MARYLAND GEOLOGICAL SURVEY

HOWARD AND MONTGOMERY COUNTIES

1964

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1964

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THE GEOLOGY OF HOWARD AND MONTGOMERY COUNTIES

BALTIMORE, MARYLAND
1964

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PREFACE

The geologic map of Howard County was published by the Maryland Geological Survey in 1940, and that of Montgomery County was issued in 1953 by the Maryland Department of Geology, Mines and Water Resources. One report covering both counties was planned because the areas include the same formations, and one report would avoid unnecessary repetition.

A thorough investigation of the rocks and structures was needed; the crystalline Piedmont Province is difficult to study because most of the formations are highly altered. In addition, since the first county was mapped in 1902, many geologists have participated in discussions about the age and correlation of the formations.

This volume presents much new material on the petrography and petrogenesis of the crystallines in Maryland, by C. A. Hopson. It is here presented in some detail because much of the controversy is due to lack of information. In addition, new approaches are being used not known 30 years ago to study crystalline rocks.

The problems discussed here had to be taken up by geological units, such as the Baltimore Gneiss, the Glenarm Series, or the Baltimore Gabbro, rather than by counties, and the rock types were studied where they were best exposed or most typically developed. Also, the maps of Baltimore and Harford Counties are quite old, and previous discussions centered there.

It is therefore necessary to consult the geologic maps of all the counties (Montgomery, Howard, Carroll, Baltimore, Harford) for detailed locality references. Many of the important localities are shown on Plate 7.

The introductory chapter by Cloos includes historical information and reviews some well-known data from the literature.

A chapter by Hopson on age dating provides the results of work done during the last 10 years with special emphasis on the Maryland portion of the Piedmont.

The structural geology is described by Ernst Cloos and is based on many years of work in Maryland.

George W. Fisher describes the Triassic rocks which occur in Western Montgomery County, and Emery T. Cleaves briefly reviews the mineral resources which are, unfortunately, not economically significant. Both Howard and Montgomery Counties will, in the future, probably provide living space for the population increase of the Baltimore-Washington metropolitan district. The trend is clearly shown in the population data.

This is primarily a geologic report and deals with the geologic units that occur in the two counties. Reports on forestry, wild life, soils, and water resources, which were part of earlier county reports, are omitted. A complete Water Resources Report appeared as Bulletin 14 in 1954. A detailed soil survey was published in 1961 by the Soil Conservation Service of the U. S. Department of Agriculture. When the older County reports were published there existed no special agencies such as the Departments of Forests and Parks, Game and Inland Fish, and the Weather Bureau to prepare and publish their own findings.

We wish to express our thanks to The Johns Hopkins University for providing space and equipment for the laboratory studies, to Mrs. Mary Gill for her assistance, to Mrs. Sakiko Olsen who did most of the modal analyses, and last but not least to Miss Agnes Creagh who edited the manuscript.

ERNST CLOOS

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HISTORY AND GEOGRAPHY OF HOWARD AND MONTGOMERY COUNTIES

BY

ERNST CLOOS

HOWARD COUNTY

Founding of Howard County

Howard County was part of Baltimore County from 1659 to 1726. At the separation of Anne Arundel and Baltimore Counties, Howard County became a part of the former. It was identified as Howard district in 1838 and became a separate political unit in 1880.

It was named after John Eager Howard, the elder.

Since its erection in 1850, the boundaries of the County have not been changed, and its area remains at 251 square miles.

Population

In 1890, ten years after its erection, Howard County showed a population density of 65 inhabitants per square mile, or a total of 16,315. In 1900 it had risen to 16,715. From 1900 to 1930 the population declined to 16,169.

In recent years the population has increased rapidly with the development of residential areas at the periphery of the Baltimore-Washington metropolitan district. The population changes are shown in Table 1 and Figure 2.

Election Districts

Six election districts were established within "Howard District" by 1868. They are Elkridge, Ellicott City, West Friendship, Lisbon, Clarksville, and Guilford and are shown on the topographic map of the county published and sold by the Maryland Geological Survey. Their population is listed in Table 1.

Transportation Facilities

The Baltimore and Ohio Railroad follows the northern, eastern, and southeastern boundary of the County. The total county line is about 78 miles long, of which the Baltimore and Ohio follows 46 miles. Of these 13 miles are on the main line between Baltimore and Washington, and the rest follows the Patapsco River and its South Branch as a freight line between Baltimore and Brunswick. At this date passenger railroad service is almost nil, except for special trains to the racetracks at Laurel.

The main highway through the length of the county is U. S. 40—a dual highway avoiding most settlements and with but a few traffic lights at its east end near the Patapsco River and north of Ellicott City. The other main road is U. S. 29 to Washington; it branches off U. S. 40 west of Ellicott City. This highway has put the eastern portion of the county within easy commuting distance to Baltimore and Washington. All other parts of the county are readily accessible by hard-surface roads.

TABLE 1

POPULATION OF HOWARD COUNTY, 1890–1960

DATE	TOTAL	DENSITY PER SQ. MILE	DISTRICT					
			1. Elkridge	2. Ellicott City	3. West Friendship	4. Lisbon	5. Clarksville	6. Guilford
1890	16,315	65						
1900	16,715	66	2,498	3,745	2,233	3,082		
1910	16,106	64	2,365	3,403	2,100	2,931		
1920	15,826	63	2,483	3,434	1,892	2,738		
1930	16,169	64	2,977	3,558	1,798	2,645	2,031	3,160
1940	17,175	68	3,229	3,778	1,974	2,410	2,304	3,480
1950	23,119	92	4,607	5,235	2,348	2,828	2,551	5,550
1960	36,152	143	7,262	9,575	2,721	3,309	4,119	9,166

MAPS

The entire area is covered by 15-minute topographic quadrangle sheets on the scale of 1:62500 (1 inch equals 1 mile) and issued by the U. S. Geological Survey. In addition maps at the scale of 1:24000 are available as shown in Figure 3. The Maryland Geological Survey (Johns Hopkins University, Baltimore 18, Md.) publishes maps showing topography, election districts, and the geology of counties at the scale of 1 inch equals 1 mile.

Howard County is also included in the Baltimore sheet at the scale of 1:250000 and the relief map at that scale, both published by the U. S. Geological Survey, (Washington 25, D. C.).

Other maps are issued by various agencies, such as the Maryland-National Capital Park and Planning Commission.

MONTGOMERY COUNTY

FOUNDING OF MONTGOMERY COUNTY

Montgomery County was established by the Constitutional Convention in 1776 and named after General Richard Montgomery. Since then the boundaries have remained almost unchanged. Prior to 1776 its territory was the southeastern portion of Frederick County. Nearly a century earlier settlements were established in the vicinity of Georgetown along Rock Creek and in the neighborhood of Spencerville and along Sligo Creek. The area of the county is 497.04 square miles of land and 7.35 square miles of water.

POPULATION

The population of Montgomery County has grown from 18,003 inhabitants in 1790 to 340,928 in 1960 (Tables 2, 3; Fig. 4). From 1790 to 1850 the population declined and by 1860 rose again to 18,000. From then on it increased. This fluctuation may be due to the method of census taking and counting. Until 1920 there was a steady increase; after that time some of the population of the Washington area began to spill over into Montgomery County. This is most emphatically shown between 1940 and 1960.

Two election districts were plotted also for comparison: 12 (Damascus) and 13 (Wheaton). The 12th district has changed little, and less than the county average; its population has increased from 1770 to 4448 in 170 years. The 13th district grew during

FIGURE 1. Location of Howard and Montgomery Counties

FIGURE 2. Population trend in Howard County 1890–1960

the same period from 3943 to 163,330 and is now 41 times the original figure. If the town of Wheaton is added, the total population of the area is 217,965 or 55 times that of 170 years ago.

TRANSPORTATION FACILITIES

The Baltimore and Ohio Railroad traverses the county from Silver Spring to Dickerson, but the only regularly scheduled passenger stop is at Silver Spring. The county is crossed by the dual highway from Washington to Frederick, connecting there with U. S. 40 from Baltimore. Many hard surface roads criss-cross the county and render all parts easily accessible.

The Chesapeake and Ohio canal, which was an important waterway until 1924, is now abandoned and has been incorporated into the National Park System because it makes many parts of the very beautiful Potomac River valley accessible.

ELECTION DISTRICTS

In 1798 the County was divided into five election districts. In 1876 a referendum was authorized on dividing the county into 12 districts, and accordingly districts 6, 7, 8, 9, 10, 11, and 12 were created between 1878 and 1884. A thirteenth district was created in 1886.

PHYSIOGRAPHY

RELIEF

The relief and drainage pattern of the counties result from the erosion of rocks of very unequal resistance against erosion. The differences in resistance are due to geological processes such as folding, faulting, uplift, and metamorphism or intrusion of

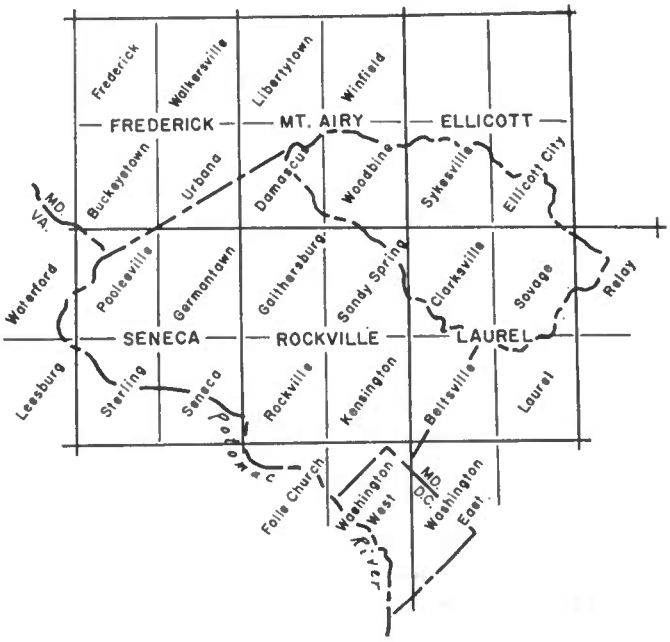

FIGURE 3. Topographic maps covering Howard and Montgomery Counties. Names are those of quadrangles published by U. S. Geological Survey. Horizontal scale 1:62,500; diagonal, 1:31,680 or 1:25,000.

TABLE 2

POPULATION OF ELECTION DISTRICTS IN MONTGOMERY COUNTY

1900–1960

DISTRICTS	1900	1910	1920	1930	1940	1950	1960
1. Laytonsville	1,981	1,866	1,599	1,687	1,813	1,749	2,133
2. Clarksburg	2,013	1,995	1,847	1,692	1,558	2,157	3,136
3. Poolesville	2,343	2,170	1,854	1,477	1,724	1,600	1,920
4. Rockville	3,488	3,459	3,442	4,684	5,995	10,558	37,896
5. Colesville	2,192	2,234	2,301	2,306	4,045	6,773	20,904
6. Darnestown	1,675	1,589	1,489	1,566	1,682	2,697	3,526
7. Bethesda	2,027	3,217	4,757	12,018	26,114	45,807	83,197
8. Olney	3,321	2,826	2,617	2,492	2,601	3,608	5,320
9. Gaithersburg	2,383	2,623	2,570	3,256	3,861	5,398	8,760
10. Potomac	1,630	1,329	1,125	1,135	1,828	1,956	4,247
11. Barnesville	1,685	1,865	1,751	1,673	1,735	1,856	2,071
12. Damascus	1,770	1,809	1,740	1,843	2,079	2,825	4,448
13. Wheaton	3,943	5,107	7,829	13,377	28,877	77,417	163,330

igneous rocks. Many different kinds of rock of varying hardness now side by side were deposited one above the other. The rocks slowly disintegrate due to weathering, and running water carries the debris away.

Montgomery and Howard Counties are almost entirely within the Piedmont Province, which spreads out southeast of the Appalachian Mountains. Along the southeastern edge of both counties, between Washington and Elkridge, Coastal plain sediments lap

TABLE 3

POPULATION OF MONTGOMERY COUNTY

	TOTAL		TOTAL
1790	18,003	1880	24,759
1800	15,058	1890	27,185
1810	17,980	1900	30,451
1820	16,400	1910	32,089
1830	19,816	1920	34,921
1840	15,456	1930	49,206
1850	15,860	1940	83,912
1860	18,322	1950	164,401
1870	20,563	1960	340,928

onto the crystalline basement rocks of the Piedmont. The boundary is here not a sharp one but an overlap of much younger formations on older ones (Figs. 5 and 8).

The lowest point in Montgomery County, at the Potomac River where it enters the District of Columbia, is 52 feet above sea level; the highest point is 873 feet, just inside Howard County at its west end south of Mt. Airy, Maryland.

Figure 5 shows the topography and relief of the two counties. The northwestern portion of the counties is the highest, and from here the average surface slopes southwestward and southeastward at approximately equal rates. The distance between the highest point (solid black in Fig. 5) and the lowest at Elkridge in the southeast is 27 miles, and the difference in elevation is 801 feet. The average gradient is therefore 29 feet per mile. From the highest point to the Potomac River at the district line the distance is 28.5 miles, and the elevation difference is 821 feet—a rate of 28.8 feet per mile. Between the highest point and the mouth of Seneca Creek the distance is 21 miles, the difference in elevation is 673 feet, and the rate is 32 feet per mile. It is rather surprising that the slope is almost the same to the southeast, south, and southwest or, in geological terms, across the general strike, diagonally across it, and parallel with it.

The surface of the Piedmont plateau is here not a simple, uniform slope gently inclined away from the Appalachian Mountains toward the coast, but in the two counties it is a surface inclined south and as much toward the Potomac River as toward the coast.

DRAINAGE

Montgomery County is drained by the Potomac River and its tributaries; Howard County is drained by the Patuxent and Patapsco Rivers. All three flow from northwest to the southeast.

The Potomac enters Montgomery County at Monocacy and is a well-entrenched broad stream. Its gradient is about 4 feet per mile above the Block House Falls about 1 mile below Seneca. From here to the District Line it forms many falls and rapids with an average gradient of 8.5 feet per mile.

The Patuxent River has its source in the Western and higher portion of Howard County at an elevation of about 850 feet. At Laurel, where it leaves the county, the elevation is about 200 feet. The gradient is therefore 23 feet, much steeper than the Potomac River gradient.

The Patapsco River has a gradient of 12 feet per mile between Elkridge and its con-

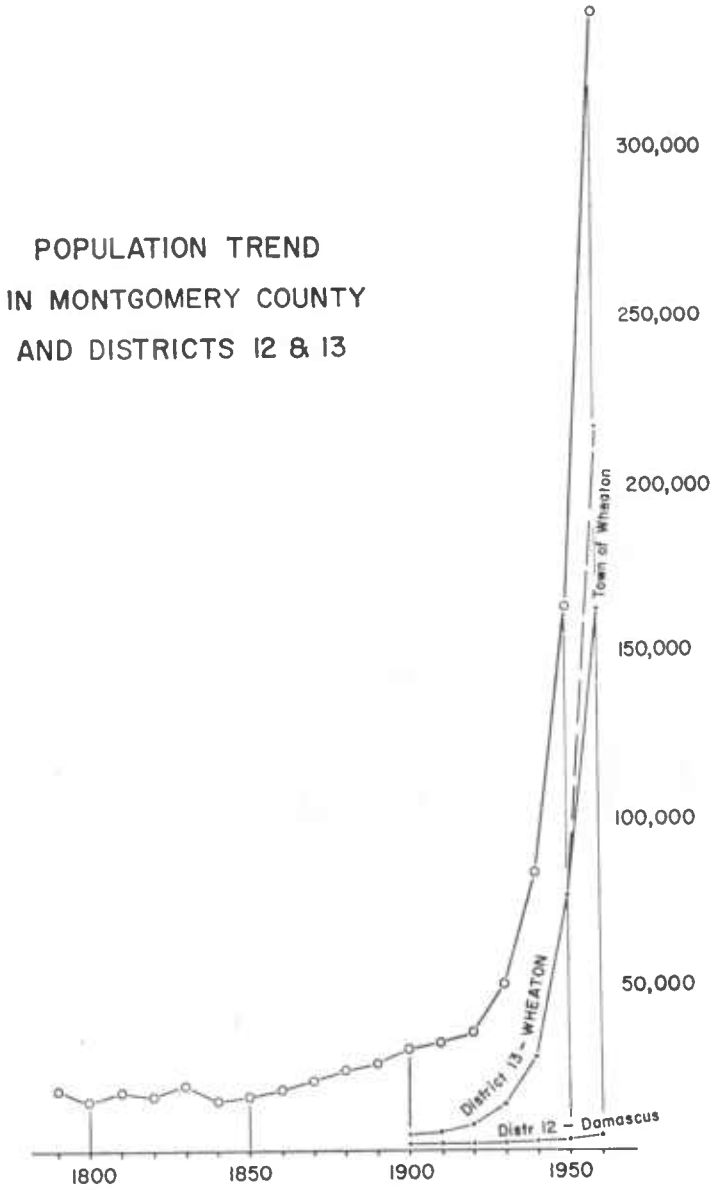

FIGURE 4. Population trend in Montgomery County and districts 12 and 13

fluence with the South Branch east of Mariottsville. From here to Parrs Spring near the western edge of the County, the gradient is 36 feet per mile.

Along all streams, prevalent falls and rapids render the valleys picturesque and have stimulated the building of dams and reservoirs.

RELIEF AND GEOLOGY

In Montgomery and Howard Counties the rocks are more altered, more crystalline, and generally harder in the southeast half or third of the counties than in the northwestern portion. To the northwest the grain size of the rocks and the intensity of meta-

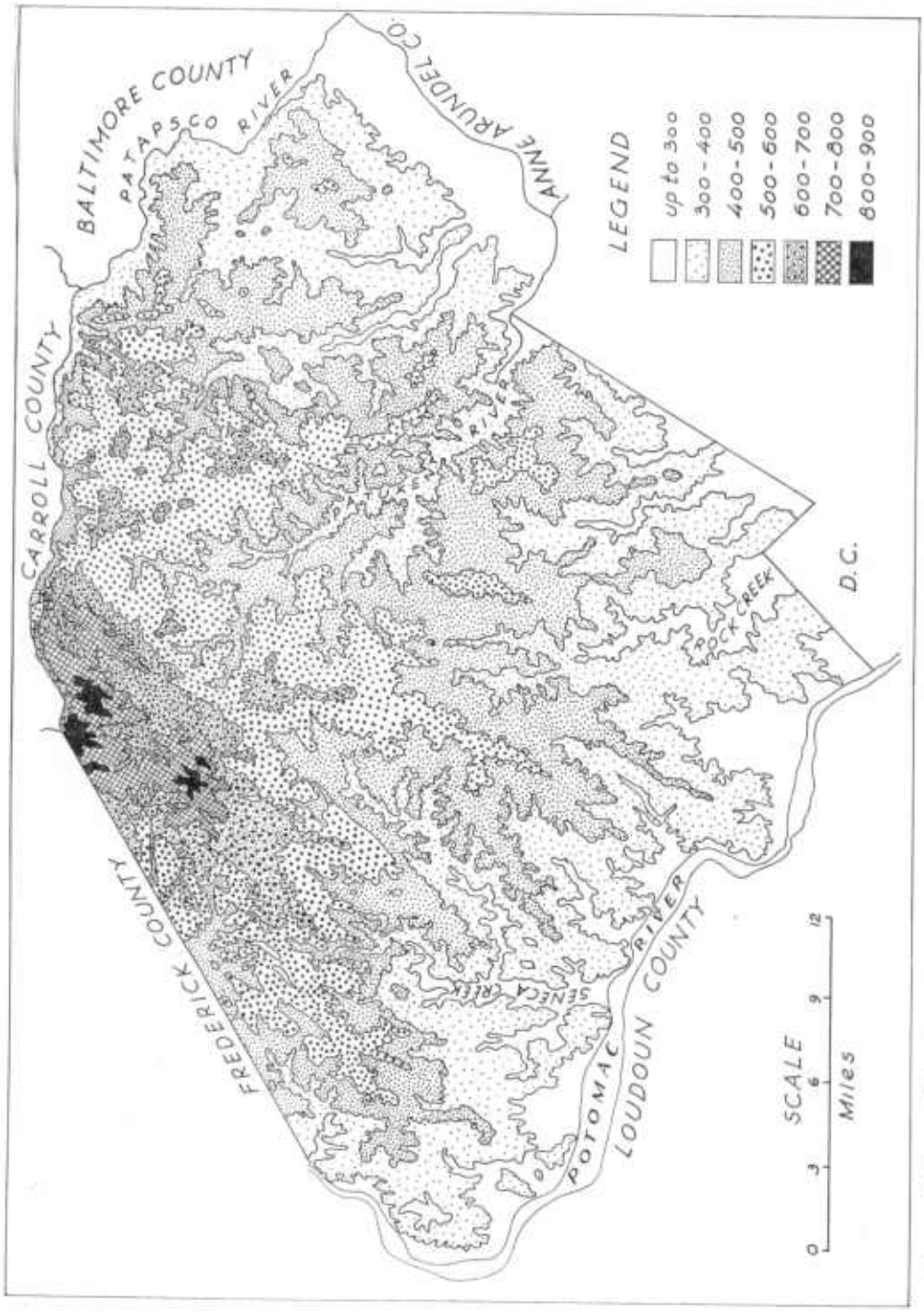

FIGURE 5. Map of Howard and Montgomery Counties showing drainage pattern and elevations. Drafted by Mrs. Josephine Spemann.

morphism decrease gradually. The southeast is made up of gneisses, quartzites, schists, and granites or mafic rocks. The northwestern portion is underlain by phyllites, slates, and much less altered formations. In the Potomac River bend the rocks are unaltered Triassic sandstones and shales.

