaining 5 percent of the clastic grains are principally feldspar (about 3 percent) and include plagioclase, microcline and microperthite. Most of the feldspars are smaller than the quartz grains and are at least partially altered. Heavy minerals comprise the bulk of the remaining 2 percent, but both schist fragments and muscovite are found. The principal heavy minerals are several varieties of tourmaline and zircon, including some with overgrowths. Rutile (?) may be present. The heavy minerals are smaller and well rounded.

<u>Origin and metamorphism of the Chilhowee Group</u>

The lower Chilhowee sediments, east of the Back Creek thrust fault zone are metamorphosed to the greenschist facies. The coarser-grained conglomerates, sandstones and siltstones of the Weverton exhibit the development of metamorphic epidote and sericite, whereas the finer-grained mudstones of the Harpers contain metamorphic sericite and penninite.

Paleocurrent analyses have been done recently by Schwab (1970), Brown (1966, 1970) and Whisonant (1970). Schwab (1970) studied the Antietam Formation in central Virginia, whereas Brown (1966, 1970) studied the Chilhowee Group between central Virginia and mideastern Tennessee. Whisonant (1970) examined the Chilhowee Group in Tennessee. Prior to these articles, Whitaker (1955b), did a paleocurrent analysis of the Weverton in northern Virginia and Maryland. A total of more than 1100 paleocurrent measurements was reported in these five articles. The principal direction indicator was crossbedding. All of these paleocurrent analyses indicate that the principal directions of transport were toward the east quadrant.

Evidence which indicates a western or northwestern source of the Chilhowee Group detritus in the northern Blue Ridge is listed below:

(1) the westward and northward facies change from the Harpers shale to the Montalto sandstone (Keith, 1893, p. 335; Stose, 1906, p. 206-207; 1932, p. 45; Stose and

Bascom, 1929, p. 7; Stose and Jonas, 1939, p. 43; Freedman, 1967, p. 25-26; Fauth, 1968, p. 33-34).

(2) the gradual thinning of the Weverton toward the south and its pinching out or possible facies change to Harpers (Whitaker, 1955a, p. 443-459).

(3) Within the study area and the adjacent region mapped by Werner (1966), a distinct change in grain size occurs. Southeast of the Back Creek-North Fork fault zone, the Harpers is a siltstone-shale phyllite, and the Weverton consists mainly of quartzose sandstone and siltstone. Northwest of the fault zone, Werner (1966, p. 23-24) reports that the Hampton consists of siltstone to sandstone and that the Unicoi consists mainly of conglomerate to pebbly quartzite. Reconnaissance by the writer into the area mapped by Werner (1966) established that the Unicoi-Hampton succession was distinctly coarser and more heterogeneous than its easterly equivalent, the Weverton-Harpers succession.

(4) The Catoctin Group apparently covered an area to the east of the main Chilhowee sedimentation basin 16 to 48 km wide at the time of Chilhowee sedimentation. However, the Catoctin volcanic detritus was not observed in the Chilhowee sediments in the study area. King (1949, p. 518, 528), Stose and Stose (1946, p. 31-34) and Werner (1966, p. 21) also noted the absence of Catoctin detritus

in the Chilhowee. The lack of Catoctin detritus indicates that either (1) the entire Catoctin basin was bypassed if the source of the Chilhowee detritus was to the east, or (2) the source of the Chilhowee detritus was located west or south of the Catoctin basin.

Tomstown Dolomite and Waynesboro Formation

Tomstown Dolomite

The Tomstown Dolomite was named by Stose (1906, p. 206) for exposures at Tomstown, Franklin County, Pennsylvania. The Tomstown has been mapped from southern Pennsylvania to central Virginia. Its southwestward equivalent is the Shady Dolomite (Butts, 1933, p. 3), named for Shady, Johnson County, Tennessee by Keith (1903). The Tomstown apparently conformably overlies the Antietam Formation; however, the contact is seldom observed. The Tomstown conformably underlies the Waynesboro Formation (Butts, 1933, p. 2-4; Woodward, 1949, p. 124-139; King, 1950a, p. 24-30). An Early Cambrian age was assigned by Butts (1933, p. 4). The Tomstown is described by Butts (1933, p. 3) as mainly bluish-gray to white coarse-grained dolomite, and thin-bedded limestones also are present (Butts, 1933, p. 3).

Within the area mapped, no exposure of Tomstown was seen; however, previous workers (Stose and others, 1919; Knechtel, 1943) indicates its presence in the subsurface (A-6; D-2). Stose and others (1919, p. 101) state that a dolomite (Tomstown) was found in the bottom of a shaft of unknown depth in Orebank Hollow (A-6). They describe the rock as being dense, light-gray dolomite and containing small pyrite crystals.

Knechtel (1943, p. 184), presumably in the same shaft, reports the shaft's depth to be less than 5 m. He also states that dolomite was found in 3 out of 10 exploratory holes drilled in the same general area during 1941-1942. He identifies this dolomite, which was encountered at depths greater than 15 m, as the Tomstown (p. 184, Plates 29, 30). Knechtel (1943, p. 183) also reports that one of the 12 drill holes located near the abandoned Lyndhurst Mine (D-2), bottomed in fresh domomite at 44 m.

Waynesboro Formation

Stose (1906, p. 209) named the Waynesboro Formation for Waynesboro, Franklin County, Pennsylvania, where its type expsoure is on a ridge just north of town. The Waynesboro's equivalent from central Virginia southward is the Rome Formation (Butts, 1933, p. 4), named for exposures near Rome, Floyd County, Georgia, by Smith (1890, p. 140). The Waynesboro, according to

Butts (1933, p. 4-6), overlies the Tomstown Dolomite and underlies the Elbrook Dolomite and is Early-Middle Cambrian in age. Butts (1933, p. 5) described the Waynesboro as mainly red and green shale.

Knechtel (1943) reported the only probable occurrence of the Waynesboro within the study area. Knechtel states (1943, p. 183) that near the old Lyndhurst Mine (D-2), 11 out of 12 exploratory holes drilled during 1942 bottomed in shale at depths of 15 to 50 m. Knechtel (1943, <u>his</u> Plate 29) identifies this shale as Waynesboro.

<u>Triassic (?) Intrusions</u>

These intrusions post-date the Paleozoic deformation and metamorphism. They are not only structurally discordant with the older trends, but they also exhibit no effects which could be attributed to the regional metamorphism that affected the Catoctin greenstones, themselves of mafic bulk composition. Recent authors (Bloomer and Werner, 1955, p. 599; Werner, 1966, p. 36; and Allen, 1963, p. 48-51) consider these basic intrusions in the Blue Ridge as being Triassic (?) or younger age.