The rocks of the western portion of the counties should be more extensively eroded and lower than those in the southeast; the fact that this is not so suggests that for a considerable portion of time the western areas have been elevated with respect to the southeastern areas. This is confirmed in the general descent of the Piedmont surface southwestward below the Coastal Plain sediments. The old surface of the Piedmont province has been tilted and is still preserved below Coastal Plain sediments southeast of the county lines. Between the Coastal Plain and the Piedmont is a sudden drop, shown in the many falls and rapids in all the streams and in the removal of the Coastal Plain sediments on the Piedmont Plateau.

The distribution of high and low areas shown in Figure 5 follows three principles:

(1) An erosion surface was formed without regard for rock hardness and distribution of formations; the surface must have been mature.

(2) The surface was tilted and raised with a rejuvenation of erosion and superpositions of a drainage pattern. For the Potomac River this pattern was influenced by a stream which flowed eastward and cut into the plain. For the Patuxent and Patapsco Rivers the inclination of the plain was the determining factor since both rivers originate in the high western area.

(3) Courses of tributaries are at many places determined by the rock distribution. Figure 5 shows some of the regional rock trends. Formational boundaries generally trend northeast-southwest, and this direction coincides with some of the major valleys. One may say that the deeper structure is seen through the erosion pattern, as a road system may show through a blanket of snow.

THE TRIASSIC ROCKS OF MONTGOMERY COUNTY

BY

GEORGE W. FISHER

NEWARK GROUP

General Relations

In the western corner of Montgomery County, nearly flat-lying red Triassic sandstones and shales unconformably overlie tightly folded low-grade metamorphic rocks. The red beds are part of an extensive belt of Triassic rocks stretching southward into Virginia and northward into Frederick County and southern Pennsylvania (Fig. 6). The geology of the Triassic rocks was described by Dorsey (1918), Roberts (1928), and Palmer (1949).

Quaternary alluvium and terrace gravel disconformably overlie the red beds in places.

Topography developed on the red beds is gently rolling, except near the Potomac River, where steep-sided valleys are locally incised. As a result of the low relief, exposures are generally poor.

Triassic sandstone, quarried near Seneca, was much used as building stone during the last century. Known as "Seneca redstone," it was used in the locks of the old Potowmack Canal (1774), the Smithsonian Institution of Washington (1847), and innumerable brownstone houses in Baltimore and Washington.

Stratigraphy, Correlation, and Age

Up to 200 feet of brownish or yellow quartz pebble conglomerate lies at the base of the red beds. Interbedded lenses of red arkose and shale are common. From 900 to 2000 feet of fine-grained arkosic sandstones and coarse siltstones with numerous lenses of purple shale overlies the conglomerate. In the southern part of the outcrop area, nearly 5000 feet of deep red shale lies next above. These shales appear stratigraphically equivalent to about 4000 feet of sandstones and conglomerates in the northern half of the area, but the two sequences may be in fault contact. Thicknesses given are approximate and neglect minor faulting.

Lithologically similar beds may be traced in nearly continuous outcrop into southern Pennsylvania (Fig. 6), where the lowermost 7000 feet of the Newark Group is named the New Oxford Formation from exposures of red shale and arkosic sandstone near New Oxford, Pennsylvania (Stose, 1932, p. 56). Upper Triassic reptile bones and conifers are locally found. Because of lateral continuity with this type section, as well as lithologic and stratigraphic similarity, the red beds of Montgomery County are correlated with the New Oxford Formation.

Lithology and Interpretation

Shales and Siltstones

Most of the shales are bright red or maroon; a few patches of green occur, localized around plant fragments. X-ray analysis of a specimen from 1.0 mile southwest of Seneca shows muscovite, chlorite, illite, hematite, and quartz. Bedding is mostly indistinct or absent; it is prominent only in highly micaceous shales, which show strong bedding

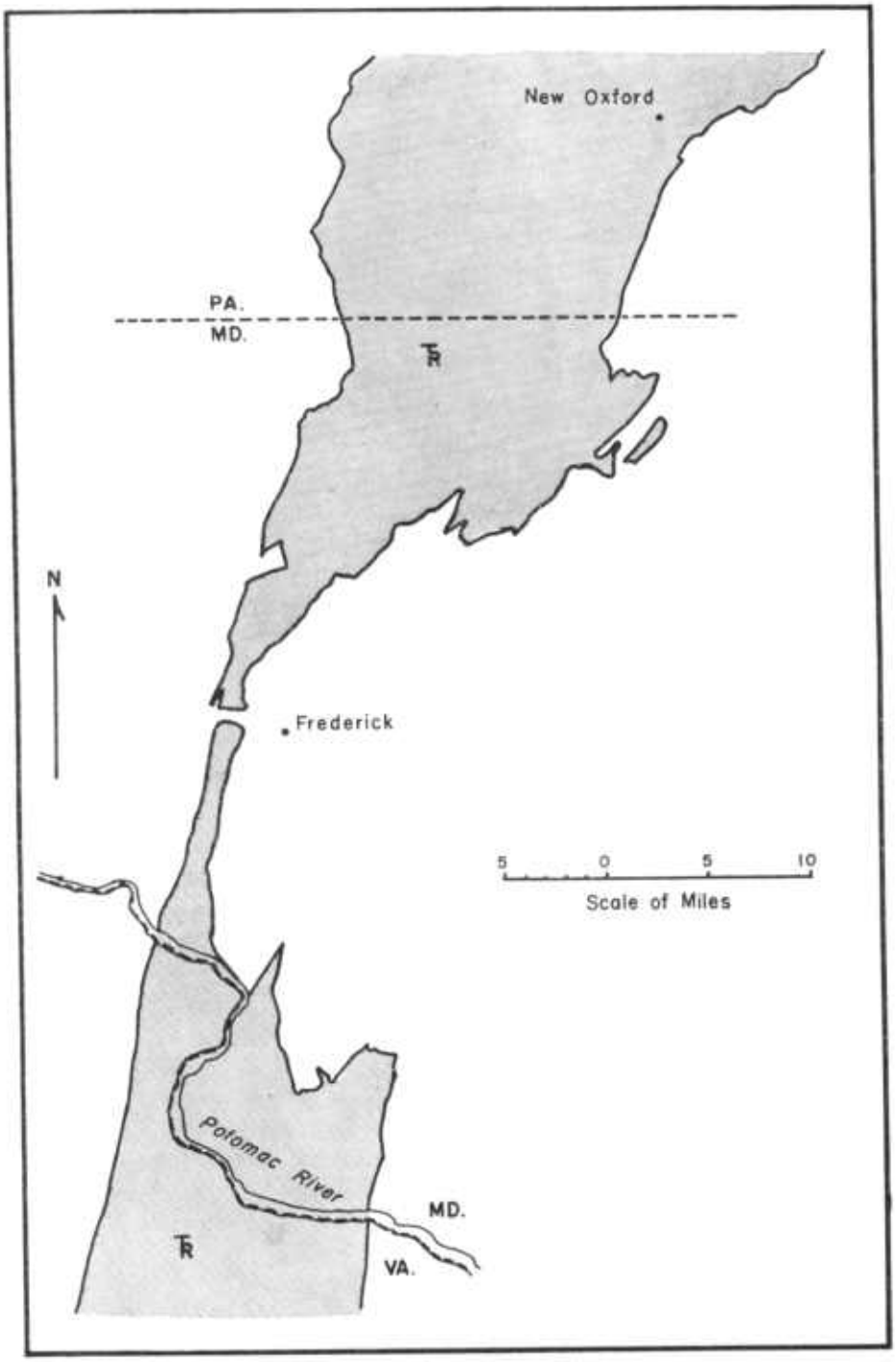

FIGURE 6. Distribution of Triassic sedimentary rocks (stippled) in Maryland and adjacent parts of Pennsylvania and Virginia.

fissility. Most nonmicaceous shales exhibit a chunky fracture only crudely related to original bedding. Rain-drop impressions, plant stems, and mud cracks are locally well developed and provide evidence of a continental depositional environment. Ripple marks are also present.

The siltstones are commonly deep red; bedding is well shown by thin color laminations (0.1–2 inches) and more widely spaced (0.2–5 inches) bedding fissility. Small-scale ripple cross-beds and ripple marks are common. Quartz and subordinate feldspar are the chief silt-sized constituents. Muscovite flakes commonly lie along the bedding planes. As indicated by the deep-red color, hematite is abundant.

The shales and siltstones are the most deeply colored of the red beds; perhaps finely divided hematite was concentrated in the fine-grained sediments by mechanical sorting. The abundance of hematite indicates deposition and lithification under strongly oxidizing conditions.

Sandstones

In hand specimen, the sandstones vary from gray or yellowish brown to light red. Average grain size is near 0.5 millimeter. Coarser sands are chiefly associated with conglomerate lenses. Dull, turbid grains of feldspar and flakes of muscovite are easily distinguished in most specimens. Fragments of phyllite 1–2 millimeters across are found in places, especially in the lowermost beds. Isolated quartzite pebbles are locally present; with increase in pebble content, the sandstones grade into conglomerates. Plant fragments are abundant.

Beds range from 6 inches to 2 feet thick; color laminations as thin as 1 inch are found. Planar and festoon cross-bedding up to 2 feet across are locally well developed. Cut-and-fill structures are common, especially in the coarsest sandstones. Ripple marks and ripple cross-bedding occur locally. These structures, together with the mud cracks and rain-drop impressions in the shales, suggest fluvial deposition.

Under the microscope, the sandstones show complex, immature mineralogy (Table 5). Following Pettijohn's classification (1957, p. 291), samples 3-A, 5-A, 11-A, and 32-A are arkoses, and 26-A is a subarkose. Specimen 5-A contains almost enough rock fragments to qualify as a subgraywacke. True lithic sandstones are rare but may occur in the lowest part of the sequence.

The abundance of detrital feldspar, chiefly plagioclase, requires a source area with feldspar-rich rocks. Plagioclase is found both as highly weathered grains, much altered to sericite and clay minerals, and as fresh grains. This association suggests strong chemical weathering at the source and rapid erosion and transport.

Epidote, chlorite, biotite, garnet, and phyllite fragments suggest a source area underlain partly by low- to medium-grade metamorphic rocks. Sericite intergrowths ("alterite") resembling pseudomorphs of kyanite or some other aluminum-excess mineral suggest the presence of high-grade metamorphic rocks. The mineralogical and textural immaturity of the sandstones suggests that the source was nearby.

These inferred characteristics of the source area most closely fit the region of metasediments and associated quartz diorites and granodiorites stretching east from the Triassic belt to the Coastal Plain.

Analysis of cross-bedding (Fig. 7) indicates that streams flowed chiefly northeast and southwest. The faults in the area show a similar trend and may have controlled the direction of streams. Fanglomerates developed along faults in Frederick County (Stose and Stose, 1946, p. 87) prove active faulting during deposition. The few cross-beds

FIGURE 7. Rose diagram showing percentage of down-current direction in each 30-degree compass sector. Based on 26 readings of cross-bedding in the sandstones and conglomerates of the Newark Series in Montgomery County.

trending northwest may represent tributary streams flowing from the source area to the southeast.

Conglomerates

Red to yellowish-brown or gray conglomerates occur as massive, thick-bedded lenses interstratified with coarse sandstones. Channels and cut-and-fill features several feet thick are well displayed. Cross-bedding is rare, probably owing to poor development of bedding, especially in weathered outcrops.

The matrix is most commonly coarse arkosic sandstone but is in places shale or siltstone. Pebbles are mostly vein quartz or metaquartzite, moderately well rounded. Phyllite pebbles are more abundant in the lowest beds. Pebble size ranges up to 4 inches and seems independent of stratigraphic level or geographic position. Data for three typical random samples are given in Table 4. These relations are compatible with an eastern source, although other interpretations are possible.

STRUCTURE

Faults are the only major tectonic structures affecting the Newark Group in Montgomery County. The three faults shown on the geologic map of Montgomery County strike north or northeast, with the east side downthrown. No fault planes have been observed; the faults are assumed to be normal. Faults are difficult to recognize, owing to a lack of distinctive marker beds which might indicate offset or repetition. In a few places, small faults of a few feet displacement are seen in outcrop.

Folds were not observed in Montgomery County but have been described from Frederick County by Palmer (1949). The dips (8°–25°) may be due to tilting of fault blocks.

Joints are well developed throughout the Newark Group, especially in the more massive arkoses. Statistical analysis of 60 joints measured in the Government quarry, 1.1 miles of Rushville, indicates at least two major sets. The most prominent set, parallel to local faults, strikes north and dips 70°E. A more complex set, perhaps actually three closely similar sets, strikes east and dips vertically. The joints may be related to the forces that produced the faults.

TABLE 4

CONGLOMERATE PEBBLES, NEW OXFORD FORMATION

	41-A	28-A	28-B
Pebble material			
vein quartz	15	10	12
metaquartzite	7	14	13
phyllite	2	1	
pegmatite	1		
Total pebbles	25	25	25
Average roundness	4.38	5.75	5.72
Average length	3.11 cms	5.03 cms	4.81 cms

(41-A) Road cut on Md 28, 0.6 miles east of Dawsonville, Md.
(28-A) Road cut on Md 419, 2.2 miles N-NW. of Beallsville, Md.
(28-B) Same as 28-A, 10 feet stratigraphically higher.

DIABASE

GENERAL RELATIONS

Diabase dikes and sills intrude Triassic and pre-Triassic formations in western Montgomery County and eastern Howard County. Although good exposures are rare, the extent of these intrusions may be traced by yellowish-brown spheroidal boulders of weathered diabase. The largest body entirely within Montgomery County is a sill near Boyds, covering roughly 3.5 square miles. A dike near Dickerson may be traced for at least 35 miles into Frederick County. In Howard County a swarm of parallel dikes northwest, west, and southwest of Ellicott City trends north-northeast and cuts across crystalline basement rocks. The dikes continue to the north into Baltimore County and to the south into Montgomery County. Chemical analyses of typical Triassic diabases from the Piedmont are given in Table 6. Modes of diabases from Montgomery and Howard Counties are given in Table 7.

Similar intrusions and lava flows are closely associated with Triassic sedimentary rocks from Virginia to Massachusetts.

FIELD RELATIONS AND PETROGRAPHY

Boyds Sill

The essentially flat-lying sill near Boyds has been intruded along the unconformity between the Newark Group and the underlying Wissahickon Schist. The sill may be up to 700 feet thick, roughly estimated from its outcrop area and inferred attitude; at least 200 feet is proven by core drilling.

A typical specimen from the center of the sill contains phenocrysts of hypersthene in a groundmass of plagioclase laths, augite, and fine-grained hypersthene. The two groundmass pyroxenes enclose each other and appear to have crystallized in equilibrium. Edges of hypersthene phenocrysts are irregular and enclose groundmass minerals. Probably these crystals continued to grow during crystallization of groundmass hypersthene.

Samples from near the margin contain considerably more olivine (mostly replaced by a light-yellow montmorillonoid, probably saponite) and correspondingly less hypersthene. The texture is fine-grained. A representative specimen from within 500 feet of the western contact contains rounded olivine pseudomorphs enclosed in large subhedra

TABLE 5

SANDSTONE MINERALOGY, NEW OXFORD FORMATION

	3-A	5-A	11-A	26-A	32-A
Quartz	M	M	M	M	M
Microcline	M	m	M	M	m
Plagioclase	M	M	M	M	M
Rock fragments	m	M	m	m	t
Orthoclase	m	t	m		m
Muscovite	m	m	m	m	m
Biotite	t	m		m	t
Epidote	t	t	m	m	m
Chlorite	m	m	t	t	m
Carbonate	t		m		
Hematite	m	m	m	t	m
Sphene	t	t	t	t	t
Leucoxene	t	t	t		t
Apatite	t	t	t	t	t
Tourmaline	t	t		t	
Garnet	t				m
Carbonaceous material	t				
"Alterite"	t	t	t	t	
Zircon	t	t			t
Magnetite	t			t	m
Rutile			t		
Chloritoid			t		
Sericite	t	t	t	t	
Quartz/feldspar ratio (from estimated mode)	2.4/1	2/1	2.4/1	3.5/1	3/1

M = 10% or more of rock; m = 1–10% of rock; t = less than 1% of rock.
(3-A) Along C & O Canal, 0.6 miles west of Rushville, Md.
(5-A) Along C & O Canal, 0.5 miles west of Rushville, Md.
(11-A) Along C & O Canal, 1.2 miles west of Rushville, Md.
(26-A) Alone C & O Canal, 1.3 miles N-NW. of Martinsburg, Md.
(32-A) Road cut on Md 419, 2.3 miles northwest of Beallsville, Md.

of hypersthene. Augite occurs as reaction rims on hypersthene, and partly enclosed plagioclase is zoned from labradorite to albite. Thin selvages of light-green hornblende are locally developed on augite and hypersthene and are partly altered to a brownish biotite. The biotite is closely associated with interstitial granophyric intergrowths of quartz and alkali feldspar. Olivine and hypersthene are partly altered to saponite (?). In this specimen, rapid cooling has "frozen" all stages in the differentiation of basaltic magma to granite. Rapid cooling can also explain the smaller grain size and the larger amount of olivine at the contact.

Dikes

Diabase dikes are intruded along northeast-trending fractures. A slight decrease in grain size toward contacts is common; otherwise the specimens examined are strikingly uniform. They differ from specimens of the Boyds sill in having one pyroxene (monoclinic) in place of augite plus hypersthene. Olivine was originally present in places but is now completely altered to yellow saponite (?). Plagioclase is in subhedral, normally

TABLE 6

CHEMICAL ANALYSES OF TRIASSIC DIABASE

	1	2	3
SiO_2	51.28	53.86	51.56
Al_2O_3	15.05	12.86	13.81
Fe_2O_3	1.12	1.75	2.08
FeO	9.31	11.80	11.32
MgO	7.97	5.12	7.40
CaO	11.42	9.66	10.08
Na_2O	2.03	2.54	2.08
K_2O	0.27	0.40	0.96
H_2O	0.48	0.46	—
TiO_2	0.78	1.13	1.48
P_2O_5	0.13	0.29	0.16
MnO	0.16	0.35	0.19
	100.90	100.63	100.00

(1) Diabase near New Market, Maryland; Shepherd (1938)
(2) Diabase cutting Cambrian schists of Albemarle Co., Va., Roberts (1928)
(3) Average diabase near Leesburg, Loudon Co., Va.; Roberts (1928)

TABLE 7

MODES OF TRIASSIC DIABASE

	H24-1	24-A	16-A
Plagioclase	54.3	58	50
Pigeonite	36.1	32	—
Augite	—	—	23
Hypersthene	—	—	20
Biotite	0.2	1	2
Hornblende	—	—	tr
Saponite	5.8	3	1
Micropegmatite	—	2	1
Magnetite/ilmenite	3.1	4	3
Celadonite	0.5	—	—
Apatite	tr	—	tr
	100.00	100	100
Points	1955	est.	est.

(H24-1) Old Frederick Road, 1½ mi north of Pine Orchard, Howard County.
(24-A) Center of large dike near Dickerson, Md.
(16-A) Near western contact of Boyds Sill, Montgomery County.

zoned laths, partly enclosed in a zoned monoclinic pyroxene, with 2V ranging from 20° at the core to nearly 40° at the rim (specimen 24-A, 1.5 miles west of Beallsville). The 2V shows that the pyroxene does not lie on the join diopside-hedenbergite but is zoned outward from pigeonite to subcalcic augite.

These data suggest that the pyroxene of the dikes began to crystallize at high temperatures, where a single pyroxene (pigeonite) was stable (Barth, 1952, p. 108). Rapid cooling prevented adjustment to equilibrium as the temperature dropped past the point where two pyroxenes should ideally form, and the pigeonite core persisted metastably, as the outermost rim followed the pyroxene solvus toward augite. In the Boyds sill,

on the other hand, slower cooling apparently maintained equilibrium to lower temperatures, so that two pyroxenes crystallized in equilibrium.

CONTACT METAMORPHISM

At their contacts the diabase dikes have produced narrow metamorphic aureoles. Recrystallization is generally slight and is limited to a zone less than 50 feet wide. Metamorphism at the Boyds sheet and along the dikes is essentially similar and will be considered together.

Arkoses and siltstones show few megascopic effects of recrystallization. Microscopically, quartz and feldspar appear somewhat recrystallized and interdigitated. Biotite is commonly broken down to chlorite, and muscovite is enlarged by recrystallization.

Pelitic rocks show more pronounced effects. Hematite in the Triassic shales is commonly reduced to magnetite; as a result, the shales are dark purple and in places black near diabase contacts. Those most intensely altered are compact and brittle and show conchoidal fracture. Spots of magnetite or epidote are locally developed. Original clays are partly recrystallized to feldspar, epidote (?), and muscovite. Quartz shows incipient recrystallization.

In contact with diabases schists show similar effects. Sericite and chlorite are slightly recrystallized so that schistosity and fracture cleavage are partly destroyed. Specimens from a salient of schist into the southeast side of the Boyds sheet show an abnormally high grade of metamorphism, presumably because the prong of schist was enclosed on three sides by diabase. The contact aureole there is at least 150 feet wide. The rocks are deep blue, dense, and structureless. Microscopic recrystallization products include cordierite, spinel, biotite, plagioclase, and corundum (?). Where quartz is present, spinel and corundum are absent.

CONCLUSIONS

Diabase dikes evidently cooled quickly, preserving a high-temperature mineral assemblage. A thick sill near Boyds was probably intruded at similar high temperatures, but cooled more slowly, resulting in a lower-temperature mineral assemblage. Equilibrium was not continuously maintained, so that a granophyric matrix developed. However, there is no significant differentiation. The slight mineralogical and textural variations within individual intrusions may be explained by more rapid cooling near contacts.

Narrow thermal metamorphic aureoles are developed at diabase contacts; only locally do the altered zones exceed 50 feet. Most important minerological changes are reduction of hematite to magnetite and slight recrystallization of clays and micas.

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REVIEW OF THE POST-TRIASSIC ROCKS

BY

ERNST CLOOS

CRETACEOUS SYSTEM

General Remarks

The Cretaceous rocks belong to the Potomac Group which has been subdivided, from base upward, into the lower Cretaceous Patuxent Formation and the upper Cretaceous Arundel and Patapsco Formations. In Montgomery and Howard Counties, only the Patuxent and Patapsco Formations have been recognized.

Patuxent Formation (Lower Cretaceous)

Name

The Patuxent Formation was named by Clark and Bibbins (1897, p. 490) after the Patuxent River in whose drainage basin the formation occurs.

Distribution

Figure 8 shows the distribution of the Patuxent Formation in Howard and Montgomery Counties along their southeast edge. The lower Cretaceous formations transgress unconformably over the much older and highly altered crystalline rocks. They rest on a plane inclined to the east at about 60 feet per mile. The drainage cuts through the sedimentary rocks into crystalline rocks, mostly of the Baltimore Gabbro complex in Howard County, and Laurel Formation and crystalline schists in Montgomery County and the District of Columbia. Coastal Plain sediments extended farther northwest before they were stripped off the older rocks by erosion.

Composition

The Patuxent Formation is composed of sand but includes well-rounded, polished pebbles and lenses of brightly colored clay. The sand is mixed with kaolin and mica. Fine and coarse sands prevail, and gravel and clay are also present. The sands are cross-bedded at many places. Ferruginous sandstones mixed with gravel are frequently seen in clay and gravel pits.

Paleontology

Organic remains are rare and consist of plants, chiefly silicified trunks of conifers and cycads. During the excavation of a reservoir at Washington a silicified trunk 50 feet long and several feet in diameter was found. A similar trunk about 4 feet thick was found near the University of Maryland (Miller, 1911, p. 86).

Berry (1911, pp. 90–94) lists 23 species of plants from Maryland and the District of Columbia, all of which occur also in the Arundel Clay or the Patapsco Formation.

Thickness

The observed thickness ranges from a few feet to 340 feet and increases eastward. On the basis of well data it may reach 500 feet farther east.

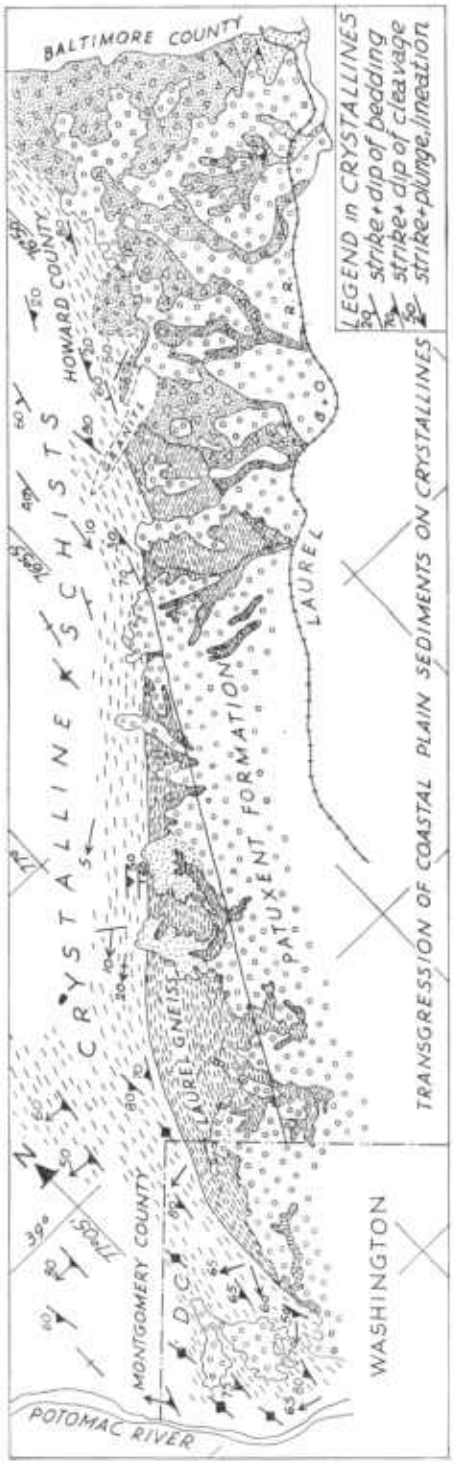

FIGURE 8. Transgression of Coastal Plain sediments on crystalline basement rocks. Drafted by Mrs. Josephine Spemann.

The Patuxent barely overlaps onto the crystalline rocks in Montgomery and Howard Counties and at no place is the complete section exposed. The formation thickens rapidly toward the southeast in Prince George's and Anne Arundel Counties. A more detailed description of it is given by Cooke (1952, p. 2).

PATAPSCO FORMATION (UPPER CRETACEOUS)

Name

The Patapsco Formation has been named by Clark and Bibbins (1897, p. 489) from typical occurrences in the valley of the Patapsco River, near Baltimore.

Distribution

The formation has been distinguished only in Montgomery County by Cloos and Cooke, 1953, where remnants of a broad band about 6 miles wide extend from the Patuxent River to the Patapsco River. The lower part of the formation is best exposed in the pits of the Washington Brick Company east of the Baltimore and Ohio Railroad in Prince George's County.

Composition

The Patapsco Formation consists chiefly of brightly colored and variegated clays which grade into lighter-colored sands and clays. Coarse sandy lenses are interstratified at some places. Some clays are dark and lignitic; others are laminated and may contain leaf impressions. Nodules of white or red clay are fossiliferous. The sands contain kaolin and are cross-bedded at some places.

Thickness

Reliable estimates of thickness are lacking, but Cooke (1953, p. 7) mentions "perhaps 200 or 300 feet" as a possible order of magnitude in Prince George's County.

Stratigraphic Relations

The Patapsco Formation unconformably overlies the Patuxent and transgresses the edge of the latter across some of the crystalline rocks. It is overlain by higher Cretaceous beds which do not reach across the county line.

Paleontology

Plant remains of the Patapsco Formation include ferns, horsetails, cycads, conifers, and the first advanced angiosperms (Cooke, 1952, p. 7).

A bone collected in 1943 in Prince George's County was identified by C. W. Gilmore as *Ornithomimus affinis*, a carnivorous dinosaur.

In addition there were reported "a few molluscan shells" (Miller, 1911, p. 90).

Brenner (1963) has reported on the spores and pollen of the Potomac Group.

TERTIARY SYSTEM

PLIOCENE SERIES

Tertiary formations appear on the geological map of Montgomery County only in the District of Columbia and immediately to the northeast at Tacoma Park.

BRYN MAWR GRAVEL

Name

Originally the term was used by Lewis (1880) for high-level gravel deposits in the vicinity of Philadelphia. Bascom (1924) included high-level gravels (390–480 feet) in Pennsylvania, Delaware, and in Cecil County, Maryland. The term was then extended to the vicinity of Washington and Burtonsville.

Distribution

There is a small patch of Bryn Mawr Gravel at Tacoma Park. A second area caps the hills between Potomac River and Rock Creek within the District of Columbia. The highest portion of this area is about 410 feet above sea level; the elevation of the base of the gravel ranges between 340 and 400 feet. These remnants are thought to be part of a funnel-shaped fan related to the course of the Potomac River.

A second fan occurs southwest of Patuxent River at Burtonsville and extends toward Laurel and the Prince George's County line. This gravel bed, about 50 feet thick, straddles the contact between the Patuxent Formation and the crystalline rocks.

Description

The Bryn Mawr Gravel consists of coarse, poorly sorted pebbles in red sand and silt. It is distinguished from the Brandywine Formation by its bright red color and its higher altitude. In the District of Columbia the gravel occurs at an elevation between 350 and 410 feet, and at Burtonsville between 450 and 500 feet.

Origin and Age

This gravel is always near the mouth of a river where it empties from the Piedmont onto the Coastal Plain. It seems to have been deposited in alluvial fans where the current slowed below the Fall Line. The Bryn Mawr Gravel may have formed a fan similar to the Brandywine, which has been studied by Schlee and is described below.

The gravel must have been deposited at the level of the river near which it is now found. In northwest Washington the Potomac River is at least 280 feet below the lower edge of the gravel and 340 feet below its top. This very substantial drop probably means a considerable lowering of the river level after gravel deposition and age for the gravel. It is customarily assigned to the Pliocene (?) but it may be Miocene or older (Cooke, 1952, p. 38).

BRANDYWINE GRAVEL

Name

The name was proposed by Clark (1915) for gravel, sand, and loam which occurs near Brandywine in Prince George's County. This gravel was included in what was later identified as Bryn Mawr Gravel by Bascom (1924) who then restricted the term Brandywine to gravels at a lower level. Cooke (1930, p. 582; 1931, p. 505) set the landward limit of the Brandywine terrace at 265 to 270 feet above sea level.

Description

The Brandywine Gravel consists mostly of well-rounded, polished pebbles of quartzite, sandstone, chert, and fairly clean quartz sand (Cooke, 1952, p. 40).

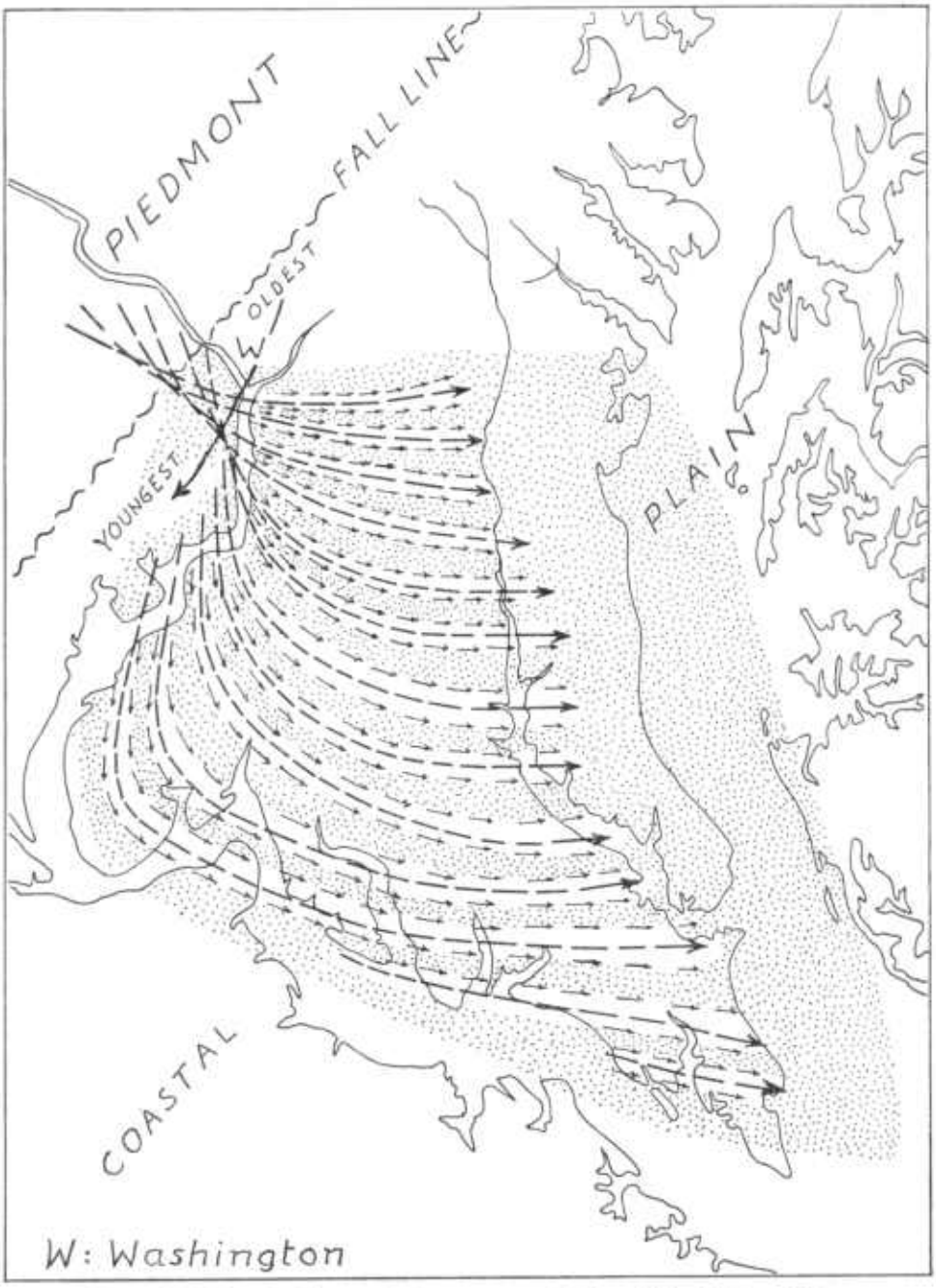

FIGURE 9. Schematic diagram of paleocurrent trends After Schlee (1957, p. 1397).

In 1955 Hack divided the Brandywine into two members, an upper loam and a lower gravel. The upper member has been eroded from the higher areas, and only the gravel is preserved. It consists of pale-dark yellowish-orange to yellowish-brown, poorly sorted gravel and sandy gravel. Bedding is visible, and the units range from 6 feet to 2–3 inches. Cross-bedding is present at many exposures.

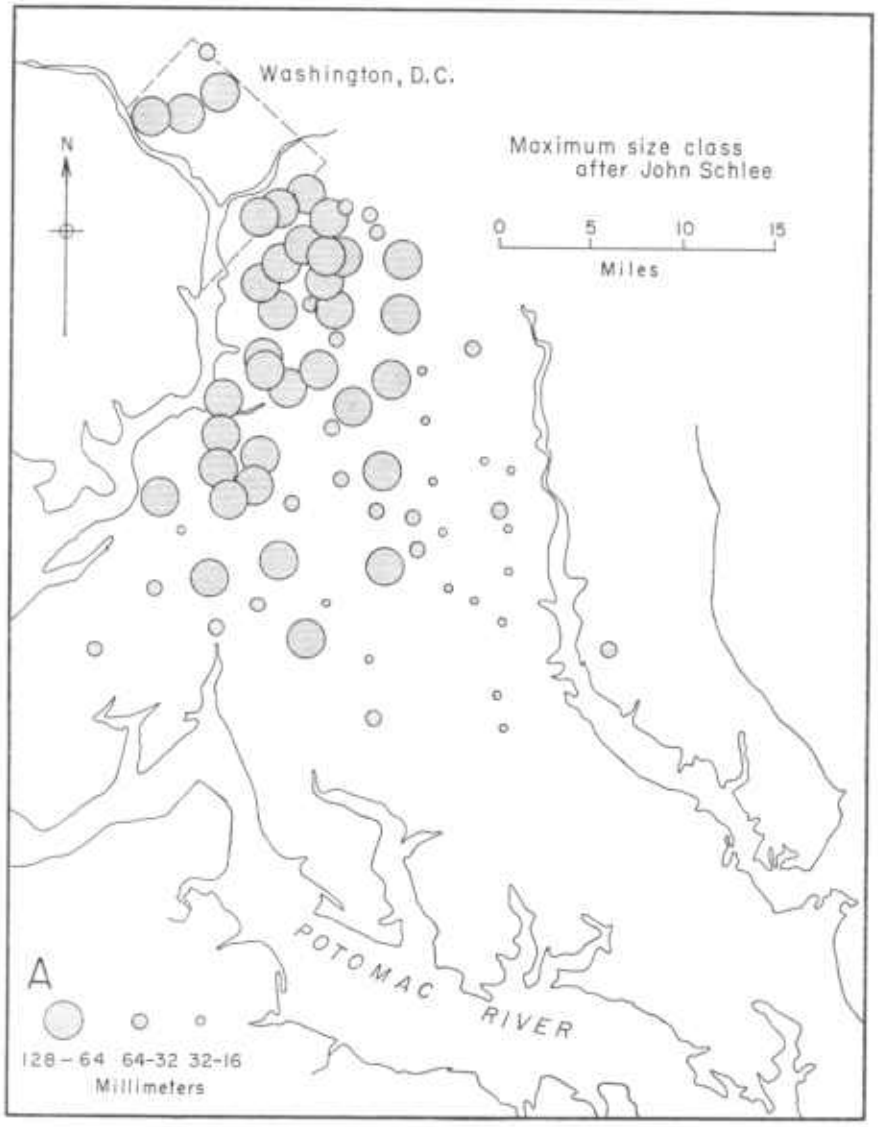

FIGURE 10. Regional distribution of gravel sizes for samples from Upland Deposits. After Schlee (1957, p. 1385).

A thorough study of this gravel and its depositional history has been undertaken by Schlee (1956; 1957).

Thickness

The Brandywine Gravel is probably 25–30 feet thick.

Origin

The gravel was "probably deposited by a graded ancestral eastward-flowing Potomac River which, in the process of lateral corrasion, spread a veneer of channel gravel and flood-plain silt. Later side- and downcutting, combined with a gradual southeastward migration of the river, resulted in the present thin gravel sheet" (Schlee, 1957, p. 1371).