The Wintergreen dike (I-14) is the best exposed of these long narrow dikes. Along State Route 627, it appears to be 1 to 2 m thick. The dike probably is vertical as suggested by its linear trend across rolling topography. Its lateral extent is inferred from the stream drainage. The rock, which is black on a fresh surface, weathers to orange-brown blocks. These blocks have all the corners and edges rounded.

In thin section, the Wintergreen dike has a subophitic texture and is composed principally of 20 to 25 percent euhedral olivine phenocrysts, 15 to 20 percent pigeonite (clinopyroxene, optically positive, small 2v, extinction angle 24°-32°), 5 percent magnetite and 45 to 50 percent plagioclase. The latter exhibits zoning from An_{64} (labradorite) to An_{33} (andesine). A combined Carlsbad-albite twinned crystal indicates a composition of An_{51} (labradorite). Minor amounts, estimated as less than 2 percent, of hematite, chlorite (?), sphene (?), and biotite (?) are present.

Three other diabase dikes were mapped (M-6, N-4, N-5). The dike along the road cut of State Route 709 (M-6) is exposed. The presence of the other two dikes is inferred from float.

Unconsolidated Deposits

The principal unconsolidated deposits, which are delineated on Plate 2, are coarse stream gravels, boulder trains and alluvial stream and flood-plain deposits.

The coarser material and the boulder trains are confined to the higher elevations, in or near the more mountainous regions. These deposits extend up ravines to the crest of the ridges.

Rockfish Valley is filled with both colluvial and alluvial fill which, except for local concentration, probably does not exceed a few meters. This thickness is indicated by the numerous occurrences of bedrock exposed in many of the stream channels on Plate 2.

Sherardo Valley apparently is completely covered with a coarse, colluvial-alluvial fill. South of the Back Creek fault, this fill is only a few meters thick, as indicated by bedrock channels in most of the streams. North of the Back Creek fault, the deposits probably are much thicker. Mining and exploratory drilling (A-6) D-2) reported by Knechtel (1943, p. 167, 183-187) indicate that fresh bedrock can be found at depths ranging from 5 to 50 m.

Ore Deposits

The two major areas in which iron and manganese deposits have been located and mined are the abandoned Lyndhurst Mine (D-2) and the abandoned Mount Torry workings (A-6). In addition to these two areas, the writer found a number of test pits on the southwest end of the hill (E-3,4) located about 0.8 km due east of Sherando.

The Lyndhurst operation was confined to a 150 m square area, whereas the Mount Torry workings were in an area about 1.6 km long and 0.8 km wide. The latter extends westward slightly beyond the mapped area. The Lyndhurst Mine operations are reviewed by Stose and others (1919, p. 99-100) and Knechtel (1943, p. 167, 181-184).

The Mount Torry operations were described by Stose and others (1919, p. 100-102) and Knechtel (1943, p. 184-187), who describes the final manganese mining and exploratory work during the late 1930's and early 1940's.

The testing done in the area east of Sherando consists of a series of shallow pits and trenches dug into the residuum developed on the Weverton quartzites and sandstones. The date of testing is unknown. Both iron and manganese oxides may be found on the slag piles.

STRUCTURAL GEOLOGY

Precambrian Structures

Within the Marshall and Pedlar rocks of the Virginia Blue Ridge Complex in the eastern and southern parts of the area are east-trending antiforms, synforms and faults which predate Late Precambrian Catoctin volcanism and sedimentation. The pre-Catoctin age of the east-trending folds is based on the fact that no east-trending folds are present in either the Catoctin Group or the Chilhowee Group. The antiforms and synforms are interpreted as anticlines and synclines, respectively, because of the large size of some of the structures. One fault which appears to have been active during Late Precambrian Catoctin volcanism is present. The fault, the Stony Creek fault, is younger than the earlier Precambrian structures.

Earlier Precambrian Folds

The largest of these structures is the Bryant Mountain anticlinorium with a Pedlar Granite core (C,D,E-12,13; Plate 3 - structure sections FF', GG') which is at least 6 km long and 4 km wide. The Bryant Mountain anticlinorium appears to plunge eastward and has at least three secondary folds superimposed on it. The three secondary folds are two anticlines (G,H-13,14; I-12) and one syncline (G,H-13) which are approximately 1.5 by 0.8 km, 1 by 0.2 km and 0.5 by 0.6 km in length and width respectively.

Another large anticlinorium probably is present beneath the Humpback Mountain (I-7)-Crawford Knob (I-9) area (Plate 3 - structure sections CC', GG'). Hence, the plunging folds south of the Johnson Mountain anticline (K-3 to N-5) are probably all secondary folds on this plunging nose. The secondary folds consist of two anticlines (L,M,N-8; L,M-6) and two synclines (L,M,N-6,7; K,L,M-5) which are approximately 2 by 1.7 km, 2 by 1 km, 2 by 0.8 km and 2 by 0.6 km in length and width respectively. Structure section A-A' (Plate 3) shows that these folds are all broad with steeply dipping flanks. The northernmost folds (L,M-6; K,L,M-5) have curved axial traces. Structure section G-G' (Plate 3) suggests that the folds plunge steeply. A small (0.8 by 0.5 km) western plunging (?) portion of an anticline is present along Pauls Creek (K-9) where erosion has cut through the Catoctin rocks. This anticline is probably the western end of one of the anticlinal folds (L,M,N-8) described above. The exposed portion of the smaller anticline (K-9) is 0.8 by 0.5 km and has a curved axial trace.

The Johnson Mountain anticline (K,L,M,N-3,4,5) is another large fold (4 by 3 km) with five secondary anticlines (L,M-5; M,N-5,6) and synclines (L,M-5; M,N-5; L,M-3) superimposed on the flanks of the plunging nose of the larger anticline. The size of the secondary folds are 1.7 by 0.3 km; 0.7 by 0.3 km; 1.5 by 0.3 km; 0.6 by 0.3 km; and 1.1 by 0.6 km respectively. One of the secondary anticlines (L,M-5), as well as the Johnson Mountain anticline, has a curved axial trace.

North of the Johnson Mountain anticline are two anticlines (L,M,N-2; L,M-1) and an intervening syncline (L,M-2) whose dimensions are 2.3 by 0.5 km; 1.8 by 0.5 km; and 1.5 by 0.5 km respectively. These folds appear to be slight flexures outlined by foliation trends rather than well developed folds.

Earlier Precambrian Faults

Associated with the east-trending folds and bounding the Johnson Mountain anticline are what are interpreted as two east-southeast-trending high angle reverse faults (M-3; M-6). Both fault traces are curved and are about 1.5 km long. The displacement on both inferred faults is unknown. Their effect, however, was to uplift the Johnson Mountain anticline relative to the folds both north and south of the faults.

Similar east-trending pre-Catoctin structures may be present in the subsurface both east and west of the map area as indicated by an east-west-trending seismic zone (Bollinger, 1969, p. 2103). Current activity along this zone may represent rejuvenation of faulting along the Precambrian east-trending structures.