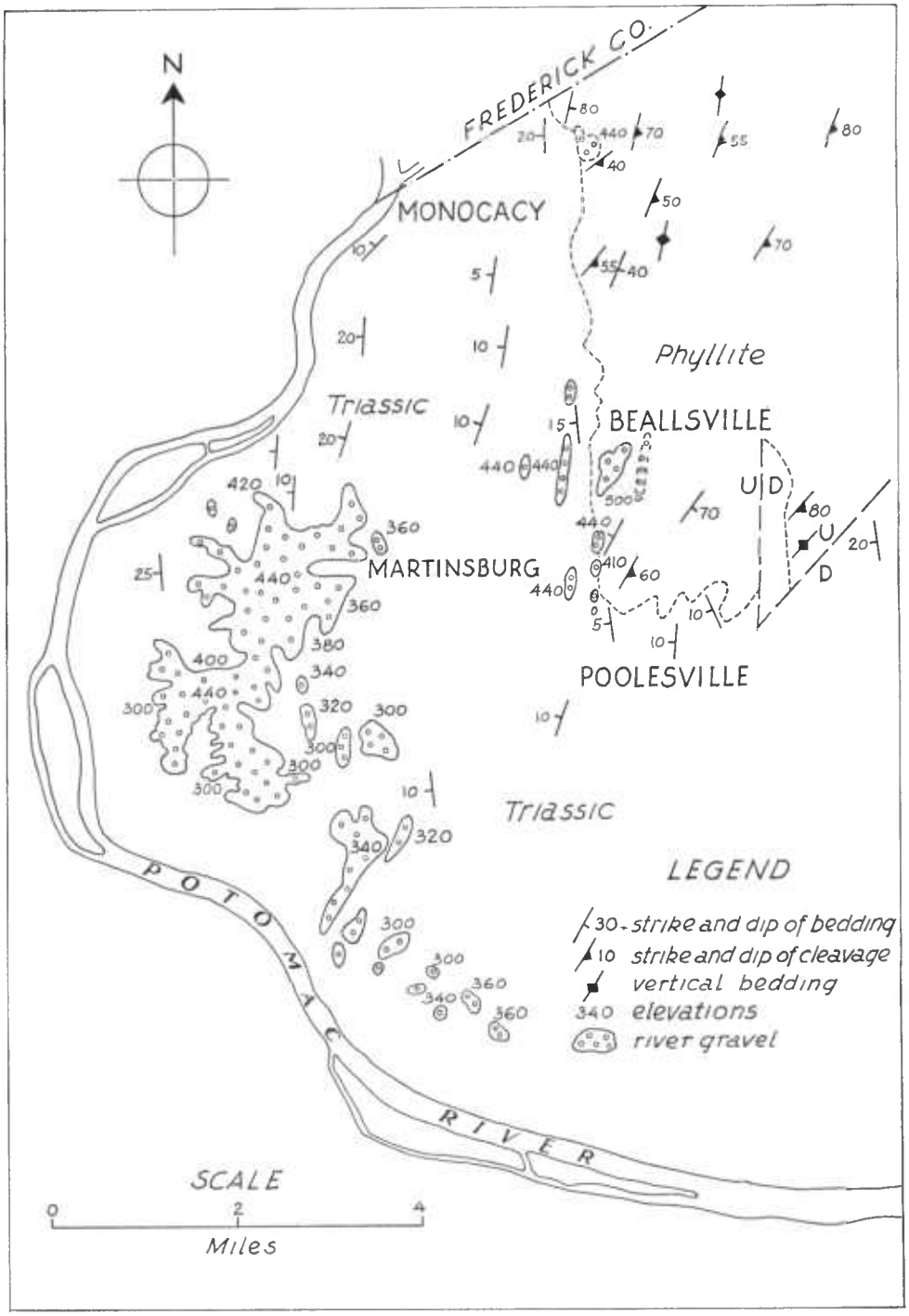

FIGURE 11. Terrace gravels in Western Montgomery County

Schlee's conclusions are based on a study of the distribution of gravel size and pebble orientation with respect to transport in the original Potomac River.

Measurements of the gravel demonstrate that pebble sizes decrease from the mouth of the River at Washington across the peninsula, showing that the large components were dropped at the mouth, whereas the smaller ones were transported up to 30 miles farther out in Charles and St. Mary's Counties (Schlee, 1957, p. 1385).

Figure 9 after Schlee, 1957, Fig. 21, is a schematic diagram showing the trend of paleocurrents during deposition of the Brandywine gravels. The Brandywine is a fine example of a fan which originates at Washington and spreads out over the area between the Potomac River and the Chesapeake Bay.

QUATERNARY SYSTEM

PLEISTOCENE SERIES

General Remarks

No attempt has been made to correlate the deposits along the streams and rivers with those below Washington where fluctuations of sea level during the Pleistocene time have been determined and described by Cooke (1952, p. 42). All of Montgomery County is above the Fall Line and upstream of the Brandywine gravel fan, and it is very doubtful that correlation can be made.

Two kinds of deposits have been distinguished on the geologic maps: terrace gravels near the Potomac River and stream-bottom deposits near all streams.

Terrace Gravels

In western Montgomery County the Potomac River has left gravels on an older erosion surface above Triassic rocks. Although not yet thoroughly studied, the gravels are very similar to Recent gravels, contain the Paleozoic rocks of the folded Appalachians, and are well rounded.

They occur as far as 3–4 miles from the present river and up to 300 feet above its present level. They are mostly above the Triassic boundary and occur on schists near Beallsville (See: Figure 11).

Figure 11, a map of the west end of Montgomery County where gravel occurs between elevations of 300 to 500 feet, suggests that the old Potomac River flowed over the old surface as far as the phyllite near Beallsville at an elevation of about 500 feet, or about 300 feet above its present level.

The little-consolidated and noncrystalline Triassic rocks were more readily eroded, and the river found its present bed gradually leaving gravels at higher elevations as it moved toward its present course. Available evidence is insufficient to distinguish terrace levels and erosion stages.

Terrace gravels were not mapped in the Patuxent and Patapsco River valleys.

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THE CRYSTALLINE ROCKS OF HOWARD AND MONTGOMERY COUNTIES

BY

CLIFFORD A. HOPSON

INTRODUCTION

Montgomery and Howard Counties together nearly span the Maryland Piedmont, and provide a representative showing of Piedmont metamorphic and plutonic rocks. Above the ancient Precambrian basement, exposed in Howard County and in the Baltimore area, there lies a nearly complete section of the metamorphosed stratified rocks—the Glenarm Series to the east and the wide expanse of dominantly phyllitic rocks, in contact with Lower Cambrian strata, to the west. Across this section the grade of regional metamorphism changes progressively, from the chlorite zone in the west to the kyanite and sillimanite zones in the east, where the beginning of anatexis is locally evident. There is also a varied assemblage of plutonic rocks, which include ultramafics and gabbro as well as a wide range of granitic rocks. The earlier plutonic rocks are in gneissic and schistose intrusive masses, while the youngest granitic rocks are in relatively undeformed late-kinematic plutons. Thus the terrane is highly varied and poses many interesting problems—stratigraphic, structural, and petrologic.

In modern work in the Maryland Piedmont much interest has centered on the deformation and metamorphism. Here more emphasis is placed on the less known premetamorphic history of the stratified rocks, and on the petrology of the plutonic rocks. Also, a special attempt is made to integrate geologic evidence with the radiogenic mineral ages recently obtained in the Baltimore-Washington area (Tilton *et al.*, 1958; 1959; 1960). Thus it is hoped to gain more information about development of the Piedmont geosyncline, and about the age and origin of its metamorphic and plutonic rocks.

In the following sections the rocks are discussed in approximate order of decreasing age: Baltimore Gneiss, Glenarm Series, crystalline schists of the western Piedmont, and the plutonic rocks. Following that is a summary and discussion of the radiogenic dating and geologic age problems.

Neither the rocks nor the problems conform to county lines, and no attempt is made to stay strictly within Howard and Montgomery Counties. Much of my field work has been done in the Baltimore area, where the gneiss domes, the gabbro complex, and some other rocks are more advantageously studied. Thus the work in this area and observations from other places outside of Howard and Montgomery Counties are included in the report.

The published geologic maps of Montgomery County (Cloos and Cooke, 1953), Howard County (Cloos and Broedel, 1940), and Baltimore County (Knopf and Jonas, 1925) form the basis for the following account of the geology of the crystalline rocks. My own field work through the years 1957–1963 involved no new geologic mapping; except for small details my observations corroborated the mapping of Cloos, Knopf, and Jonas. Needless to say, the main conclusions from the present study are dependant on the hard core of basic information that their maps give.

Ernst Cloos introduced me to Appalachian geology, provided the opportunity for my

participation in the present study, and collaborated in some phases of my part of the report. His enthusiasm and encouragement all along the line are deeply appreciated.

THE BALTIMORE GNEISS

INTRODUCTION

The Baltimore Gneiss, the oldest rock in the Maryland Piedmont, crops out in the cores of seven anticlinal domes and in a narrow strip along the Coastal Plain margin near Baltimore (Fig. 12). The gneiss in the domes lies unconformably beneath metasedimentary rocks of the Glenarm Series and is cut by younger granitic rocks. The Baltimore Gneiss was named by George Huntington Williams (1892a) for well stratified gneisses cropping out beneath Coastal Plain sediments along Jones Falls and Gwynns Falls in Baltimore City. In the gneiss domes the formation is a complex assemblage of quartzo-feldspathic gneisses, amphibolite, migmatite, and gneissic granitic rocks. The name has also been used for similar gneissic rocks in the Piedmont of southeastern Pennsylvania (Bascom, 1905; Bascom and Stose, 1932; Gray, 1960). In its present usage, therefore, "Baltimore Gneiss" includes a wide variety of rocks, possibly of different ages and origin, which lie beneath the Glenarm Series. It is synonomous with "basement complex".

The gneiss is Precambrian in the Baltimore domes. Radiogenic dating (Tilton *et al.*, 1958; 1960) indicates that the rock went through a period of crystallization about 1000–1100 million years ago.

OCCURRENCE AND GENERAL STRUCTURAL RELATIONS

The gneiss domes are situated almost entirely within Howard and Baltimore Counties (Fig. 12). They form protrusions along the crest of the major anticlinorium which passes through Baltimore and Washington, D. C. (Fig. 47). Baltimore lies at the apex of this doubly-plunging anticlinorium. To the southwest and northeast of Baltimore, along the regional strike, the gneiss disappears beneath younger formations.

Mathews (1907) was first to note the general form of the gneiss domes, and Broedel (1937) studied their structures in detail. Some of the domes are roughly circular, suggesting vertical uplift, while others are elongate or anticlinal. For simplicity they are all called domes. Because of faulting the Clarksville dome in Howard County appears narrow and elongate. Before faulting the dome probably had the shape of a pear, broad and rounded at the southern end and tapering to a narrow neck on the north. The Towson-Chattolanee-Woodstock-Mayfield-Clarksville string of domes are all protrusions along the crest of a single curving anticline, while the Phoenix dome is a separate uplift to the north (Fig. 12). The Texas dome is a bulge on the saddle between the Phoenix and Towson domes. The limbs of the domes are generally steep, and some are overturned. Overturning is mostly to the east in Howard County and to the south in Baltimore County. The general relationships are shown effectively by Broedel (1937, Pl. 27).

Rimming the gneiss domes is quartzite and feldspathic mica schist of the Setters Formation, which rests unconformably on the gneiss. Above it are Cockeysville marble and calc-silicate rocks, and Wissahickon pelitic schist and psammitic granulite. Evidence for the unconformity is discussed at the beginning of the section on the Glenarm Series. Locally the Setters is missing, and one of the higher formations rests against the gneiss. This is due to faulting along the eastern edge of the Clarksville dome, but in other cases its absence could be better explained by non-deposition or by pre-Wissa-

FIGURE 12. Generalized geologic map of the gneiss domes in Howard and Baltimore Counties. Geology taken from the Maryland Geological Survey maps of Howard, Baltimore, and Carroll Counties and from Broedel (1937).

hickon erosion. This problem is discussed in the sections on Setters and Cockeysville Formations.

Eskola (1949) first called attention to the problem of mantled gneiss domes. They are unique structures, having features which suggest both that the gneiss is the older basement rock, mantled by younger strata, and that the gneiss is younger, intrusive into its mantle. Evidence that the gneiss is older is as follows:

(1) It is overlain by a succession of sedimentary strata, with the same formation at the base. If the gneiss were intrusive, why should it always stop at the same horizon?

(2) In some of the Finnish domes the basal formation is conglomerate, carrying pebbles of the underlying gneiss.

Favoring the second alternative, that the gneiss is intrusive, is the following:

(1) The banding and foliation in the outer part of the gneiss is parallel to the base of the mantling strata. In Eskola's (1949, p. 463) words: "It was difficult to understand how, if the contact surface had once been the earth's surface, the banding of the gneiss could be parallel to it and invariably concordant with the superimposed strata".

(2) In some places marginal facies of the gneiss invade the mantle as dikes.

Eskola brilliantly resolved these contradictions by showing, for the Finnish domes, that the gneiss was indeed the older basement rock, but that during later orogeny it was reactivated by migmatization, becoming mobile and punching upward as domical intrusions into its own mantle. During a brief visit to Maryland in 1946 Eskola visited the Baltimore gneiss domes and concluded that they had a history similar to that of the Finnish mantled domes (1949, p. 470).

Domes of Baltimore Gneiss concordantly mantled by Setters and Cockeysville Formations are only found west and north of the Baltimore Gabbro Complex (Fig. 12). Gneiss also occurs southeast of the gabbro, but crops out in a narrow northeast trending strip between younger intrusive Relay Quartz Diorite and overlapping Coastal Plain sediments, so that domal form is not evident. In southwestern Baltimore this strip of gneiss is concordantly overlain by a narrow wedge of garnet-oligoclase-quartz-mica schist, shown as Setters Formation on the Baltimore County geological map. This rock is not Setters, however, as is evident from its quite different mineralogy and composition; it is like Wissahickon schist, and is shown as such on Figure 12. Thus the gneiss in southwestern Baltimore—its type locality—does not lie directly beneath the lowest Glenarm formations, nor is there any certainty that it forms a mantled dome.

In Harford County a northeast trending belt of rocks passing east of Belair is shown as Baltimore Gneiss on the Maryland Geological Survey maps (Mathews and Johannsen, 1904; Mathews, 1933). These rocks do not occur beneath Setters and Cockeysville Formations either, and for the most part they are lithologically different from the gneiss in the mantled domes. Some of this rock closely resembles the stratified paragneiss in southwestern Baltimore, but there is also much mica schist (see Hershey, 1937, Pl. XX) like the Wissahickon. Probably the well-stratified gneiss in Harford County and in southwestern Baltimore are the same, for they lie along strike and resemble one another closely. More work is needed to test this possibility, however, and to establish their relationship to the gneiss in the mantled domes northwest of the gabbro complex.

LITHOLOGY

General Remarks

Mineralogically the Baltimore Gneiss is simple and monotonous. About 95 per cent of it is biotite-quartz-feldspar gneiss, with occasional minor muscovite or garnet. Biotite-hornblende gneiss and amphibolite is widespread but subordinate.

Texturally and structurally, however, the gneiss is extremely varied. Granitic gneiss, veined gneiss, augen gneiss, and banded gneiss are common distinctive varieties, but all gradations occur, and in places they are complexly intermingled. Moreover, these varieties may be intimately mixed with alaskite or pegmatite, forming migmatite. Well-stratified paragneiss is found only in southwestern Baltimore, peeking out from beneath the Coastal Plain cover. Some of the banded gneiss and migmatite in the other domes appear to have originally been similar to the paragneiss, but now they are greatly deformed and granitized.

Paragneiss

At its type locality along Jones Falls and Gwynns Falls in Baltimore City the Baltimore Gneiss consists of well-bedded granulite, gneiss, and amphibolite that is less strongly transformed than the complex gneissic and migmatitic rocks of the other domes. When viewing these rocks one is impressed by the perfection of their stratification; from a distance one might at first mistake them for unmetamorphosed sedimentary rocks. The individual layers are in sharp contact, in contrast to the irregular or gradational contacts, sometimes blurred by feldspathization, found in the banded and migmatitic gneisses. Moreover, many of the layers are straight and continuous for long distances, whereas in gneisses of the other domes the layers generally pinch out, fray out, or fade out within a few feet or even a few inches. In the Campbell Corporation quarry, along Gwynns Falls at W. Baltimore Street, beds only a few inches thick can be traced continuously for many tens of feet along the quarry wall. The persistence and perfection of the stratification led Knopf and Jonas (1929, p. 147) to conclude that these rocks were of sedimentary origin. I concur that the layering is primary. Metamorphic differentiation cannot account for the striking difference in composition between many adjoining layers, for the sharpness of their contacts, or for their uniform thickness and great lateral continuity.

The strata, although tilted in places up to 50°, show relatively little deformation. They lack the complicated drag folds, ptygmatic folds, and the irregular buckling, stretching, and swirling that the banded gneisses and migmatites in the other domes commonly display. Foliation is invariably parallel to bedding, and linear structures are weak or lacking. Hornblende prisms and biotite segregations commonly show no preferred orientation when viewed on the bedding surface (*ab* plane). One gains the impression that these rocks underwent very little penetrative deformation, and perhaps acquired their foliation by static mimetic recrystallization parallel to the bedding.

Most common is fine-grained biotite-quartz-oligoclase gneiss and granulite. Some contains minor amounts of microcline and garnet. The term granulite is used here in the textural sense, to describe quartzo-feldspathic metamorphic rocks that are relatively fine-grained and that are weak or lacking in gneissic structure (Harker, 1939, p. 246). It does not connote metamorphic facies.

The gneiss and granulite occur in beds from 1 inch to more than 20 feet thick. Interstratified with them are thin beds of hornblende-quartz-oligoclase granulite, with or without biotite, microcline, or garnet, and amphibolite. Even the most delicately stratified rocks are thoroughly recrystallized, with no vestige of original texture. Modal analyses of some of the typical paragneisses are given in Table 8.

Granite pegmatite appears locally in the stratified gneiss and granulite, chiefly as discordant dikes. Some of them are sharp-edged intrusive dilation dikes; others have gradational contacts and retain ghostlike traces of discordant wall-rock structure passing through them, indicating replacement. Both types were primarily controlled by

fractures, as shown by their straight, tabular form. A few pegmatites form sharply bounded pods, with the bedding deflected around them. The pegmatites do not intimately vein or permeate the gneiss, but form discrete, generally discordant bodies spaced far apart, and they are absent over wide areas. They are not indicative of pervasive granitization. The high-grade metamorphism of the paragneiss was evidently isochemical.

The well-preserved bedding indicates a supracrustal origin for the gneiss sequence in southwestern Baltimore. The rocks may have been ordinary sandstones and shales or bedding tuffaceous rocks. Their chemical composition supports the second alternative. The possibility that they were lava flows is rejected because of their thin stratification.

TABLE 8

COMPOSITION OF THE PARAGNEISS

Baltimore Gneiss

	1	2	3	4	5	6
Modal Analyses						
Quartz	37.1	38.4	46.5	23.0	30.3	1.8
Microcline	9.9	1.6	—	4.0	—	—
Plagioclase	48.2	53.0	48.5	53.0	48.3	50.0
Biotite	3.9	6.0	3.7	2.4	0.5	2.1
Hornblende	0.1	t	—	16.4	18.2	46.0
Almandite	0.4	—	—	—	—	—
Epidote	0.1	0.3	0.3	0.4	—	—
Allanite	—	—	0.1	t	—	—
Apatite	t	t	t	0.1	0.3	0.1
Sphene	0.2	t	0.1	0.7	0.7	—
Zircon	t	0.1	0.1	t	—	—
Magnetite	0.1	0.6	0.7	—	1.7	—
Total	100.0	100.0	100.0	100.0	100.0	100.0
Points	1320	1271	1079	1212	1283	1349
Plagio. Comp.	An_{24}	An_{25}	An_{23}	An_{22}	An_{25}	An_{24}

Chemical Composition, Calculated from the Modes

	1	2	3	4	5	6	7
SiO_2	74.9	74.1	77.8	67.0	67.5	55.1	74.66
TiO_2	0.3	0.2	0.2	0.5	0.4	0.4	0.25
Al_2O_3	14.0	13.9	12.0	14.8	12.7	15.2	14.02
Fe_2O_3	0.3	1.0	1.1	0.8	2.8	1.8	0.37
FeO	1.3	1.9	1.4	2.5	3.1	5.3	1.44
MgO	0.3	0.4	0.3	2.9	3.1	7.5	0.17
CaO	2.5	2.7	2.4	5.2	5.4	8.6	2.08
Na_2O	4.3	4.7	4.3	4.7	4.3	4.6	4.20
K_2O	2.0	0.9	0.4	1.0	0.2	0.3	2.01
H_2O	0.1	0.2	0.1	0.5	0.4	1.1	0.77
P_2O_5	t	t	t	0.1	0.1	0.1	0.09
Total	100.0	100.0	100.0	100.0	100.0	100.0	100.06

TABLE 8—*Continued*
CIPW Norms

	1	2	3	4	5	6	7
Quartz	35.9	35.8	44.8	20.5	27.2	—	37.3
Orthoclase	11.7	5.6	2.8	6.1	1.1	1.7	11.7
Albite	36.2	39.8	36.2	40.4	36.2	38.8	35.6
Anorthite	12.5	13.3	12.0	15.6	15.0	20.0	9.7
Corundum	0.2	0.2	0.1	—	—	—	1.1
Diopside	—	—	—	7.6	8.7	17.7	—
Enstatite	0.8	1.0	0.8	4.9	4.7	10.7	0.4
Ferrosilite	1.6	2.2	1.2	2.1	1.7	4.5	2.0
Olivine	—	—	—	—	—	2.0	—
Ilmenite	0.6	0.5	0.5	0.9	0.6	0.8	0.5
Magnetite	0.5	1.4	1.6	1.2	4.2	2.6	0.5
Apatite	—	—	—	0.3	0.3	0.3	0.2
Total	100.0	99.8	100.0	99.6	99.7	99.1	99.0
Plagio. Comp.	An_{25}	An_{25}	An_{25}	An_{28}	An_{29}	An_{31}	An_{21}

(1) Gneiss (BD1-1).
(2) Gneiss (BD1-2).
(3) Granulite (BD1-4).
(4) Granulite (BD2-1).
(5) Granulite (BD1-3).
(6) Amphibolite (BD2-3).
(7) Chemically analyzed biotite gneiss. Schwinds quarry, 29th and Remington Ave., Baltimore, Md. Penniman & Browne, analysts. (Knopf and Jonas, 1929, p. 149, col. 1).
Specimens 1–6 from the Campbell Corporation quarry, Gwynns Falls at West Baltimore Street, Baltimore.

The chemical compositions of several typical rocks from the paragneiss sequence are given in Table 8. All but no. 7 were calculated from the modes. Plagioclase compositions used for the calculations were determined optically. The biotite and hornblende compositions are the analyzed Baltimore Gneiss biotite from Table 14 and the hornblende from Table 34, col. 2. The calculated values for FeO, Fe_2O_3, MgO, and TiO_2 in the rocks are probably the least accurate; however, these errors do not significantly affect the position of points plotted in Figure 13, where only salic constituents are involved.

An outstanding feature is the high Na/K ratio of the paragneisses. This and other chemical characteristics are brought out by plotting (Fig. 13) their normative orthoclase, plagioclase, quartz, and corundum (from Table 8), thus facilitating comparison with a number of possible parent rocks. These four salic minerals constitute more than 80 per cent of most of the rocks plotted; only subordinate femic minerals are omitted. This plot is a convenient way to compare the relative amounts of K, Na, Al, and Si, which are present in different proportions in various rocks of sedimentary and igneous origin.

Figure 13 shows that the paragneiss has a much higher Na/K ratio (indicated by normative plagioclase/orthoclase) than the average shale and lacks its excess alumina (normative corundum). The gneiss also has much less excess silica than the more quartzose varieties of sandstone. Only the average graywacke closely approaches the chemical composition of the paragneiss. Graywacke, however, does not ideally fulfill the requirements for the parent rock either, for the following reasons:

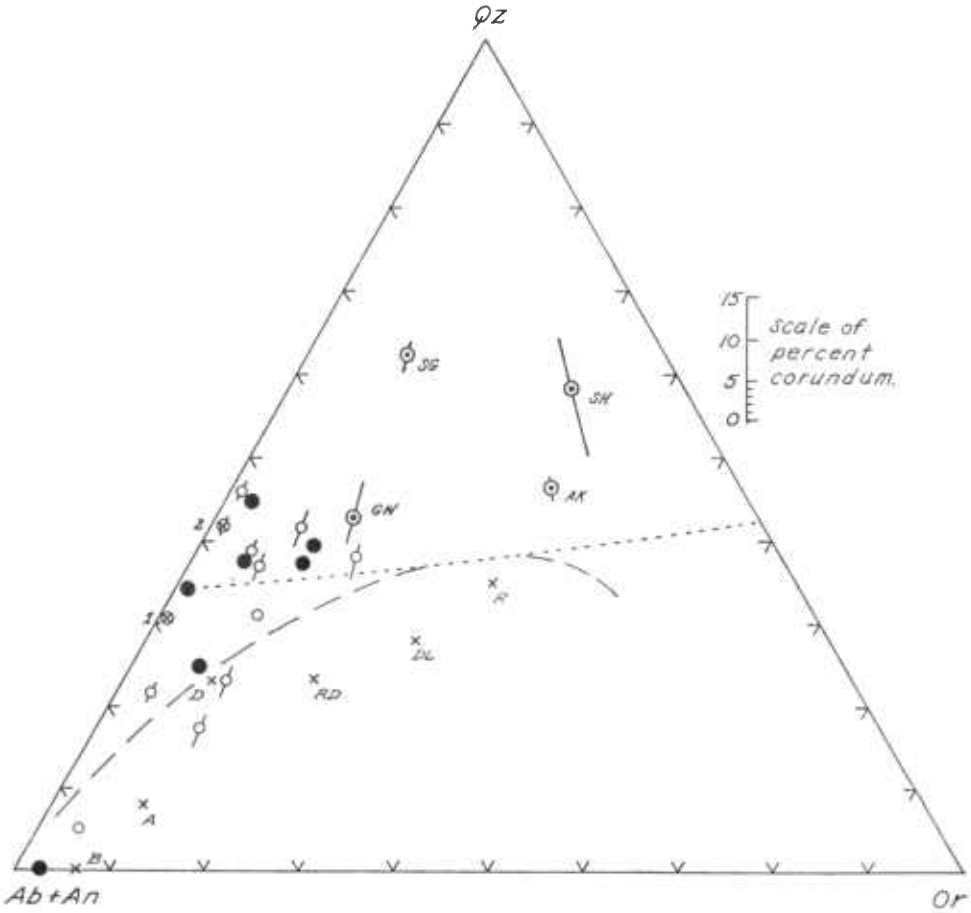


FIGURE 13. Plot comparing the normative compositions of Baltimore paragneiss (solid circles) with those of the following possible parent rocks: average shale (SH), average arkose (AK), average subgraywacke (SG), average graywacke (GW), a series of average calc-alkaline lavas (R, rhyolite; DL, dellenite; RD, rhyodacite; D, dacite; A, andesite; B, tholeiitic basalt), 10 altered marine volcanic siltstones and sandstones (small open circles), and 2 quartz keratophyres formed by alteration of volcanic ash (circles with ×). Normative quartz (Qz), orthoclase (Or), and plagioclase (Ab + An) are plotted on the triangle, and the line through each point marks normative corundum (percentage given by the vertical distance between the ends of the line). The dotted line separates the fields of SiO_2 and alkali feldspar on the liquidus of the "dry" system AB-OR-SiO_2; unaltered volcanic rocks plot in the feldspar field (Bowen, 1937), whereas shales and nonvolcanic sandstones (except some arkoses) generally plot in the SiO_2 field. The long-dashed curve approximately separates the region where calc-alkaline lavas and unaltered volcanic clastic rocks plot (below the curve) from that in which albitized and zeolitized pyroclastic and epiclastic volcanic rocks plot. The Baltimore paragneisses fall with the latter group.

The Baltimore paragneisses are those shown in Table 8. The average shale, arkose, subgraywacke, and graywacke are from Pettijohn (1957, p. 344, col. A; p. 324, col. F; p. 319, cols. A-D; p. 307, col. A). The average lavas are from Nockolds (1954, p. 1012, col. II, p. 1014, cols. II, and IV, p. 1015, col. VI, p 1019, col. II, and p. 1021, col. VII). The quartz keratophyres are #1 from Dickenson (1962, p. 263, col. 7) and #2 from Albers (cited in Dickenson, p. 263, col. 10). The volcanic sandstones and siltstones, all from the Eocene Ohanapecosh Formation, Washington, are albitized and zeolitized marine deposits consisting of rhyodacitic to andesitic pyroclastic debris (unpublished chemical analyses provided by R. S. Fiske; H. Haramura, analyst).

(1) Graywackes, with their argillaceous matrix, normally have excess alumina but the paragneisses have almost none (Fig. 13; Table 8, *see* normative corundum values).

(2) Graywackes are usually found interbedded or closely associated with shale, but the paragneiss sequence appears to have lacked shale. Instead, the paragneiss contains numerous thin interbeds of mafic rock (amphibolite), which is atypical for graywacke sequences.

(3) Graded beds and other relict structures common to graywackes are not found in the paragneisses, although they are well preserved in Wissahickon metagraywackes that have suffered equally strong deformation and metamorphism (Pls. 11–13).

Turning to volcanic rocks, there again is a lack of correspondance to the paragneiss. In Figure 13 the normative composition of the gneiss is compared with that of average calcalkali volcanic rocks. The paragneisses show more normative albite and quartz than volcanics of intermediate to silicic composition, reflecting a higher Na/K ratio and more excess silica. The long-dashed line approximately separates the two groups. At the base of the triangle an amphibolite from the paragneiss sequence plots near basalt.

There is a class of altered marine volcanic sediments, however, whose chemical composition does closely match the Baltimore paragneiss. These are the albitized and zeolitized volcanic clastic rocks of intermediate to silicic composition. Examples of such rocks are the two quartz keratophyres plotted in Figure 13, both derived from albitized tuff. Other examples are taken from the Upper Eocene Ohanapecosh Formation, (Washington), an altered submarine deposit of andesitic to rhyodacitic pyroclastic debris about 15,000 feet thick (Fiske, Hopson, and Waters, 1963). Analyses of 10 Ohanapecosh volcanic sandstones and siltstones, plotted in Figure 13, completely overlap the field of the paragneiss. Such rocks are by no means rare. Zeolitized and albitized volcanic epiclastic and pyroclastic rocks are widespread and voluminous in the Cascade Mountains and Coast Range in Oregon and Washington, and they also form thick sections in New Zealand (Coombs, 1954), Japan (R. S. Fiske, personal communication), and other parts of the Circumpacific belt. Their alteration, and in some cases metasomatism, was effected by resurgent connate waters during diagenesis and incipient metamorphism accompanying deep burial (Fiske, Hopson, and Waters, 1963; Dickenson, 1962; Waters, 1955; Coombs, 1954).

Such marine volcanic sediments commonly resemble the Baltimore paragneiss in structure as well as composition, forming massive to thin-bedded deposits with interstratified mafic material. For instance, tuffaceous shale, siltstone, and rhyolitic to andesitic sandstone form thin beds and thick massive units interlayered with palagonite tuff and tuffaceous basaltic sandstone in the thick marine sequences of Eocene-Oligocene age in western Oregon and Washington (Snavely and Wagner, 1963; Snavely *et al.*, 1951; Warren *et al.*, 1945; Wells and Waters, 1935), and in the Miocene of Japan (Matsuda and Mizuno, 1955; R. S. Fiske, personal communication). Such rocks, if they were metamorphosed to the amphibolite facies, would closely resemble the gneiss, granulite, and amphibolite in southwestern Baltimore both in structure and composition.

The Baltimore paragneiss, therefore, probably represents a thick sequence of rhyodacitic to basaltic sediments, which may have been albitized prior to high-grade metamorphism. A second but far less likely alternative is that the paragneisses came from graywacke.

Banded Gneiss

Interlayered dark and light gneiss, resembling the stratified paragneiss, is widespread although not abundant. It is best displayed in the Phoenix, Towson, and Woodstock domes, but examples are found in the others as well. These banded rocks may have originally been paragneiss like that in southwestern Baltimore, but if so they have been coarsely recrystallized, granitized, and strongly deformed. In the Baltimore County report this rock is described as ribbon gneiss (Knopf and Jonas, 1929, p. 143).

The dark layers are hornblende- and biotite-rich gneisses and amphibolite. Table 9 lists modal analyses of some typical examples. These dark layers are not so sharp and continuous as are those in the paragneiss. Many of the contacts are fuzzy, and the thinner layers fray out or fade out within short distances along their strike or dip. The dark banded gneiss layers also appear to be richer in K_2O; many have more biotite than the paragneisses of comparable color index, and more of them contain microcline. (*Compare* Table 8, cols. 3–5, with Table 9.) In many cases, nevertheless, these differences are not great.

The contrast between banded gneiss and paragneiss is much stronger for the light-colored layers. The light banded gneiss layers are often coarser-grained or even quite coarsely porphyroblastic. Metamorphic differentiation has commonly expelled the

TABLE 9

MODAL ANALYSES OF HORNBLENDE GNEISS AND AMPHIBOLITE

Baltimore Gneiss

	1	2	3	4	5	6	7	8
Quartz	0.3	24.2	27.9	1.5	16.8	5.4	10.3	8.3
Plagioclase	45.1	32.1	32.7	34.0	38.5	41.5	25.8	35.2
Microcline	4.7	—	—	—	2.9	2.0	—	0.8
Diopside	9.1	—	—	—	—	—	—	—
Hornblende	36.5	20.7	5.4	43.3	8.6	11.1	33.9	5.6
Biotite	—	20.0	27.7	19.7	26.1	30.1	21.0	38.8
Chlorite	—	0.1	—	0.5	—	—	—	—
Garnet	—	—	1.1	—	—	—	—	—
Clinozoisite	2.4*	—	1.1	—	1.8*	1.6	3.7	6.9
Allanite	—	—	0.1	—	0.7	—	—	0.1
Apatite	—	0.2	0.4	0.5	0.9	2.5	0.9	0.3
Sphene	1.9	1.8	2.1	—	2.6	4.6	4.0	0.7
Zircon	—	0.1	t	—	t	t	—	—
Magnetite	—	0.4	1.5	0.5	1.1	0.5	0.4	—
Calcite	—	0.4	—	—	—	0.7	—	3.3
Total	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0
Points	1302	1792	1231	1132	544	1103	1091	1113

(1) Diopside amphibolite (B-65). Woodstock dome, near Marriottsville.
(2) Biotite-hornblende gneiss (H28-1). Woodstock dome, near Marriottsville.
(3) Garnet-hornblende-biotite gneiss (B-64). Woodstock dome, near Marriottsville.
(4) Biotite amphibolite (TD10-2). Towson dome, near Cromwell Bridge.
(5) Hornblende-biotite gneiss (TD-2). Texas dome, near Merryman Branch.
(6) Hornblende-biotite gneiss (PD1-4). Phoenix dome, west of Verona.
(7) Biotite-hornblende gneiss (PD3-3). Phoenix dome, north of Glencoe.
(8) Hornblende-clinozoisite-biotite gneiss (PD6-3). Phoenix dome, near Western Run.
* Epidote.

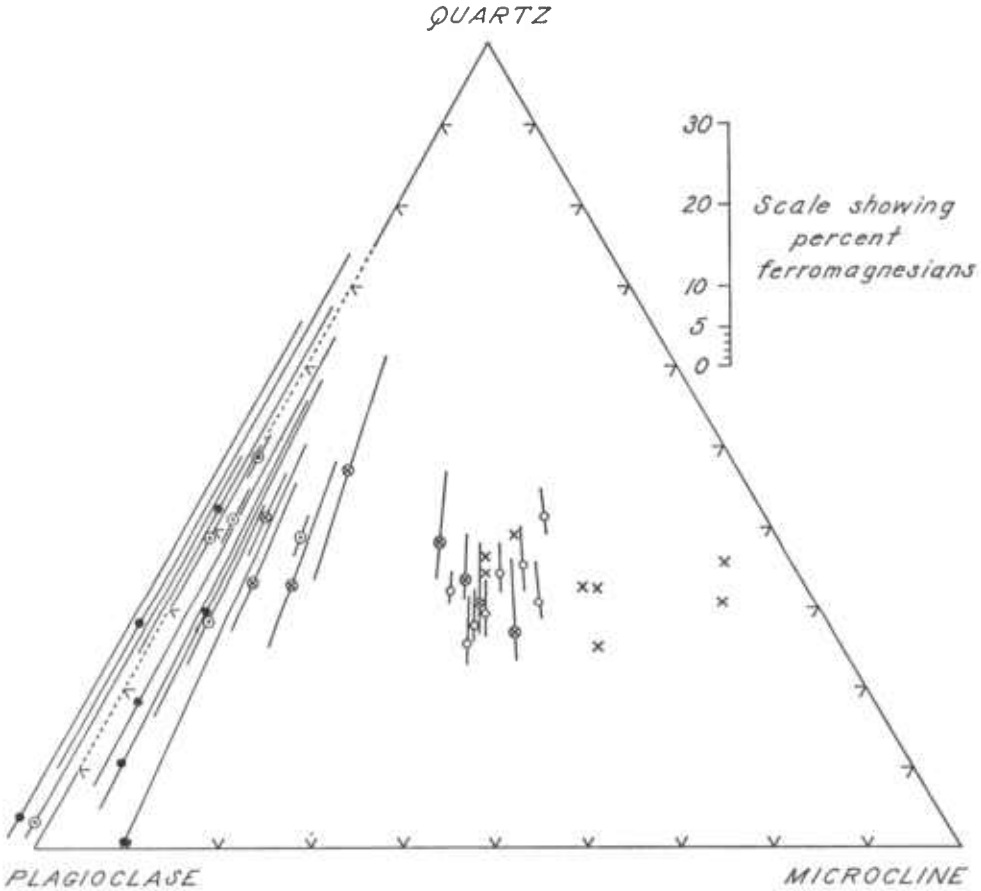


FIGURE 14. Modal compositions of different varieties of Baltimore Gneiss. Solid circles, hornblende gneiss and amphibolite; circles with dot, paragneiss; circles with ×, biotitic portion of migmatite and veined gneiss; ×, light-colored layers and veins in migmatite and veined gneiss; small open circles; granitic gneiss. Top of line through each point marks percentages of microcline + plagioclase; bottom of line marks percentage of quartz. Length of line marks percentages of ferromagnesian + accessory minerals. Some points that plot on the left edge are shifted just outside the triangle to show their ferromagnesian lines separately.

biotite, concentrating it into clots within the layers or rimming their margins. The biggest contrast, however, is in composition. The light-colored banded gneiss layers are rich in potassium feldspar, whereas the light-colored paragneiss has little or none (compare Figs. 13 and 14).

Deformation is also more pronounced. The banded gneiss, in contrast to the straight or gently warped paragneiss layers, is often complexly folded, or plastically dragged and smeared out. The amplitude of folding ranges from tens of feet down to a few inches or less (Pl. 3, fig. 1). Some folds are simple flexures, but others show thickening and thinning due to plastic flowage. Some have even been smeared out and partly obliterated, and the remnants refolded. Lineation is intense in many places. Recrystallization and segregation of dark minerals into streaks and clots accompanied the deformation.

As the banded structure becomes increasingly obscured by flowage and feldspathization these rocks grade into migmatite.

Migmatite

Migmatite nomenclature used here follows the definitions recommended by Dietrich and Mehnert (1961). The *neosome* is the "younger part of a composite rock (for example the injected, exuded, or metasomatically introduced material)." The *paleosome* is the "older part of a composite rock (*i.e.*, the remaining or pre-existing parts)."

Large tracts within the gneiss domes consist of migmatite—mixtures of light-colored igneous-appearing rock and darker gneiss. The igneous-appearing fraction (the neosome) forms veins, layers, and cross-cutting bodies in the darker metamorphic rock (the paleosome). The appearance of the migmatite varies from place to place, depending on the character of the paleosome and neosomes and on the amount of deformation. The paleosome consists of banded gneiss at some localities, but more commonly it is plain biotite gneiss (Pl. 1, fig. 1) or dark, mafic-rich gneiss (Pl. 1, fig. 2; Pl. 3, fig. 2). The neosomes, which vein and permeate these host rocks, are fine-to-medium grained granite or alaskite (Pl. 2, fig. 1), coarse porphyroblastic alaskite (Pl. 1, fig. 1), or more rarely pegmatite.

According to Knopf and Jonas (1929, p. 151) the migmatite, which they called injection gneiss, formed by injection of granitic magma into gneiss of sedimentary origin, prior to deposition of the Glenarm Series. In my opinion, however, the structural relations indicate that the neosomes developed in the solid state by metasomatism or exudation. Some of the commonly observed criteria are as follows:

(1) The neosomes are chiefly nondilation bodies. Where they cut across planar structures in the host rock those structures are not offset but continue undeviated on the other side.

(2) Many of the neosomes have irregular, gradational contacts (Pl. 2, fig. 1; Pl. 4, fig. 1). Wispy biotitic septa from the host rock commonly extend out into the neosome with little or no deflection (Pl. 2, fig. 1; Fig. 15).

(3) The neosomes contain undisturbed skialiths of the host rock. No rotated inclusions are found. Also, nebulitic concentrations of biotite may extend through the neosome, retaining the same orientation as the foliation of the surrounding host rock (Pl. 2, figs. 1, 2; Fig. 15).

(4) Dark mafic minerals (chiefly biotite) are concentrated as rims along many of the neosome bodies or as segregations within them (Pl. 2, fig. 2). This might represent mafic material that was expelled as the neosomes grew by replacement, or material that was left behind and segregated as the neosomes developed by exudation. It cannot be explained, however, by magmatic injection.

(5) Some very thin neosomes are "constipated" veinlets. They contain microcline crystals that are larger than the width of the veinlet and that penetrate the wall rock on both sides without deflecting its foliation. These crystals are porphyroblasts, which grew by replacement.