Earlier Precambrian Foliation

The foliation of the Pedlar and Marshall rocks typically appears to parallel the lithologic boundaries, thus defining the folds. By contrast, cleavage formed during Paleozoic time does not parallel lithologic boundaries and, hence, does not outline Paleozoic folds.

Late Precambrian Stony Creek Fault

The Stony Creek fault is about 5 km long and is located in the Stony Creek Gorge (G-9 to J-13). It is interpreted as a steep southwestward-dipping normal fault (Plate 3 - structure section G-G'). Stony Creek fault terminates abruptly to the northwest where it disappears beneath Catoctin rocks (G-8,9).

The Stony Creek fault and later northwest-trending joints probably are responsible for the striking parallelism of the trends of Stony Creek Gorge (G-9 to J-13) and Laurel Springs Branch water gap (E-5 to F-7) on the east and west sides of the Blue Ridge respectively. However, even though Stony Creek fault does not appear to cut Catoctin Group rocks in the vicinity of Laurel Springs

Gap (F-7), it probably is present in the subsurface. Hence, the fault may have had an indirect effect on the formation of the Laurel Springs Branch water gap because of post joint-formation movement along the fault.

Stony Creek fault does not appear to cut across the Lovingston Gneiss present in the southeastern corner of the area (J,K-14,15). Thus the fault probably dies out beneath the alluvial cover of Rockfish Valley. The possibility exists, however, that the Lovingston Gneiss is younger than the Stony Creek fault.

Faulting along the Stony Creek fault probably has occurred at two widely separated times. The first period of activity appears to have been during early Catoctin time. This early activity is suggested by the apparent pinch out of about 300 m of the lower Catoctin Formation in the vicinity of the fault(G-8,9). Hence, the Stony Creek fault is believed to have experienced about 300 m of displacement and concurrently to have caused the formation of a basin on its southwestern side in early Catoctin time. The second period of faulting is discussed later with the Triassic (?) high angle faulting.

Paleozoic Structures

Structures formed during Paleozoic metamorphism and deformation include: northeast-trending and north-trending major and minor folds; a northeast-striking thrust fault zone; a northwest-striking tear fault; and northeast-striking, southeast-dipping cleavage. In addition, many smaller features such as slickensides, ptygmatic features, kink bends and other small scale folds are believed to have originated during Paleozoic deformation.

Folds

The larger folds typically trend northeast, parallel to the median axis of the Blue Ridge anticlinorium (plate 1). The median axis of the Blue Ridge anticlinorium lies about 10-15 km east of the map area. For descriptive purposes, the folds are divided into the following groups: (1) northeast-trending Torry Ridge anticline and other parallel folds northwest of the North Fork-Back Creek Fault zone (A-4 to A-9 to D-7); (2) the Sherando syncline (D-6 to H-1); (3) the Devils Knob syncline (A-12 to F-7); (4) the broad, breached anticlinorium beneath the Humpback Mountain-Crawford Knob area (G-6 to I-12 to J-5); (5) the Pauls Creek syncline (L-10 to K-7); (6) the Spruce Creek syncline (F-14 to H-12); (7) smaller folds in the Catoctin and Chilhowee rocks; and (8) small folds superimposed on the earlier Precambrian folds.

The folds northwest of the North Fork-Back Creek fault zone (A-10 to E-1) consist of the northeastward plunging Torry Mountain anticline (A-8 to C-6), four smaller northeastward plunging folds north of Torry Mountain anticline (A-4 to A-7), and secondary folds superimposed on the Torry Mountain anticline (A-9). The axis of the Torry Mountain anticline is approximately 4 km long and trends N60° E. The other fold axes are subparallel to the Torry Mountain axis and are 1 to 3 km in length. All of these folds are located on the northeastward plunging nose of a larger anticlinorium mapped by Werner (1966, his Plate 1). The folds are interpreted as open structures with gently dipping flanks (Plate 3 - structure sections E-E', F-F'). The wavelength between the larger anticlinal axes appears to be about 1 to 1.5 km. The larger folds are topographically expressed by anticlinal ridges of Antietam quartzite and sandstone and intervening synclinal valleys with colluvial-alluvial fill covering Tomstown Dolomite (Plate 3 - structure section E-E').

The Sherando syncline (D-6 to H-1) within the map area is the southern end of a large synclinal structure which topographically appears to extend 5 to 10 km northeast to the vicinity of Waynesboro, Virginia. The southeastern flank of the Sherando syncline is overturned to the northwest and forms the northwestern flank of the

Blue Ridge anticlinorium (Plate 1 and Plate 3 - structure section D-D'). The axial trace of this tightly folded plunging syncline is about 5 km long within the area, but it is terminated on the southwestern end by the Back Creek thrust fault (D-5). The axial trace of the syncline trends approximately N55° E. The syncline is about 5 km wide, and the axial portion is underlain by Harpers Formation and the flank portion by Weverton Formation. Within the Sherando syncline are small scale structural features such as kink bends (NW¼E-5) and small plunging folds (SW¼G-3 and SE¼H-2). The axial traces of the smaller folds are subparallel to the axial trace of the Sherando syncline.

The Devils Knob syncline (A-12 to F-7) is a long (7 km) doubly plunging asymmetric structure in which only Catoctin Group rocks are exposed. The axial trace trends about N30° E and about N50° E along the southern and northern parts of the fold respectively. The northwestern flank of the syncline dips more gently than the southeastern flank (Plate 3 - structure sections E-E', F-F'), and a small anticlinal flexure, terminated by the Laurel Springs-Reed Gap fault zone, is present on the northwest limb. The southeast limb is truncated by the Devils Knob fault (A-13 to F-10; Plate 3 - structure sections E-E', F-F'). The fold widens northward from about 1 km (A-12) to about 3 km (E-8 to F-9).

A large breached anticlinorium exists beneath the Humpback Mountain-Crawford Knob area (G-6 to I-12 to J-5). This structure is a broad open arch with smaller undulatory secondary folds superimposed on its present surface (Plate 3 - structure sections B-B', C-C', G-G'). The axial traces of the smaller folds are 1.5 to 2.5 km long, and the wavelength of these folds is about 1 to 1.5 km. The overall dimension of the arch appears to be on the order of 6 by 6 km. The axes of the eight folds on the northeastern flank (F-8 to H-8 to I-3) all trend N40° to 50° E, whereas the five fold axes on the southeastern flank (I-11 to J-12 to K-10) trend N60° to 80° E. The three secondary folds on the steeply plunging northeast nose of the arch (Plate 3 - structure section G-G') are subparallel to the Pauls Creek syncline, described below, and trend north-northwest.