(6) All intermediate stages can be seen in the development of neosomes formed by the coalescence of swarms of feldspar porphyroblasts which grew metasomatically in the gneiss. Where the process was arrested early, the feldspars (chiefly microcline) appear in the gneiss as isolated prophyroblasts, as strings of disconnected prophyroblasts, and as swarms of porphyroblasts still separated from one another by matrix of biotitic gneiss. In the more advanced stages the porphyroblastic feldspars have partly coalesced, but in many places they retain interstitial segregated clots of biotite; the contacts between the porphyroblast swarm and the host rock are still vague and ill-

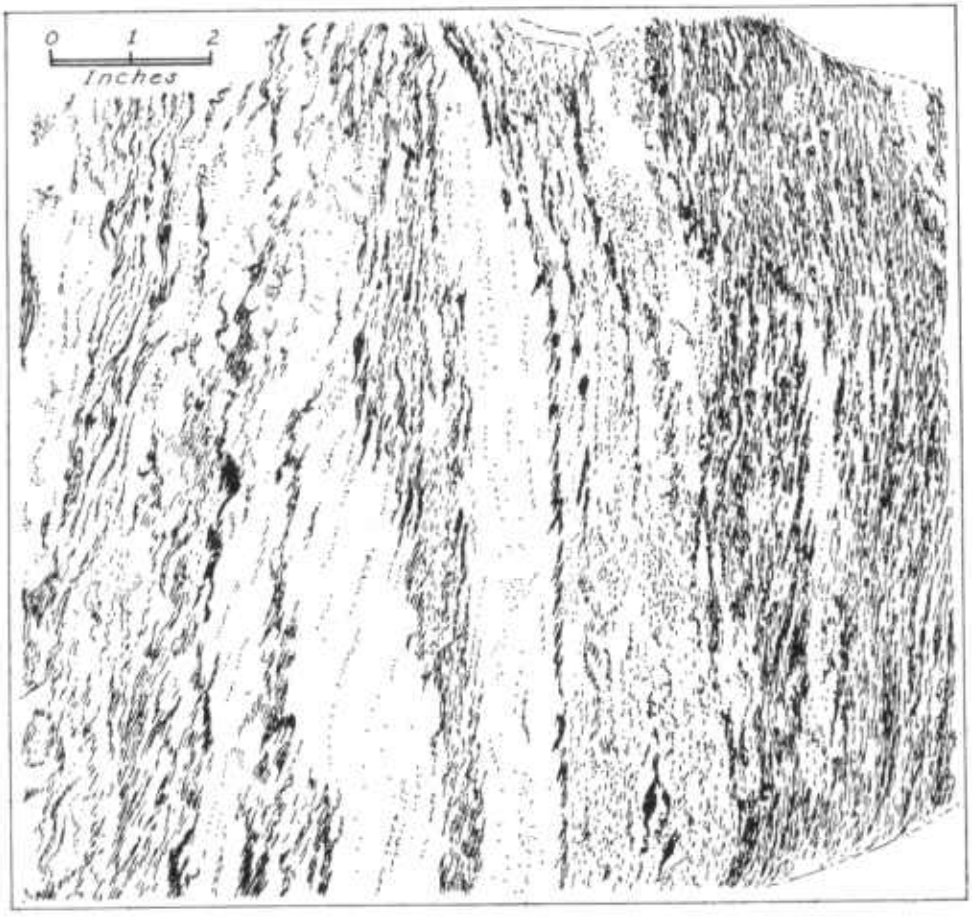


FIGURE 15. Sketch of migmatite, showing light-colored granite replacing dark biotite gneiss. Contacts are gradational, and undisturbed wisps of biotite extend out into the granite. Faint relict foliation from the gneiss runs through the discordant granite. Phoenix dome, half a mile west of Verona. Also see Plate 2, figure 1.

defined (Pl. 4, fig. 1). Finally, coalescence is nearly complete, and the biotite is expelled and concentrated along well-defined borders of the neosome (Pl. 1, fig. 1). Thus all stages of the growth of these neosomes are preserved.

The migmatization was commonly accompanied by strong deformation but locally outlasted it. The migmatitic layering is folded (Pl. 3, fig. 1), ptygmatically buckled (Pl. 3, fig. 2), stretched, pulled apart, or sheared out and partly obliterated. Large porphyroblasts were crushed, drawn out, and recrystallized. Locally the small-scale folding becomes "wild", suggesting that the migmatite was in an incompetent, highly plastic condition. In some places new, undeformed neosomes or swarms of microcline porphyroblasts cut across earlier the folds or flowage structures, showing that migmatization continued where movement had locally ceased.

Structural evidence that the migmatite flowed plastically suggests that anatexis may have been in progress. With the one exception described below it seems unlikely that

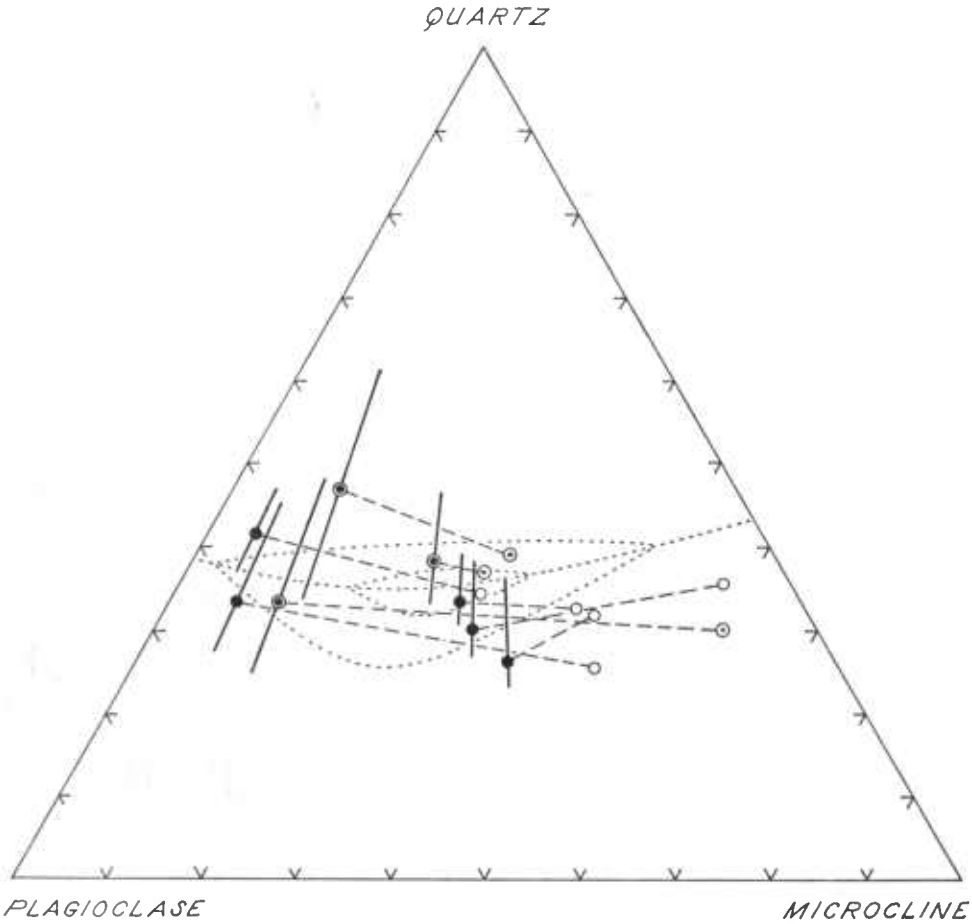


FIGURE 16. Modal compositions of paired dark and light portions from veined gneiss and migmatite. Each pair is joined by a dashed line. The solid circles are veined gneiss dark fraction (paleosome), the open circles are veined gneiss light fraction (neosome). The circles with large dots are the migmatite dark fraction (paleosome); circles with [small] dots are the migmatite light fraction (neosome). The length of the vertical line through each point shows percentages of mafic + accessory minerals. Dotted lines mark the projected position of the minimum-melting trough for the synthetic granite system at 2,000 bars P_{H_2O}; the inner line bounds the lowest point. Note that for each pair the neosome is consistently richer in microcline but does not necessarily plot nearer the minimum-melting composition.

an appreciable fraction of the rock actually melted, although much of it may have become "juicy". The evidence of fluidity displayed in some migmatite terranes, such as swirled foliation, or mafic layers disrupted into blocks and disintegrated into trains of dark minerals, is lacking here. Moreover, there are no petrographic features that suggest crystallization from a melt.

In some migmatites and veined gneisses the quartzo-feldspathic neosomes are thought to have crystallized from anatectic liquid that was pressed and strained from the enclosing darker rock (Turner and Verhoogen, 1960, p. 384–388). Such cannot be the case here because (1) the structural evidence (summarized above) indicates that the neosomes grew in the solid state, and (2) many of them have inappropriate compositions. The bulk composition of the first-melting fraction of the gneiss should plot near the lowest-melting composition of the synthetic granite system (Tuttle and Bowen, 1958).

TABLE 10

MODAL ANALYSES OF MIGMATITIC GNEISS

Baltimore Gneiss

Numbers 1, 2, and 4 are paleosome–neosome pairs that occur in contact.

	1a	1b	2a	2b	3	4a	4b	5
Quartz	33.8	37.6	25.0	30.2	27.4	33.0	35.6	25.9
Plagioclase	30.3	26.8	42.2	9.3	37.5	30.9	31.2	49.8
Microcline	7.9	31.5	8.8	59.3	22.5	21.8	29.5	7.5
Myrmekite	0.2	0.2	0.6	0.9	0.8	1.1	1.2	—
Muscovite	2.0	0.5	—	t	0.5	0.5	0.7	—
Biotite	25.7	3.0	17.8	0.3	11.1	11.7	1.6	14.8
Chlorite	—	0.1	0.4	t	—	—	0.1	—
Epidote	—	—	3.7	t	—	0.2	—	0.2
Allanite	—	—	0.1	t	t	—	—	0.6
Apatite	—	0.1	0.4	—	0.2	0.3	t	0.7
Sphene	—	—	0.5	—	—	0.3	t	0.5
Zircon	—	—	0.1	t	t	t	t	—
Magnetite	0.1	0.2	0.4	t	t	0.2	0.1	—
Total	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0
Points	1062	1205	1054	2404	1798	1736	1823	1076

(1a) Biotitic layer in migmatite (H17-1a).
(1b) Alaskitic layer in migmatite (H17-2).
} Clarksville dome; U. S. 40 west of Pine Orchard.

(2a) Biotitic layer in migmatite (H26-2).
(2b) Alaskitic layer in migmatite (H26-6).
} Woodstock dome, 1 mile east of Marriottsville.

(3) Biotite gneiss layer in migmatite (H26-4). Woodstock dome, 1 mile east of Marriottsville.

(4a) Biotitic layer in migmatitic augen-gneiss ("Hartley Gneiss") (HG1-2b).
(4b) Alaskitic layer in migmatitic augen-gneiss ("Gunpowder Granite") (HG1-1).
} Towson dome, at Hartley.

(5) Biotitic zone in migmatitic gneiss (PD6-1). Phoenix dome, NW. of Western Run.

In Figure 16 the modal compositions of eight paleosome-neosome pairs from the migmatites (Table 10) and veined gneisses (Table 11) are plotted on a triangular diagram similar to that used for the synthetic granite system. Figure 16 differs from the synthetic granite system in that plagioclase (An_{20-25}) takes the place of albite, and microcline the place of sanidine.

Some of the neosomes plot near the minimum-melting composition, but more of them are displaced toward the potassium feldspar corner. For some pairs the paleosome actually plots much closer to the granite minimum than does the neosome—a relationship incompatible with partial fusion. For all of the pairs, however, the neosome is consistently richer in potassium feldspar than is the paleosome, regardless of where they plot with respect to the granite minimum. This argues against liquification and strengthens the conclusion obtained from structural evidence that the neosomes developed by metasomatic growth of potassium feldspar.

In some places, however, the migmatite and other gneisses are cut by small discordant dikes of granite pegmatite that approach the minimum-melting composition. They do not appear to be genetically related to any intrusive granite body, but simply to have come from deeper within the gneiss. Perhaps these are the products of partial fusion, squeezed off to higher levels.

TABLE 11

MODAL ANALYSES OF VEINED GNEISS

Baltimore Gneiss

Tabulated as paleosome-neosome pairs (gneiss and vein material, in contact).

	1		2		3		4		5	
	(a)	(b)	(a)	(b)	(a)	(b)	(a)	(b)	(a)	(b)
Quartz	31.0	32.5	22.8	31.6	37.0	32.0	26.6	35.6	27.5	23.7
Plagioclase	32.6	23.7	29.0	21.2	48.0	29.8	31.3	6.3	49.8	24.3
Microcline	27.6	43.4	33.6	43.8	4.2	30.4	29.6	57.0	4.4	47.0
Myrmekite	0.4	t	0.8	1.0	0.6	3.1	0.4	—	t	2.5
Muscovite	1.5	0.1	—	1.1	—	0.5	0.5	—	2.0	0.1
Biotite	6.4	0.2	11.7	1.3	9.8	4.2	10.8	0.9	14.2	2.1
Garnet	—	0.1	—	—	—	—	—	—	—	—
Clinozoisite	0.2	t	0.4	—	—	—	—	—	0.9	0.1
Allanite	—	—	0.5	—	t	t	0.1	—	0.2	0.1
Apatite	0.3	t	0.1	—	0.4	—	0.3	—	0.2	0.1
Sphene	t	—	1.0	—	—	t	0.4	0.2	0.8	—
Zircon	t	t	t	—	t	t	t	t	t	t
Magnetite	—	—	0.1	—	—	—	—	—	—	—
Total	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0
Points	1100	1022	1050	1344	1045	637	921	570	949	654

(1a) Biotite gneiss (H69-1).
(1b) Alaskitic layer (H69-2).
} Clarksville dome, south of Clarksville.

(2a) Biotite gneiss (H34-1).
(2b) Pegmatitic layer (H34-2).
} Clarksville dome, 1 mile NW. of Pine Orchard.

(3a) Biotite gneiss (H26-5).
(3b) Alaskitic lens (H26-5).
} Woodstock dome, 1 mile east of Marriottsville.

(4a) Biotite gneiss (PD7-1).
(4b) Alaskitic lens (PD7-1).
} Phoenix dome, near Philopolis.

(5a) Biotite gneiss (PD5-1).
(5b) Alaskitic layer (PD5-1).
} Phoenix dome, at Butler.

Veined Gneiss

The largest exposed fraction of the Baltimore Gneiss consists of well-foliated biotite gneiss veined and striped with leucocratic material. Such veining may be weak and inconspicuous, or it may be so pronounced that the gneiss grades imperceptibly into migmatite (Pl. 1, fig. 1). There is little difference, in fact, between the coarsely veined gneiss and some of the migmatite except for the scale of veining. As the veining becomes weak and inconspicuous, the veined gneiss passes gradually into rock described as granitic gneiss.

Mineralogically the veined gneiss is similar to most of the other kinds of Baltimore Gneiss but generally lacks hornblende. The modal compositions of some paleosome-neosome pairs from typical specimens of veined gneiss are given in Table 11. It is evident, from Table 11 and from Figure 16, that the veins are consistently richer in potassium feldspar than their host is.

Textural and structural relations show that the neosomes in the veined gneiss grew by replacement, as they did in the migmatite. The same criteria apply, but it is difficult to establish whether the neosomes are just segregations of quartzo-feldspathic,

especially potassic, material that exuded and collected from the immediate surroundings or if they represent new material that migrated in from a distant source. The same problem holds for the migmatites. If the veined gneisses and migmatites are highly transformed paragneisses like those in southwestern Baltimore, as the structures locally suggest, then the difference in their overall bulk composition requires that a large amount of K_2O was added from an outside source.

Augen Gneiss

Coarse augen gneiss, with large crystals of microcline, is a widespread but minor phase of the Baltimore Gneiss. It is best developed in the northeastern part of the Towson anticline, but good examples are also found in the Texas, Chattolanee, Woodstock, and Clarksville domes. The augen gneiss is most abundant at the margins of the domes, just beneath the Setters contact.

In Baltimore County Knopf and Jonas (1925; 1929, p. 124) named such rock the Hartley augen gneiss, after its good exposures in the Towson anticline near Hartley, and they tried to map it separately. It is not well adapted for large-scale mapping, however, because it occurs chiefly as narrow zones, a few inches to a few tens of feet wide, that are gradational into other kinds of gneiss. Hartley-like augen gneiss, in small patches, is much more widespread in Baltimore County than is shown on the geologic map; conversely, those areas mapped as Hartley Gneiss are actually a mixture of several different gneissic rocks among which augen gneiss is especially prominent. In Howard County no attempt was made to map the augen gneiss separately. It is best displayed in the Woodstock dome, three-quarters of a mile east of Marriottsville along the South Branch of the Patapsco River, and east of Clarksville.

Where best developed the augen gneiss consists of large microcline ovoids closely spaced in a matrix of fine-grained biotite gneiss (Pl. 5, figs. 1, 2). Some of the augen are large single crystals of microcline that have been crushed and elongated, with tails of recrystallized mortar at each end. Others are augen-shaped clusters of quartz, microcline, and plagioclase. Masses of tiny biotite crystals wrap around the augen and form elongate segregations in the matrix (Pl. 5, fig. 2). The matrix tends to contain little potassium feldspar; most of it is concentrated in the augen.

Knopf and Jonas (1929, p. 124–125) thought the augen gneiss had been a porphyritic granite, intruded into the Baltimore Gneiss before it was denuded and covered by the Glenarm Series and later, during Glenarm metamorphism, crushed and recrystallized into augen gneiss. For the following reasons I believe the augen gneiss is a local porphyroblastic phase of the Baltimore Gneiss, formed chiefly by potassium metasomatism along movement zones beneath the Setters quartzite, in part during post-Glenarm doming and metamorphism:

(1) The augen gneiss is not in sharp contact with other kinds of Baltimore Gneiss, nor does it show any other intrusive features. Instead, it grades into the adjoining rocks, where the augen become fewer and smaller and finally disappear.

(2) The augen gneiss does not occur in separate well-defined masses like an intrusive granite but rather in diffuse, ill-defined zones, mostly quite small, that grade into nonporphyritic gneisses.

(3) The augen were not phenocrysts but porphyroblasts, and clusters of smaller grains coalescing into porphyroblasts. In some of these the microcline had begun to join into skeletal masses, enclosing islands of plagioclase and quartz. The spotty distribution of the augen also indicates that they are porphyroblasts rather than pheno-

cryts. In one part of an outcrop the augen are crowded together—so closely, in places, that they coalesce into patches of coarse granite (Pl. 5, fig. 1) or pegmatite-like stringers—but a few feet or even a few inches away only a few isolated augen stud the gneiss. Phenocrysts in magmatic granite would be dispersed more uniformly. Where the gneiss forms alternating light and dark layers, as at Long Green Creek below Hartley, some light-colored augen-rich layers may resemble granitic sills, but many of the dark "host" layers between them are also studded with microcline augen. H. H. Read (1944, p. 80–87) has discussed many similar occurrences and concluded that the big feldspars are metasomatic porphyroblasts. Finally, the microcline augen, or microclines just like them, transgress the unconformity above the gneiss, forming discordant coarse porphyroblastic zones in the base of the Setters Formation. This also can be seen along Long Green Creek, below Hartley (Towson dome).

(4) If the augen gneiss had been intrusive into the Baltimore Gneiss before it was deeply eroded and the Glenarm Series laid down, it is difficult to see why it is now localized along the margins of the gneiss domes, just beneath the Glenarm unconformity. It appears, rather, that the Glenarm cover controlled its formation and distribution.

Structural evidence shows that the Setters quartzite behaved as a competent, rigid unit during the doming, while the gneiss beneath it was an incompetent, plastically deforming mass in which new foliation was developing parallel to the mantle. Possibly, therefore, granitizing solutions migrated along this active movement zone, causing feldspathization and the growth of augen along zones of granulation, shearing, and recrystallization. The newly formed augen were then in turn crushed, drawn out, and recrystallized, as movement and granulation continued.

Granitic Gneiss

The outer portions of the gneiss domes consist mostly of dark biotite-rich gneiss, banded gneiss, veined gneiss, augen gneiss, and migmatite, but many of the domes become more nearly uniform toward their cores, where the rock is chiefly light-colored veined gneiss and finer-grained, more nearly homogeneous granitic gneiss.

This gneiss is a fine- to medium-grained, weakly to moderately foliated rock with microcline, plagioclase (An_{20-25}), and quartz in about equal proportions (Table 12; Fig. 14). Biotite is the chief dark mineral, but small amounts of epidote and muscovite are commonly present. Garnet is rare. In hand specimen (Pl. 4, fig. 2) the rock resembles some intrusive granites. The texture is crystalloblastic, however, showing thorough recrystallization. The term granitic gneiss is purely descriptive and neither implies nor excludes a magmatic origin.

No intrusive contacts between the granitic gneiss and the other gneisses have ever been found. In the field the granitic gneiss is intimately mixed with veined gneiss, and the two commonly grade into one another, even within a single outcrop. The only difference between them is that the veined gneiss has a more coarsely segregated texture. Contorted layers and schlieren-like masses of dark gneiss are locally intercalated with the granitic and veined gneisses, and they appear to have deformed and flowed plastically together.

The origin of the granitic gneiss and associated veined gneiss is a matter of speculation. The remnant layers of locally intercalated dark gneiss suggest that the rock may have come from biotite paragneiss, like that preserved in southwestern Baltimore, and that it has been plastically deformed and thoroughly granitized, becoming much richer in potassium feldspar in the process. (*Compare* the composition of paragneiss and grani-

TABLE 12

MODAL ANALYSES OF GRANITIC GNEISS

Baltimore Gneiss

	1	2	3	4	5	6	7	8
Quartz	30.6	38.9	22.8	32.1	25.7	32.2	28.2	27.0
Plagioclase	36.3	22.2	35.8	29.4	34.9	26.2	27.0	33.5
Microcline	28.1	32.7	31.7	31.1	31.8	32.6	36.3	31.3
Myrmekite	1.2	0.3	1.1	1.1	0.4	1.1	0.4	0.4
Muscovite	0.2	0.3	1.9	0.2	1.9	3.0	4.6	2.4
Biotite	3.4	5.0	5.0	5.4	4.7	4.8	3.2	5.0
Chlorite	—	—	t	—	—	0.1	0.3	0.3
Garnet	t	0.2	—	—	—	—	—	—
Epidote	0.1	0.1	1.1	0.2	0.6	—	—	t
Allanite	0.1	0.1	0.1	0.1	—	t	t	—
Apatite	t	0.1	0.1	0.2	t	t	t	0.1
Sphene	t	0.1	0.4	0.2	t	—	t	—
Zircon	t	t	t	t	t	t	t	t
Magnetite	—	t	—	—	—	—	t	—
Total	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0
Points	1818	1763	1836	1756	1609	1780	1746	1133

(1) Fine-grained granitic biotite gneiss (H26-3). Woodstock dome, 1 mile east of Marriottsville.

(2) Granitic biotite gneiss (H25-2). Woodstock dome, Old Frederick Road, SW. of Davis.

(3) Granitic muscovite-biotite gneiss (H25-3). Woodstock dome, Old Frederick Road, SW. of Davis.

(4) Coarse-grained granitic biotite gneiss (H33-1). Clarksville dome, 1 mile west of Bethany Church.

(5) Fine-grained granitic muscovite-biotite gneiss (H68-2). Clarksville dome, 1½ miles SE. of Highland.

(6) Granitic muscovite-biotite gneiss (TD6-2). Towson dome, along Gunpowder Falls near power line.

(7) Coarse-grained biotite-muscovite gneiss (TD5-1). Towson dome, along Gunpowder Falls near power line.

(8) Fine-grained granitic muscovite-biotite gneiss (TD11-1a). Towson dome, Charles Street Extended near Joppa Road, Towson.

tic gneiss, Figure 14.) Alternatively, these rocks may represent an old intrusive complex, where masses of granitic magma had invaded earlier layered rocks, but in which later metamorphism and plastic flowage erased the igneous texture and intrusive contacts. In that case the reconstitution has been very complete, for no hint of such textures or contacts remain. Still other possible origins can be visualized, but these rocks are so thoroughly recrystallized and so poorly exposed that the problem is far from solved.

Gunpowder Granite

Along Gunpowder Falls in Baltimore County Knopf and Jonas (1925; 1929, p. 125–127) mapped a gneissic granite that cuts across the overturned eastern limb of the Towson dome and can be traced for nearly a mile outward into the Wissahickon schist. This mass, which they named the Gunpowder Granite, is of exceptional interest, because here the Baltimore Gneiss became rheomorphic and broke through the Glenarm mantle.

The geologic relations are complicated at the type locality along Gunpowder Falls, but there is no doubt that the granite has intruded the Glenarm series. It split apart the steeply dipping Wissahickon mica schist, penetrating it in numerous sheetlike bodies that range from several inches to tens or perhaps hundreds of feet wide. The con-

tacts are generally sharp but locally are blurred by feldspathization. Near the Harford Road bridge over Gunpowder Falls feldspathization spread for long distances along the foliation planes of the schist, producing nebulous granite-like layers (Pl. 6, fig. 1). Where the granite cuts Cockeysville Marble there are narrow reaction zones, marked by silicification and concentrations of phlogopite.

Near its type locality the Gunpowder Granite, which is actually a quartz monzonite, is a distinctive fine- to medium-grained gneissic rock with dark, streaky biotitic schlieren. Microcline, plagioclase (An_{20-26}), and quartz are about equally abundant, and muscovite and biotite are the chief varietal minerals (Table 13). Its texture, in thin section, is crystalloblastic and resembles that of the Baltimore Gneiss. If the granite were once partly molten—as suggested by its intrusive contacts and its reaction with the marble—it has been thoroughly recrystallized.

The granite is easily identified and its intrusive relation to the Glenarm Series is clear cut at the margin of the Towson dome, but toward its core the relations become complex. The appearance of the granite inside the dome is more variable, and it is often difficult to decide whether one is looking at Gunpowder Granite or Baltimore Gneiss. Intrusive relations were found at two places: on the north side of Gunpowder Falls near the power line fine-grained gneissic granite cuts discordantly across coarser-textured gneiss (Pl. 6, fig. 2), while on the hillside across the river gneissic granite dikes

TABLE 13

MODAL ANALYSIS OF GUNPOWDER GRANITE

	1	2	3	4	5	6
Quartz	28.8	28.7	32.2	29.0	33.4	35.6
Plagioclase	25.7	26.2	26.6	27.5	24.0	31.2
Microcline	34.2	36.7	27.3	31.8	28.0	29.5
Myrmekite	0.4	0.5	0.9	1.5	0.2	1.2
Muscovite	4.6	3.0	7.7	4.9	6.9	0.7
Biotite	5.6	4.6	4.8	4.2	6.5	1.6
Chlorite	—	0.3	0.1	0.4	0.2	0.1
Clinozoisite	—	—	—	0.3	—	—
Allanite	—	—	—	—	0.3	—
Apatite	0.4	t	—	t	0.3	t
Zircon	0.1	t	t	t	t	t
Magnetite	0.1	—	t	t	0.2	0.1
Calcite	0.1	—	0.4	0.4	—	—
Total	100.0	100.0	100.0	100.0	100.0	100.0
Points	1772	1777	1591	1534	1706	1823

(1) Muscovite-biotite quartz monzonite gneiss (TD5-4). Towson dome; Notch Cliff Road, one-half mile west of Harford Road.

(2) Muscovite-biotite quartz monzonite gneiss (TD8-1). Towson dome; Harford Road, 0.4 miles east of Gunpowder Falls bridge.

(3) Biotite-muscovite quartz monzonite gneiss (B-51). Towson dome; Gunpowder Falls at Harford Road. Drill core, depth 230 feet.

(4) Biotite-muscovite quartz monzonite gneiss (B-52). Towson dome; Gunpowder Falls quarry. Drill core, depth 192 feet.

(5) Biotite-muscovite quartz monzonite gneiss (B-50). Towson dome; Harford Road, 0.2 miles east of Gunpowder Falls bridge.

(6) Alaskitic layer in migmatitic augen-gneiss (HG1-1). Towson dome; at Hartley.

cut layered migmatite. Elsewhere in the area, however, there is a complete gradation from typical Gunpowder Granite to coarser-textured granitic gneiss and layered migmatite. These gradations can be seen on the hillsides along Gunpowder Falls, but they were best displayed underground, during excavation of the Baltimore Gas and Electric Company propane storage cavern near Harford Road.

The granite is evidently a rheomorphic offshoot from the Baltimore Gneiss. Where it breaks across the Glenarm mantle the granitic material is relatively uniform, having probably been homogenized by partial melting, mixing, and flowage. Only the dark schlieren remain as evidence of migmatitic parentage. Traced back toward its roots in the core of the dome, however, the granite is less uniform and begins to take on the characteristics of the gneiss. The texture coarsens, with more segregation of light and dark minerals. The dark schlieren are less streaked out, and some can be recognized as remnants of mafic-rich migmatite layers. Still deeper within the dome the granite becomes quite difficult to recognize, and coarse granitic gneiss and migmatite predominate. Here the structure of the gneiss is "wild", with abundant signs of plastic flowage, such as wavy foliation and streaked-out mafic layers. It is evident that the migmatitic gneiss in the anticlinal core was in a mobile, plastically flowing state.

These structural relations are one line of evidence linking the granite to the Baltimore Gneiss. In addition, the mineralogy and bulk composition of the two rocks are about the same (Fig. 17), so that only flowage and recrystallization are requisite for converting the gneiss to Gunpowder granite. The more abundant muscovite in the granite may be explained by a high partial pressure of H_2O, perhaps developed in connection with anatexis, or crystallization of anatectic melt. It is uncertain what fraction of the rock may have actually melted, because recrystallization has erased any textural evidence of its solidification. Evidently the granite was emplaced during deformation and regional metamorphism.

The Wissahickon schist has been locally granitized at Gunpowder Falls (Pl. 6, fig. 1). It is difficult to decide how much of this is related to intrusion of the Gunpowder Granite. Some of the granitization appears to have been earlier, for some granitized layers are folded and truncated by intrusive granite. Basal Setters schist appears to have been granitized in the same way at the eastern tip of the Towson dome near Hartley, although Knopf and Jonas (1929, p. 124) interpreted this as intrusive granite. The migmatization of the Baltimore Gneiss, the local granitization of schists near the base of the Glenarm Series, and the diapiric intrusion of mobilized migma (Gunpowder Granite) were probably all closely related phenomena, which occurred during the rise of the gneiss domes.

The Geological Map of Baltimore County (Knopf and Jonas, 1925) shows Gunpowder Granite occupying wide areas in the Towson, Texas, and Chattolanee domes. Light-colored granitic gneiss (Pl. 4, fig. 2) that resembles the granite at Gunpowder Falls is prominent in these areas, but veined gneiss, augen gneiss, banded gneiss, and migmatite are also present. Knopf and Jonas (1929, p. 126) recognized the mixture of rocks but showed only the dominant type, as the scale of mapping and the poor exposures made their separation impractical. Re-examination of these areas, however, suggests that it would be better to show it all as Baltimore Gneiss, as has been done on the Geological Map of Howard County (Cloos and Broedel, 1940). The granitic gneiss superficially resembles the granite at Gunpowder Falls but lacks its abundant muscovite and differs in other petrographic details. More important, it does not appear to be intrusive, either into the other types of gneiss or the Glenarm mantle. The diapiric in-

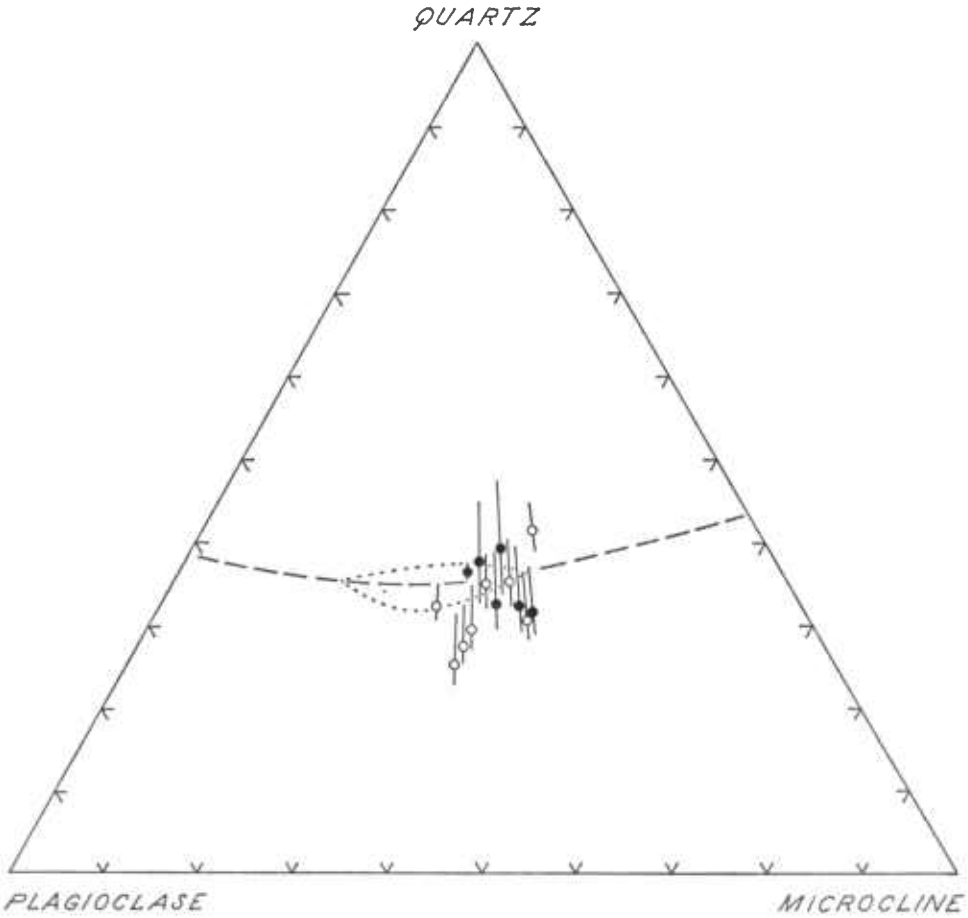


FIGURE 17. Modal composition of Gunpowder Granite (solid circles) and granitic gneisses from the Baltimore Gneiss (open circles). Top of line marks percentages of microcline + plagioclase; bottom of line marks percentages of quartz. Vertical distance between ends of line marks percentages of mafic + accessory minerals. Dashed line marks the projected minimum-melting trough of the synthetic granite system at 2000 bars P_{H_2O}, and the dotted line bounds the granite minimum composition.

trusive mass at Gunpowder Falls appears to be something truly different, and it would be better to restrict the name Gunpowder Granite to this mass.

PETROGRAPHY

Despite the wide variety of structural forms taken by the gneiss its mineralogy and microscopic textures are much the same.

Most of the gneiss consists mainly of plagioclase, quartz, and potassium feldspar, with biotite as the chief dark mineral. Potassium feldspar is scarce in the darker gneiss but equals or exceeds plagioclase in rocks where ferromagnesian minerals are scarce or absent (Fig. 14). Muscovite, clinozoisite (or epidote), allanite, apatite, sphene, zircon, and magnetite are common minor accessories. Other tiny accessory minerals, some of which form pleochroic halos in biotite, were not identified. A few of the gneisses contain garnet, but other aluminum-excess minerals, such as sillimanite or cordierite, are absent. Green hornblende is common in the more mafic gneisses, and diopside is found

in some of the amphibolite. Small amounts of chlorite locally replace biotite. Details of mineralogy of the different types of gneiss can be seen from the modal analyses (Tables 8–12).

The plagioclase is chiefly oligoclase, An_{20-28}. In some of the hornblende gneiss and amphibolite it is sodic andesine. In this respect the Baltimore Gneiss amphibolites can be distinguished from those of the Baltimore Gabbro Complex, in which the plagioclase is labradorite and bytownite. Pericline and albite twinning are common. None of the plagioclase in the Baltimore Gneiss shows well-defined zoning.

The potassium feldspar is exclusively microcline, with well developed grid twinning. Microperthite is rather uncommon, and the perthitic lamellae are tiny and irregularly developed. Some optically homogeneous microcline was X-rayed and found not to be cryptoperthitic. Probably the feldspar is largely unmixed because of continuous recrystallization during plastic flowage of the gneiss. The microcline is locally invaded by small lobes of myrmekite.

Biotite, in well-formed flakes, is pleochroic from deep chocolate brown or blackish brown to pale straw yellow. A chemical analysis of typical biotite from biotite-rich gneiss in the Towson dome is given in Table 14, along with the calculated structural formula. The Baltimore Gneiss does not contain any reddish biotite, found in many other high-grade metamorphic terrains. The muscovite occurs in small flakes and shreds but appears to be in equilibrium with the feldspar.

The gneisses are thoroughly recrystallized. Whatever the original character of these rocks may have been, no trace of their primary textures remain. The feldspars and quartz are xenoblastic, with irregular, interlocking borders. Metamorphic differentiation, which segregated the dark and light-colored minerals into streaks and thin bands, is pronounced in some of the gneisses but inconspicuous in others.

The thoroughly crystalloblastic texture of the gneiss, and evidence that mineralogical equilibrium was approached, reinforce the conclusion from structural evidence that the gneiss flowed plastically—by granulation and recrystallization, like the flowage of glacial ice. Other textural features also suggest such movement. Feldspar grains are bent and broken, quartz is strongly undulatory, and mica flakes are twisted and sheared. But in most of the rocks these effects are partly effaced by annealing or recrystallization;

TABLE 14

CHEMICAL ANALYSIS OF BIOTITE FROM BALTIMORE GNEISS

SiO_2	35.32	K_2O	9.57
TiO_2	3.45	H_2O^+	3.09
Al_2O_3	15.30	H_2O^-	0.02
Fe_2O_3	3.33	F	0.18
FeO	22.94		
MnO	0.26		99.98
MgO	6.18	$-O\equiv F$	0.08
CaO	0.20	Total	99.90
Na_2O	0.14		

H146-1. Biotite from biotite-microcline-quartz-oligoclase gneiss, Towson dome. Charles Street at Sheppard Pratt Hospital entrance, Baltimore County. O. von Knorring, analyst.

Structural formula:

$$\left[\underset{+0.27}{\underline{(Al_{0.15}Ti_{0.20}Fe^{'''}_{0.20}Fe^{''}_{1.49}Mg_{0.72}Mn^{''}_{0.02})}}\ \underset{-1.26}{\underline{(Si_{2.74}Al_{1.26})O_{10}(OH)_2}}\right]^{-0.99}\ \underset{+0.98}{\underline{(K_{0.95}Na_{0.02}Ca_{0.01})^{+0.98}}}$$

true cataclastic textures are scarce. The micas—which are delicate and most easily deformed—recrystallized most readily: in many rocks in which the feldspars retain bent twinning and other evidence of strain the micas that are crowded between them have perfectly sharp, straight extinction. Narrow zones of granulation appear in some thin sections, providing evidence of shear. Streaky masses of fine-grained undulatory quartz and feldspar and tiny twisted shreds of mica form thin trails of mortar, which may parallel the foliation or cut across it. Most of these shears are partly healed by recrystallization, with new unstrained minerals growing into or across the granulated material.

Textural evidence shows also that during and following the deformation potassium feldspar was chemically mobile, migrating and replacing other minerals. Groups of large microcline grains form stringers and knots with biotite densely clustered around the margin. These are analogous to some of the replacement neosomes in the migmatite but are developed on a microscopic scale. Microcline also forms large irregular porphyroblasts, which enclose earlier quartz and feldspar and penetrate along the grain boundaries of the enclosing minerals. In many of the finer-grained gneisses thin rinds of microcline appear between plagioclase and quartz grains, replacing their margins. Some of the sheared, granulated zones described above are penetrated by amoeba-like masses of new, unstrained microcline; after shearing the granulated rock was evidently infiltrated and partly replaced by the microcline. These and other features suggest that flowage of the gneiss, by pervasive shearing and recrystallization, was accompanied by widespread migration of potassium.

STRUCTURE

The structure of the gneiss domes has been studied in detail by Broedel (1937). His maps (Pls. 28–31 of his paper) give a wealth of information on the planar and linear structures in the gneiss and its mantle. From these data and from my own observations some generalizations may be drawn.

The foliation and banding in the gneiss dips steeply and strikes parallel to the long axes of the anticlinal domes. The mineral lineation plunges gently to steeply within the foliation plane. Toward the margins of the domes the gneissic structure tends to conform to the contact with the overlying Glenarm Series. The planar structures dip parallel to the limbs and wrap around the noses of the anticlinal domes. This fact, together with the observation that the axes of small-scale folds in the gneiss and adjacent Setters quartzite are often parallel (Fig. 18; also Broedel, 1937, p. 175), shows that the gneiss and mantle were deformed and moved together. The gneiss appears to have developed a new foliation parallel to the Glenarm contacts. Inward from the contact the gneissic structure rapidly becomes more variable.

The circular Woodstock dome and some of the other fatter domes show evidence of an upward, stretching type of movement. Steep migmatitic banding is thinned and drawn out. The lineation, especially mineral streaking, plunges obliquely or steeply down the foliation planes. Commonly this is paralleled by steeply plunging streaks on bedding surfaces of the adjacent Setters quartzite. These features, and also the rounded shape of the domes, suggest that the gneiss was moving upward. In the more elongate anticlinal domes the lineation is commonly subhorizontal and plunges gently down their noses. Perhaps here lateral compression or horizontal stretching was more important.