The Pauls Creek syncline (L-10 to K-7) axis is about 3 km long and trends N35° E for about 0.5 km at the southern end and bends northward to trend N30° W approximately along the northern part. The eastern flank of the structure is overturned to the west (Plate 3 - structure section B-B', G-G'). This tight fold dies out to the northwest (K-7) and appears to be truncated by the Humpback Mountain tear fault zone (L-10) to the south.

The Spruce Creek syncline (F-14 to G-12) which is about 3 km long and 1 km wide has a curved axial trace which trends about N60° E along the southern end then N20° E along the central portion and finally N50° E along the northern portion. The syncline has a peculiarly cylindrical shape, decidedly similar to Valley and Ridge type folds and not similar to isoclinally folded rocks of the Piedmont. This shape, which has been used in construction of the structure sections (Plate 3 - structure sections E-E', F-F'), is evident at the southwestern end of the structure (F-15) where several secondary folds have warped the flank of the syncline. Furthermore, the units, of relatively constant thickness, can be traced sufficiently within the syncline along strike to determine that flowage folding has contributed little to the formation of this structure. Another syncline is present, along with an intervening anticline, southeast (H-13 to I-12) of the Spruce Creek syncline. The axes of these two folds are subparallel to the trend of the northern part of the Spruce Creek synclinal axis.

Three small north-northwest-trending folds are present in the Catoctin rocks between the Devils Knob and Pond Hollow faults (E,F-10,11) and south of the Pond Hollow fault (G-10). The axes of these folds are about 1 km long. Additional small folds not previously described are: (1) a north-trending syncline, which has an axis about 1 km long, located between the South Fork and Reed Gap faults (A-11,12); (2) a northeast-trending syncline (3 km long) located between the North Fork and South Fork faults; and (3) two secondary northwest-trending folds (0.1 km long) superimposed on the previously described fold adjacent to the South Fork fault (A-10,11).

The folds north of a line extending from A-10 to K-7 all appear to have axes which trend northeast, whereas the folds south of this line commonly have east, northeast, north, or northwest trends present along a single fold axis. The northeast-trending folds such as the Sherando and Devils Knob synclines and the Torry Ridge anticline are of Paleozoic age and related to the similar trending structures throughout the Appalachians. The formation of these folds probably corresponds with the initial development of the arch presently known as the Blue Ridge anticlinorium (Plate 1).

The north-trending folds, which are known to involve only rocks as young as Weverton, probably formed during the same period of deformation which formed the northeast-trending folds. Thus the two fold trends may represent a conjugate set. An alternative explanation is that the north-trending folds represent the result of a left-lateral stress system acting on the previously formed northeast-trending folds. This alternative is discussed later in the section dealing with the Humpback Mountain tear fault zone. In either case, a north-trending cleavage, associated with the north-trending secondary folds of the Spruce Creek syncline (F,G-14) suggests that these north-trending folds may represent a separate phase of deformation. This north-trending cleavage may be related to the development of fibrous actinolite parallel to the secondary crinkle foliation noted in 13 thin sections. This crinkle foliation, as noted previously, was superimposed on the earlier formed albite-chlorite cleavage which is observed in all type two greenstones.

At two localities (F-14; K-8) small ptygma-like quartz layers are present. The axial planes of these ptygma-like folds are more or less parallel to the cleavage in each area. However, the quartz layers exhibit very little thickening andthinning characteristic of axial planar flow-cleavage. Formation of these ptygma-like folds probably was coincident with the formation of the microscopic greenstone crinkle foliations described previously.

The Pedlar-Marshall rocks have small northeast, north or northwest-trending folds superimposed on the older east-trending structures. The smaller Paleozoic fold axes are 0.5 to 2.5 km long. The folds are found at four localities (H,I-13; I-14; L,M,N-8; K,L-5,6). At the first locality are three folds whose axes trend about N35° E. The anticline and syncline at the second locality trend N45° E. Axes of the third and fourth localities trend N10° to 70° W respectively. Four complimentary folds are present at both localities. The fold axes at all four locations are subparallel to Paleozoic fold axes in the adjacent Catoctin Group rocks.

Cleavage

The dominant southeast-dipping cleavage probably postdates the principal period of folding inasmuch as the cleavage cuts across all folds and does not appear to be axial-planar in all folds. Non-axial-planar cleavage is evident particularly in the broad anticlinal arch of the Humpback Mountain region. There the folds are not overturned to the northwest which they would have to be if the southeast-dipping cleavage were axial planar (Plate 3 - structure sections B-B', C-C'). Likewise, Griffin (1971, p. 422) states that this is a later-formed southeast-dipping cleavage approximately parallel to the statistical axial plane of the overturned folds of the Chilhowee Group subfabric. Thus the conclusion of Cloos (1957, p. 837-838) that both the cleavage and folds were produced by a single deformation plan does not appear applicable to the Blue Ridge of central Virginia.

Thrust Faults

A single thrust fault zone, consisting of six individual faults, extends across the northwestern corner of the map area (A-10,11,12 to E-1 to G-4). This zone of thrusting, named the Back Creek fault by Stose and others (1919, their Plate 1), may be the Blue Ridge frontal thrust inasmuch as metamorphosed rocks of the Blue Ridge are displaced over unmetamorphosed Paleozoic rocks of the Valley and Ridge. However, inasmuch as the Blue Ridge thrust fault has been neither traced this far north from its known locality (near Roanoke, Virginia) nor demonstrated to connect with any of the faults in this area, the term Back Creek fault will be retained for the portion of the zone along Back Creek (D-7 to E-1). The other branches of the zone have been given the following names: (1) North Fork fault (A-10 to D-7); (2) South Fork fault (A-11 to C-8); (3) Reed Gap fault (A-12 to D-7); and (4) Laurel Springs fault (B-9 to G-4). In addition, one unnamed fault exists between South Fork and Reed Gap faults (B-9 to G-4). The Back Creek thrust zone was traced northeastward by Stose and others (1919, their Plate 1) for a distance of 20 km.

The N60° E striking North Fork fault, the N45° E striking South Fork fault and the N40° E striking Reed Gap fault merge northeastward to form the N10° E striking Back Creek fault.

On the other hand, the Laurel Springs fault branches off the Reed Gap fault (C-9) and strikes N40° E for about 7 km to where it is terminated by the Humpback Mountain tear fault zone (G-4). The thrust fault zone is interpreted as a low-to-moderate-dipping zone (Plate 3 - structure sections B-B', C-C', D-D', E-E', F-F'). Inasmuch as all the faults appear to merge within the map area (B-9 to D-7), it seems likely that they also merge at depth as depicted in structure sections E-E' and F-F' (Plate 3). If the faults merge at depth, then it is clear that the structurally higher faults must dip more steeply than the basal Back Creek-North Fork fault; hence, in structure sections E-E' and F-F' (Plate 3) the structurally higher faults are depicted as branching off subparallel to the basal thrust and then steepening rapidly.

The Back Creek thrust zone truncates three northeast-trending folds (D-5, A-9, B-9) and three north-northwest-trending folds (A-11, A-11, A-10). Thus, thrusting along the Back Creek fault zone definitely is younger than the folding phase of deformation. Likewise, thrusting was either coincident with the development of the southeast-dipping cleavage or postdates cleavage development as indicated by the presence of slickensides on cleavage surfaces adjacent to the thrust zone (A-11, A-12, E-5). If the slickensides are the result of thrusting, then the upper block has moved northwestward (about N40°W) relative to the lower block. Minimum displacement on the thrust fault zone appears to be on the order of 3000 m based on the apparent stratigraphic displacement (Plate 3).

Humpback Mountain Tear Fault Zone

The Humpback Mountain tear fault zone (G-5 to L-10) appears to consist of a series of vertical or nearly vertical *en echelon* faults (Plate 3 - structure section G-G'). The southwestern side of the fault is interpreted as having been uplifted relative to the northeast side, based on the deeper level of erosion apparent on the southwest side. The zone appears to have experienced strike-slip as well as vertical movement as indicated by slickensides (H-6). Minimum displacement on the tear fault zone appears to be on the order of 300 to 600 m.

A discrepancy with the interpreted upward and northward movement of the western block is the north-striking cleavage on the eastern block (H,I-6 to I-8). The bending of the cleavage from northeast-striking (I-4) to north-striking suggests that the fault zone has experienced left lateral movement instead of right lateral movement. An alternative explanation of the cleavage anomaly is that the north-trending cleavage is genetically related to the north-trending folds in this area.

Triassic (?) Structures

High-angle Faults

A series of northeast-trending, high-angle faults 2 to 12 km long are present in the southwestern part of the area. These faults are: the Devils Knob (A-13 to G-9), Pond Hollow (C-12 to G-10), Grassy Ridge (G-12 to H-11), Little Stony Creek (G-12 to I-12) and Wintergreen (H-14 to I-13) faults. The latter three faults are interpreted as southeast-dipping normal faults, whereas the Pond Hollow fault apparently dips northwestward, and the Devils Knob fault probably is vertical (Plate 3 - structure sections E-E', F-F', G-G').

Movement along these faults appears to have included both normal and strike-slip components as indicated by both slickensides and relative displacements (Plate 3 - structure sections E-E', F-F', G-G'). The slickensides related to movements along these faults typically are developed on joint surfaces rather than on cleavage or bedding surfaces (A,B,C,E-12).

As a result of movement along this series of faults, horst and graben structures have been developed. These structures resemble the much larger Triassic block-faulted basins of the Appalachians, hence the age of these faults is considered to be Triassic (?).

The high-angle faults postdate movement along the thrust fault zone inasmuch as the Devils Knob fault cuts the Reed Gap thrust just west of the western boundary of the area (A-12,13). The relationship between the Wintergreen fault and the Triassic (?) Wintergreen diabase dike (I-14; Plate 3 - structure section E-E') is uncertain. However, tha faulting probably precedes the dike intrusions because the dike rocks exhibit very little alteration. If the faults had cut the dikes, then much more alteration would be expected inasmuch as the faults typically have shear zones up to 30 to 60 m wide developed along them (C-12, H-12).

Movement along the Stony Creek fault also occurred during this episode of faulting, because none of the high-angle faults are recognizable east of the Stony Creek fault. This episode of adjustment along the Stony Creek fault had the same relative sense of movement as the earlier faulting (west side moved relatively downward). The Pond Hollow (D-12)-Black Rock Mountain (F-11)-Potato

Patch Mountain (E-11) area appear to have been rotated upward relative to the dropped adjacent western side of Stony Creek fault (G-10,11). The total fault movement appears to have included: (1) a rotational movement about a northwest axis; (2) normal faulting along a northeast trend; and (3) minor strike-slip faulting along this northeast trend.

Joints

As previously stated, slickensides apparently related to the period of high-angle faulting are developed along joint surfaces. Inasmuchas the slickensides are present on all four of the major joint trends, it is likely that joint development preceded or coincided with the high-angle faulting. Likewise, because the joint trends appear to cut across all Paleozoic structures, the joints are interpreted as post thrust faulting. The age of these joints, therefore, is considered to be Triassic (?).

Although joint measurements were not a primary objective of this project, a sufficient number were measured to establish the trends in all rock types in the area. The four principal trends are: (1) northeast strike with a steep southeast dip; (2) northeast strike with a steep northwest dip; and (3) northwest strike with a steep southwest dip; plus (4) north-northwest to north-northeast strike with a dip from moderate northwest to vertical to moderate northeast.

The joint trends in the Catoctin Group typically are well developed and prominent on air photos and are primarily responsible for the blocky stairstep type outcrops of greenstone. The joints in the Chilhowee Group are well developed but not prominent on air photos. Within the more massive quartzites, the joint trends are of value as an aid in determining bedding. Joint trends in the Virginia Blue Ridge Complex are difficult to establish because of weathering effects on the previously formed foliations, cleavages, dikes and small faults.

SUMMARY OF GEOLOGIC HISTORY

The relative sequence of events is given below. Relative ages are indicated for each event.

(1) Deposition of Pedlar and Marshall sediment and subsequent metamorphism to the granulite facies accompanied by alteration of Pedlar Gneiss to Pedlar Granite and/or intrusion of Pedlar Granite into Pedlar Gneiss (earlier Precambrian).

(2) Retrograde metamorphism of Pedlar and Marshall formations to as low as greenschist facies. Folding of Pedlar and Marshall formations about east-west-trending axes (earlier Precambrian).

(3) Movement along Stony Creek fault. Establishment of the northeast-trending Catoctin Group volcanic basin. Eruption locally of basic to intermediate lava flows from northeast-trending dikes; accumulation of felsic pyroclastics derived from a western or northwestern source; deposition in localized basin of Swift Run and Catoctin sediments derived principally from an eastern and southeastern source; deposition of Loudoun and upper Catoctin arkosic sediments derived principally from a western or northwestern source and transported southwestward along the western edge of the Catoctin basin (Late Precambrian).

(4) Deposition of the sediments of the Lower Cambrian Chilhowee Group derived from a western or

northwestern source.

(5) Deposition of an unknown thickness of Cambrian Tomstown Dolomite, Waynesboro Shale and other younger Paleozoic rocks.

(6) Folding about northeast-trending axes, followed by folding about north-trending axes.

(7) Development of greenstone cleavage concurrent with greenschist facies metamorphism during which the following minerals formed:

a) Catoctin Group metasediments--epidote, chlorite, sericite and microcrystalline quartz
b) Catoctin Group greenstones--epidote, chlorite, albite, quartz and actinolite
c) Catoctin Group pyroclastic phyllites--epidote, chlorite, sericite and microcrystalline quartz
d) Chilhowee Group metasediments--epidote, chlorite, sericite and microcrystalline quartz.

(8) Northwestward thrusting of the Blue Ridge over the adjacent Paleozoic rocks of the Valley and Ridge.

(9) Development of principal joint trends.

(10) Normal and strike-slip movement along northeast-trending high-angle faults and renewed movement along

Stony Creek fault.

(11) Intrusion of Triassic (?) diabase dikes into the Marshall-Pedlar rocks.

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APPENDIX

List of samples, sample locations, thin section, x-ray patterns and point counts. Sample locations on Plate 5.

Sample	Location	Formation	Thin Sections	X-ray	Point Count
1	F-11	Pedlar			
2	F-11	Pedlar			
3	F-11	Pedlar			
4	E-11	Pedlar	1		
5	E-11	Catoctin			
6	F-8	Catoctin	1		
7	F-8	Catoctin	2		
8	F-8	Catoctin	1		
9	F-8	Catoctin	1		
10	F-8	Catoctin	1		
11	D-9	Catoctin	1	X	X
12	D-10	Catoctin	1		
13	C-12	Catoctin			
14	C-12	Catoctin	2		
15	C-12	Catoctin	1		
16	C-12	Catoctin	2		
17	C-12	Pedlar	1		
18	C-12	Pedlar	1		
20	C-12	Catoctin	2		
21	C-12	Catoctin	1		
22	C-12	Catoctin	1		
23	F-8	Catoctin	1		
24	F-8	Catoctin	1		
25	F-8	Catoctin	1		
26	F-8	Catoctin	1	X	
27	F-8	Catoctin	2		
28	F-8	Catoctin	2		
29	F-8	Catoctin	3		
30	E-7	Catoctin	1		
31	E-7	Catoctin	1	X	
32	E-8	Catoctin	1		
33	E-8	Catoctin	2	X	X
34	D-9	Catoctin	1		X
36	E-8	Catoctin	1		
37	E-8	Catoctin	1		
38	E-8	Catoctin	1	X	
39	G-12	Pedlar	1		
40	G-11	Swift Run	3		
42	G-11	Swift Run	1		

Sample	Location	Formation	Thin Sections	X-ray	Point Count
43	G-11	Swift Run	1		
44	D-11	Swift Run	1		
45	G-10	Catoctin	1		
46	E-10	Catoctin	1		
47	E-9	Catoctin	1		
48	E-9	Catoctin	1		
49	E-10	Catoctin	1		X
50	E-10	Catoctin	3	X	
51	E-10	Catoctin	7		
52	F-11	Pedlar	2		
53	F-11	Catoctin	2	X	
54	F-11	Pedlar	-		
55	F-11	Swift Run	1		
56	F-11	Pedlar	1		
57	F-11	Pedlar	1		
58	F-11	Pedlar	1		
59	F-11	Catoctin	1		
60	F-11	Catoctin	1		
61	F-11	Catoctin	1		
62	F-11	Catoctin	-		
63	G-11	Catoctin	1		
64	G-10	Catoctin	1		
65	F-11	Catoctin	1		
66	D-10	Catoctin	1		
67	D-10	Catoctin	1	X	
68	D-10	Catoctin	1		
69	D-10	Catoctin	1		
70	E-8	Catoctin	1	X	
71	E-8	Catoctin	1		
72	F-8	Catoctin	1		
73	F-9	Catoctin	2	X	
74	F-9	Catoctin	3	X	
75	E-11	Pedlar	1		
77	E-11	Pedlar	2		
78	F-10	Catoctin	2	X	X
79	F-10	Catoctin	2		
80	F-10	Catoctin	3	X	
81	F-10	Catoctin	1		
86	D-11	Swift Run	1		
90	F-9	Swift Run	1		
91	F-9	Swift Run	1	X	X
92	G-9	Swift Run	1		
94	G-9	Swift Run	1		
107	D-11	Swift Run	1		
112	D-11	Swift Run	1		
113	D-11	Pedlar	2		

Sample	Location	Formation	Thin Sections	X-ray	Point Count
114	J-8	Catoctin	8	X	
115	E-11	Pedlar	3		
117	F-13	Catoctin	1		
121	G-11	Pedlar	1		
123	G-12	Pedlar Catoctin	3	X	
124	H-12	Pedlar Catoctin	3		
125	G-11	Pedlar	2		
126	G-11	Swift Run	1	X	
128	G-11	Pedlar	1		
130	G-8	Catoctin	2	X	
131A	G-8	Catoctin	3	X	X
133	I-8	Catoctin	1		
134	I-8	Pedlar	1		
135A	I-8	Pedlar	1		
137A	K-13	Marshall	1		
137B	J-8	Catoctin	1		
138	E-12	Pedlar	1		
139	F-13	Catoctin	1		
140	G-13	Swift Run	1		
141	G-13	Pedlar	5		
142	H-13	Pedlar	1		
143	H-13	Marshall	1		
144	H-13	Marshall	1		
145	H-13	Marshall	1		
146	I-13	Marshall	1		
148	I-13	Marshall	1		
149	I-13	Marshall	1		
152	I-14	Triassic dike	2		
164	I-12	Pedlar	4		
169	H-12	Swift Run	2		
176	F-8	Catoctin	1		
177	F-7	Catoctin	1		
178	E-8	Catoctin	1		
180	I-2	Loudoun	2		
181	I-2	Loudoun	1		
183	I-12	Swift Run	1		
184	I-12	Marshall	1		
186	I-12	Swift Run	1		
187	I-11	Swift Run	-		
193	I-10	Marshall	2		
194	I-12	Marshall	1		
196	I-13	Marshall	1		
197	I-13	Marshall	1		

Sample	Location	Formation	Thin Sections	X-ray	Point Count
199	K-5	Pedlar	1		
201	K-5	Pedlar	4		
202A	K-5	Pedlar	1		
203	K-5	Pedlar	-		
204	E-11	Pedlar	1	X	
206	K-5	Pedlar	1		
207	K-5	Marshall	1		
208	K-5	Pedlar	1		
209	K-5	Pedlar	1	X	
210	K-5	Marshall	1		
211	K-5	Marshall	2		
212	K-6	Pedlar	2		
214	K-6	Swift Run	1	X	
215	K-6	Pedlar	2		
217	L-5	Pedlar	2		
218	M-5	Marshall	1		
219	L-8	Marshall	2		
222	K-8	Catoctin	-	X	
224	F-11	Pedlar	1	X	
226	J-9	Pedlar	1		
227	F-11	Catoctin	1		
228B	D-11	Catoctin	1	X	
229	D-11	Catoctin	-		
237	I-2	Loudoun	1	X	X
266	A-9	Antietam	1		
269	A-6	Antietam	1		
277	A-11	Loudoun	2		
281	B-10	Loudoun	15		
283	A-12	Catoctin	5	X	
287	A-12	Catoctin	5	X	
288	A-12	Catoctin	1		
289	A-12	Catoctin	1		
296	F-3	Catoctin	1		
301	H-1	Harpers	1	X	
307	E-5	Weverton	1		
312	F-5	Loudoun	6	X	
313	F-6	Loudoun	4		
314	F-6	Loudoun	1		
315	E-6	Harpers Weverton	4	X	
316	E-6	Weverton	2		
325	E-5	Weverton	1		
335	G-4	Weverton	1		
343	M-5	Pedlar	1	X	
344	M-5	Pedlar	2		
370	A-15	Marshall	-		

Sample	Location	Formation	Thin Sections	X-ray	Point Count
435	M-3	Marshall	-		
438	N-4	Pedlar	-		
439	N-4	Triassic dike	-		
447	M-6	Marshall Catoctin	-	X	
449	I-15	Lovingston	-		
474	N-10	Marshall	-		
478	M-10	Lovingston	-	X	
488	C-13	Pedlar	-	X	

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GEOLOGY OF THE HUMPBACK MOUNTAIN AREA OF THE BLUE RIDGE IN NELSON AND AUGUSTA COUNTIES, VIRGINIA

Mervin Jerome Bartholomew

Abstract

Mapping of a 190 square km area along the western flank of the Blue Ridge anticlinorium in central Virginia has defined three major rock groups: (1) the earlier Precambrian Virginia Blue Ridge Complex; (2) the Late Precambrian Catoctin Group; and (3) the Early Cambrian Chilhowee Group. The Virginia Blue Ridge Complex is subdivided from oldest to youngest into the Pedlar, Marshall and Lovingston formations. The Pedlar and Marshall formations, partially of metasedimentary origin, were metamorphosed to the granulite facies, retrograded to as low as the greenschist facies, and deformed into a series of east-trending folds prior to Late Precambrian. The relationships between the two metamorphic events and structural deformation was not determined. An angular unconformity separates the Catoctin Group from the Virginia Blue Ridge Complex upon which 300 m of topographic relief was developed.

The Catoctin Group is subdivided from oldest to youngest into the Swift Run, Catoctin and Loudoun formations which are subdivided into phyllitic units of both sedimentary and pyroclastic origin separated by sequences

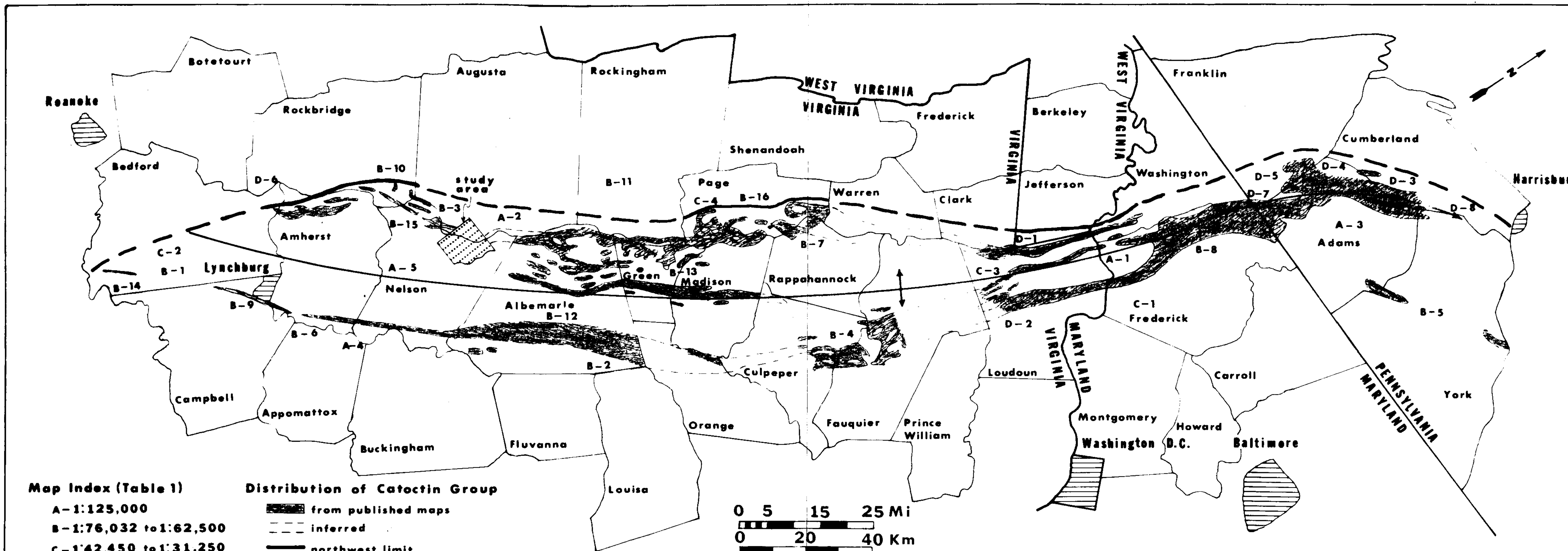
of greenstone flows. Early Catoctin Group volcanism, originating *in situ* from northeast-trending dike complexes, was accompanied by normal faulting along the northwest-trending Stony Creek fault. Swift Run and lower Catoctin Formation sediments were transported principally from the southeast quadrant or were derived *in situ* from saprolite and colluvial deposits. Loudoun and upper Catoctin Formation arkosic sediments were derived from localized sources, west of the Blue Ridge, and transported in a southwesterly direction parallel to the Appalachian structural trend. Pyroclastics of the Catoctin Group probably were derived from a northwest source. The upper Catoctin Formation is older than 700 my.

Paleozoic metamorphism of the Catoctin Group altered the lava flows toward the following greenschist assemblages of high oxidation state: quartz and albite plus (1) epidote, penninite and magnetite; or (2) penninite, magnetite and calcite. Subsequently, metamorphism indicative of a lower oxidation state altered the lavas toward the greenschist assemblage of: quartz, albite, epidote and actinolite. Actinolite content increases with depth at the expense of relict pyroxene, penninite and magnetite. Metamorphosed pyroclastics are characterized by sericite, quartz, magnetite and/or hematite.

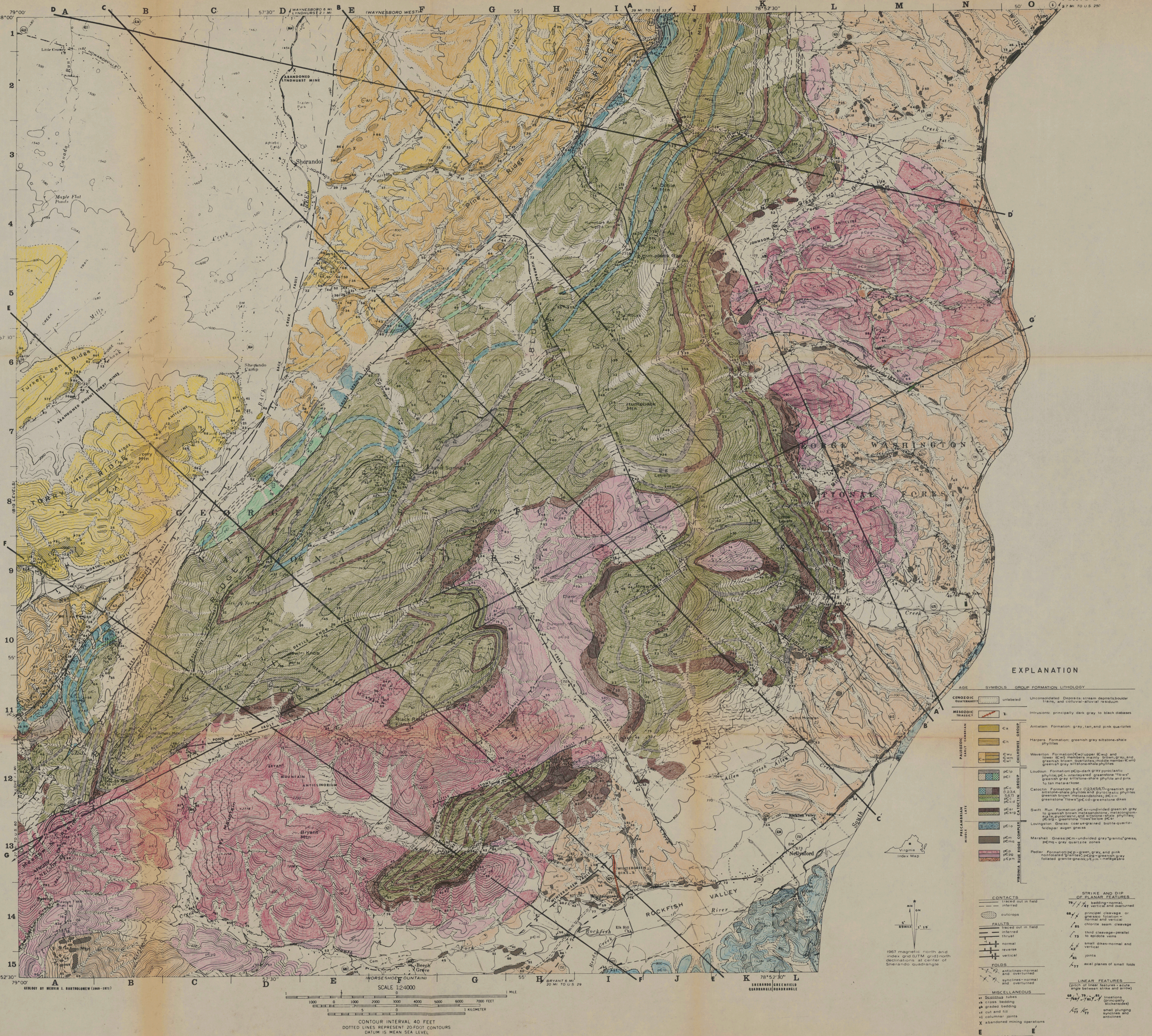
The Chilhowee Group is subdivided, from oldest to youngest, into the Weverton, Harpers and Antietam formations. A thrust fault zone separates metamorphosed Weverton and Harpers clastics from unmetamorphosed Antietam clastics. Metamorphosed Chilhowee and Catoctin sediments are characterized by formation of the following metamorphic minerals: epidote, penninite, sericite, and microcrystalline quartz. Chilhowee detritus was derived from the west-northwest and transported eastward and/or parallel to the Appalachian structural grain.

Post-Precambrian deformation includes folding about northeast and north-trending axes followed by development of southeast-dipping cleavage during Paleozoic metamorphism. Cleavage development preceded northwestward thrusting of metamorphosed rocks of the Blue Ridge over unmetamorphosed Paleozoic rocks of the Valley and Ridge. Subsequently the principal joint trends were established followed by normal and strike-slip movement along northeast-trending high-angle faults. Intrusion of Mesozoic dikes post-dates high-angle faulting.

PLATE



Roanoke
Botetourt
Rockbridge
Augusta
Rockingham
WEST VIRGINIA
VIRGINIA
Frederick
Berkeley
WEST VIRGINIA
VIRGINIA
Franklin
Cumberland
Shenandoah
Bedford
D-6
B-10
study area
B-3
B-11
Page
C-4
B-16
Warren
Clark
Jefferson
Washington
D-5
D-4
D-3
Harrisburg
D-7
D-8
B-15
A-2
D-1
A-3
Adams
Amherst
C-2
B-1
Lynchburg
A-5
B-7
A-1
B-8
Green
B-13
Madison
Rappahannock
C-3
B-14
Nelson
B-9
Albemarle
B-12
C-1
Frederick
D-2
B-5
B-6
A-4
B-4
B-2
Culpeper
Loudoun
VIRGINIA
MARYLAND
Carroll
PENNSYLVANIA
MARYLAND
York
Campbell
Appomattox
Orange
Fauquier
Prince William
Montgomery
Howard
Washington D.C.
Baltimore
Buckingham
Fluvanna
Louisa
N
Map Index (Table 1)
A - 1:125,000
B - 1:76,032 to 1:62,500
C - 1:42,450 to 1:31,250
Distribution of Catoctin Group
from published maps
inferred
northwest limit
0 5 15 25 Mi
0 20 40 Km



GEOLOGIC MAP OF THE HUMPBACK MOUNTAIN AREA, VIRGINIA

PLATE 3

EXPLANATION

apparent dip of bedding; patterns of Loudoun and and younger formations are parallel to bedding

apparent dip of cleavage or foliation; patterns of Catoctin and older formations are parallel to cleavage or foliation

formation symbols correspond to map symbols

GEOLOGIC STRUCTURE SECTIONS IN THE HUMPBACK MOUNTAIN AREA, VIRGINIA


PLATE 5
EXPLANATION
AGE SYMBOLS GROUP FORMATION LITHOLOGY
CENOZOIC QUATERNARY?
unlabeled
MESOZOIC TRIASSIC?
PALEOZOIC EARLY CAMBRIAN
CHILHOWEE GROUP
GEORGE WASHINGTON NATIONAL FOREST
ROCKFISH VALLEY
Sherando
Nellysford
Beech Grove
GEOLOGY BY MERVIN J. BARTHOLOMEW (1969–1971)
SCALE 1:24000
CONTOUR INTERVAL 40 FEET
DOTTED LINES REPRESENT 20 FOOT CONTOURS