But these generalizations tend to obscure the fact that the internal structure of the gneiss is often extremely variable and irregular and does not reflect a simple movement pattern. For instance, the streaky mineral lineation may vary in intensity and change

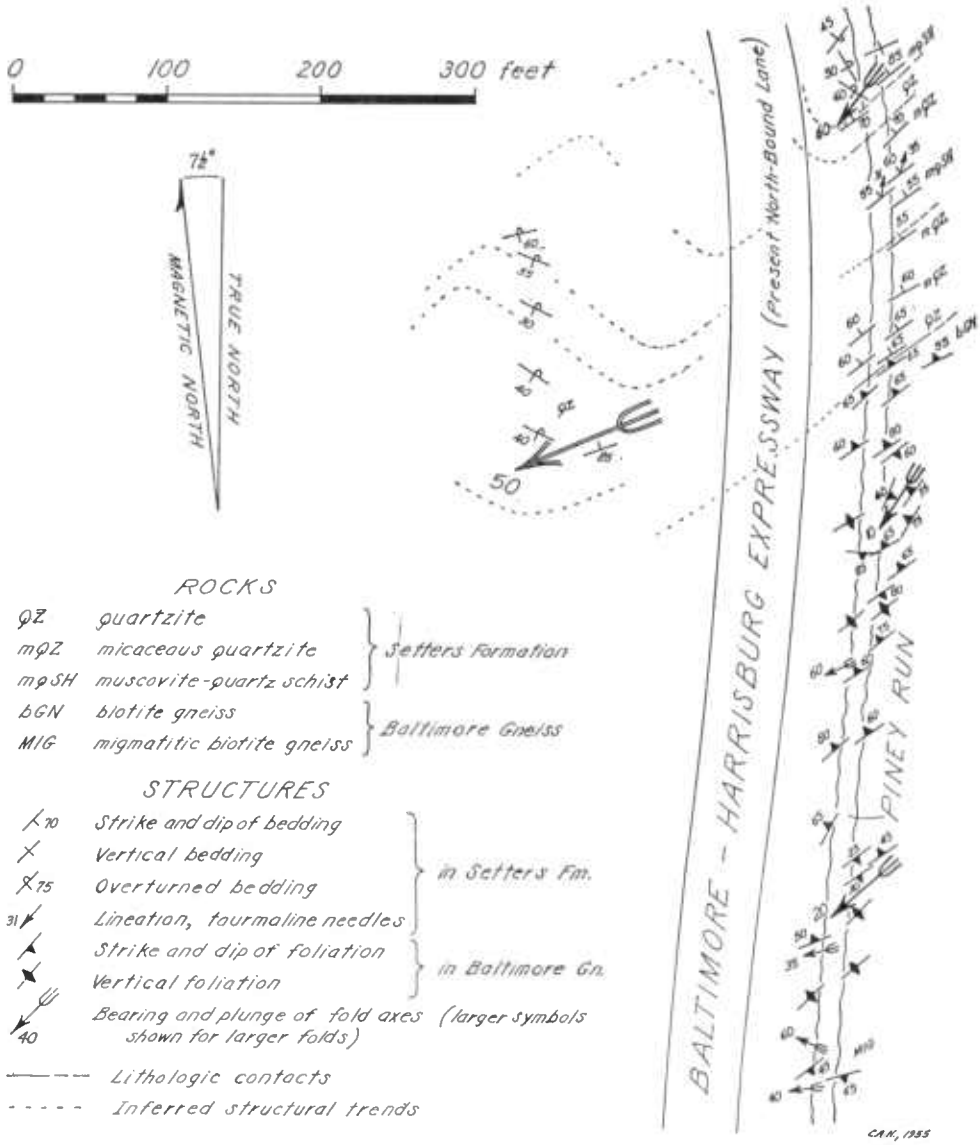


FIGURE 18. Map of geologic structures near the Setters Formation–Baltimore Gneiss contact. Northern flank of Phoenix dome, half a mile west of Verona.

its direction abruptly within a few feet or tens of feet. Some folds have been refolded, so that their axes bend from horizontal to nearly vertical within a few feet, or follow highly sinuous trends like soft strands of spaghetti. Some folds are truncated by bands or streaky masses of gneiss that cut obliquely across them; here the gneiss was dragged and smeared out in a new direction, obliterating the earlier folds. Evidently local movements, changing in intensity and direction, occurred repeatedly as the incompetent, plastically deforming gneiss rose as domes and anticlines.

Evidence for plasticity is greatest where the gneiss is migmatitic or strongly veined with quartzo-feldspathic material. Wavy irregular banding, recumbent and refolded

flowage folds, ptygmatic folds, and smeared out structures of all kinds are most characteristic of the migmatitic zones. This relationship has been consistantly observed in mobilized basement complexes throughout the world (Read, 1955, p. 411–414), lending strong support to the hypothesis that granitizing fluids increased the ability of the gneiss to deform and flow plastically.

SUMMARY AND CONCLUSIONS

The Baltimore Gneiss consists of gneissic and migmatitic Precambrian rocks that evidently had a long and complex history even before they were denuded and covered by the Glenarm Series. Some of the gneiss probably formed from volcanic sediments, but most of it has been so thoroughly transformed that its original character is beyond recognition. Following the deposition of the Glenarm Series the gneiss and overlying rocks were up-arched into the Baltimore-Washington anticlinorium and regionally metamorphosed. During the process the basement rocks flowed plastically, and along the crest of the anticlinorium the most mobile material punched upward as domes and arches into the overlying rocks. Migmatization accompanied and probably facilitated the movement. In at least one place the gneiss became rheomorphic and broke intrusively into the Glenarm mantle (*i.e.*, Gunpowder Granite).

Similar examples of mobilized basement rocks, long recognized in Europe, Fennoscandia, Greenland, and elsewhere, have been summarized by Wegmann (1935) and H. H. Read (1955). A paragraph from Wegmann's paper, describing the interaction between mobilized migmatitic rock (infrastructure) and its nonmigmatitic, more rigid cover (superstructure), is especially pertinent to the Baltimore domes (translated from Wegmann, 1935, p. 334):

"The folding of the superstructure can occur in two ways: a) by tangential compression; b) by the infrastructure forming anticlines and arching it [the superstructure] upward. According to present experience both may occur together. Thus the superstructure glides rather firmly over the infrastructure, or the infrastructure flows under the superstructure. If it flows into anticlines, quite a confusing picture results. The dip of the movement planes in the larger-scale features of the infrastructure are steeper than the inclination in the superstructure. In the tightly folded anticlinal zones of the infrastructure recrystallization and formation of nebulites are often especially intense. At some places apparently granitic stocks break through the folded gneisses. Detailed investigation proves that many of these, however, are the recrystallized continuation of the strongly folded migmatite, whose inner structure has disappeared."

In the case of the Baltimore domes, the boundary between superstructure and mobilized infrastructure falls at the unconformity between the Glenarm Series and the Baltimore Gneiss.

If migmatization facilitated plastic flowage of the gneiss during post-Glenarm doming then perhaps it is surprising that the lowest formations in the Glenarm Series are so little affected. The reason may be twofold:

1. These rocks are chiefly quartzite and marble, notable resistors to granitization (Read, 1957, p. 346–347). Perhaps these rocks formed a barrier, as Read (1957, p. 357) has suggested for a similar case in Ireland, that protected the overlying Wissahickon Formation from the granitizing substances that permeated the underlying gneiss. Where granitization did extend through to the Wissahickon, near Gunpowder Falls (Towson dome), the quartzite is locally missing, and the marble is thinned to 30 feet or less as shown in drill cores taken at Gunpowder Falls near Harford Road.

2. The gneiss may have already been migmatitic from an earlier metamorphic cycle,

and then had a second, perhaps milder migmatization superimposed upon it during the Glenarm doming. This could explain why the gneiss is so much more thoroughly veined, granitized, and plastically deformed than the lowest Glenarm rocks. The dating of porphyroblastic microclines from migmatite and augen gneiss in the Phoenix and Towson domes at about 1100 million years (Tilton *et al.*, 1959) supports this idea.

Certain features of the gneiss domes invite comparison with intrusive magmatic plutons. There is a rise in metamorphic grade around the domes, marked by the appearance of kyanite in the pelitic schists, that might be likened to an aureole of contact metamorphism. Also, swarms of large granite pegmatite sills and dikes cluster around the gneiss domes (Fig. 44), somewhat analogous to apophyses from granitic plutons.

Nevertheless, the gneiss cannot be interpreted as magmatic granite wholly younger than the Glenarm Series: (1) In southwestern Baltimore it consists of well-bedded paragneiss. Banded gneiss and migmatite, which are widespread in the other domes, are more thoroughly transformed than the paragneiss but still record a supracrustal origin. (2) The gneiss is concordantly overlain by the Glenarm beds; in most places it is capped by the same unit (Setters Fm.). (3) The gneiss does not contain xenoliths of Glenarm rocks. (4) The gneiss lacks any petrographic features, such as relics of euhedrally-zoned plagioclase, that might indicate a magmatic origin. Had the gneiss solidified from magma when the Glenarm rocks were domed, some evidence of it would surely survive.

On the other hand, all features of the gneiss and the overlying rocks are explained if the gneiss is the older basement which was heated, granitized, and made sufficiently mobile to flow plastically and dome its mantle. Anatectic fluids, constituting but a very minor fraction of the rock, rose from the gneiss to form pegmatites in the adjacent Glenarm Series (see discussion of pegmatites, pp. 186–187, and Fig. 45). The fringe of higher grade, kyanite zone metamorphism which surrounds the domes might be partly a thermal effect related to upward migration of heat-carrying granitizing fluids, but pressure probably had a greater influence than heat. Pressure in excess of 10 kilobars seems necessary for stable development of kyanite at regional metamorphic temperatures (Bell, 1963, Fig. 1; Khitarov *et al.*, 1963, Fig. 5). Pressures of this magnitude cannot be accounted for in the Baltimore area solely by lithostatic load, which would necessitate at least 30 km of cover. Tectonic overpressures (Clark, 1961, p. 647–649; Coleman and Lee, 1962, p. 589–592) were therefore probably involved. Such overpressures would likely have developed where the domical masses of gneiss were squeezed upward from deeper root zones in the basement.

My interpretation of the Baltimore gneiss domes is essentially the same as that made by Eskola (1949) for the mantled domes in Finland, and which he also suggested for the Baltimore domes. The present study only adds details, most of which strengthen his hypothesis.

An important conclusion from Eskola's theory was that mantled gneiss domes appear only in regions that have undergone two orogenic revolutions. This condition is fulfilled in the Appalachians, where radiogenic dating (Tilton *et al.*, 1958; 1959; 1960; Long *et al.*, 1959) has demonstrated that Paleozoic orogeny was preceded by a Precambrian regional metamorphism about 1000–1100 million years ago.

An unsolved problem concerns the age and stratigraphic position of the paragneiss in southwestern Baltimore. Are these well bedded volcanoclastic gneisses correlative with the more strongly transformed banded and migmatitic gneisses in the other domes? The paragneiss might have resembled these rocks quite closely, had it undergone the same degree of potassic granitization and flowage.

But another interpretation is possible. Northeastward along strike from the paragneiss are similar rocks mapped as Baltimore Gneiss in Harford County (Mathews and Johannsen, 1904; Mathews, 1933), and beyond lies the metavolcanic complex of Cecil County (Bascom, 1902; Marshall, 1937). Some of the Harford County gneiss is well layered and strikingly resembles the paragneiss in Baltimore. The Cecil County metavolcanics are also similar in composition and structure but less strongly metamorphosed. The Baltimore paragneiss might be correlative with these rocks, since it lies on strike with them and is also derived from volcanic material. None of these rocks lie directly beneath the Setters and Cockeysville Formations, and none have been proven to be part of the 1-billion year old basement. More work is needed to resolve their age and their stratigraphic relation to the Glenarm Series.

GLENARM SERIES

INTRODUCTION

The Glenarm Series was named by Knopf and Jonas (1922; 1923), whose definition states (1923, p. 45):

> "Overlying the Baltimore gneiss is a series of pre-Cambrian sediments here named the Glenarm series, from its typical development near Glenarm, 13 miles northeast of Baltimore. The Glenarm series comprises the Setters formation, the Cockeysville marble, the Wissahickon formation, the Peters Creek schist, the Cardiff conglomerate, and the Peach Bottom slate. The total thickness of the series probably amounts to between 8,000 and 10,000 feet, although no accurate estimates can be made, for the middle formations have been repeated by close folding. So far as now known, deposition of this series was not interrupted by erosion or by orogenic deformation, although the earlier formations are overlapping shore deposits."

Later work (Stose and Jonas, 1939, p. 103–106) indicates that the Cardiff Conglomerate and Peach Bottom Slate lie unconformably above the Peters Creek Formation. Accordingly, the Cardiff and Peach Bottom were withdrawn from the Glenarm Series.

The work in Howard and Montgomery Counties suggests the need for further revisions. The Peters Creek Formation, when traced southward to Howard County, loses its distinctive character and appears to grade laterally into the Sykesville Formation. The two formations are continuous along strike (Fig. 30) and may represent different contemporaneous facies. Near Washington, D. C., the Sykesville wraps around the nose of the Baltimore-Washington anticlinorium and reappears on the eastern limb as the Laurel Formation. The Sykesville and Laurel Formations are thought to be correlative because they are nearly traceable into one another in Washington, D. C. and Arlington, Virginia, they both directly overlie the eastern sequence of the Wissahickon Formation, and they have the same distinctive lithology.

Knopf and Jonas (1923; 1929) and most subsequent workers have assumed that the entire Wissahickon Formation lies beneath the Peters Creek. This is founded on the belief that the Peters Creek Formation occupies the core of a major fold, called the Peach Bottom syncline, and that the Wissahickon oligoclase-mica schist which bounds the Peters Creek on the southeastern side reappears as the albite-chlorite schist on the northwestern side (Fig. 19A). In Howard and Montgomery Counties, however, there is no evidence for such a syncline, and the Glenarm sequence appears to be continuous and to become younger to the west. The existence of a small syncline at Peach Bottom, outlined by the canoe-shaped Cardiff Conglomerate, is well established (Knopf and Jonas, 1929; Agron, 1950) but its visible extent is small, and if the Cardiff Conglomerate

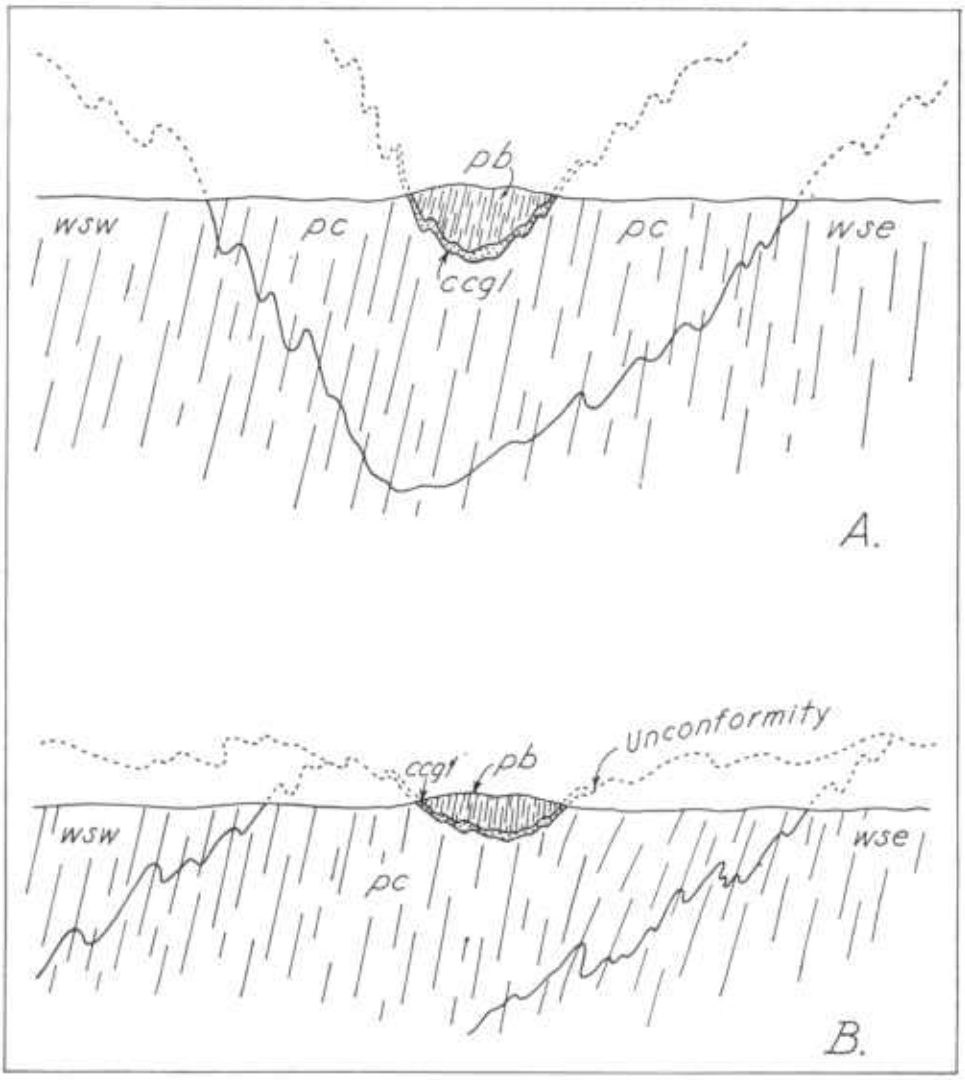


FIGURE 19. Two possible interpretations of structure in the Glenarm Series at the Peach Bottom syncline.

A. The Glenarm formations (Peters Creek, Wissahickon) are synclinally folded with the overlying rocks. The Wissahickon lies beneath the Peters Creek Formation, and its eastern and western facies are stratigraphically equivalent.

B. The Cardiff Conglomerate and Peach Bottom Slate rest unconformably on steeply-dipping, homoclinal Glenarm formations. The Peach Bottom syncline (center) is a second-order fold that does not significantly affect the Glenarm structure. The Wissahickon western facies is younger than the Peters Creek Formation and the Wissahickon eastern facies.

The units shown are: Pb, Peach Bottom Slate; ccgl, Cardiff Conglomerate; pc, Peters Creek Formation; wsw, Wissahickon Formation, western facies, wse, Wissahickon Formation, eastern facies.

rests unconformably upon the Peters Creek Formation (Stose and Jonas, 1939) then the Glenarm formations are not necessarily synclinal, and the eastern and western belts of the Wissahickon need not be stratigraphically equivalent. Figure 19B shows this possible alternative interpretation.

The Glenarm Series in Howard and Montgomery Counties consists of the sequence: Setters Formation, Cockeysville Marble, Wissahickon Formation (eastern sequence), Sykesville Formation = Laurel Formation, Wissahickon Formation (western sequence). The top of the series is arbitrarily placed at the base of the Ijamsville Phyllite, but further studies may show that this unit should be included.

The age of the Glenarm Series has long been controversial, but since 1935 most geologists have favored an early Paleozoic age (Miller, 1935; Mackin, 1935; Cloos and Hietanen, 1941; Swartz, 1948; Watson, 1957; McKinstry, 1961). However, the present study shows that the Glenarm sequence in the Baltimore-Washington area must be Cambrian or earlier, and that it is probably latest Precambrian.

UNCONFORMITY AT THE BASE OF THE GLENARM SERIES

Beginning with E. B. Mathews (1904; 1905) all geologists who have studied the Maryland Piedmont have recognized a major unconformity between the Glenarm metasedimentary rocks and the underlying Baltimore Gneiss. This is perhaps surprising, because at no place is there actually an angular discordance. The arguments for the unconformity are therefore summarized, and the evidence brought up to date.

According to Knopf and Jonas (1929, p. 103–104): "The Baltimore gneiss had been intruded by gabbro and by granite before the deposition of the Glenarm Series." They (1929, p. 154) also state:

"Evidence of the unconformable nature of the contact between the two formations is seen in the character of the deformation which is more pronounced in the Baltimore gneiss than in the Setters formation. There is also a divergence in average strike between the two formations that amounts to about 20 degrees."

According to Broedel (1937, p. 162):

"The unconformity in Maryland has been largely obliterated by intense folding during uplift of the domes. The best evidence for a general unconformity is shown by the much greater deformation and igneous injection in the gneiss as compared with the Glenarm Series. In addition the Hartley augen gneiss and the hornblende gneiss, which occur in the Baltimore gneiss, are absent in the Setters and overlying formations. In contrast to the gneiss with its high degree of minor isoclinal folding and its ptygmatic folds, the quartzite is relatively free from such minor folds except locally. It was protected by and folded along with the old gneiss surface during post-Glenarm deformation."

Some of these reasons appear to be dubious or invalid. There is no clear-cut evidence in this area that plutonic rocks intruded the Baltimore Gneiss before the Glenarm Series was laid down. The Hartley augen gneiss is not a metamorphosed pre-Glenarm granite as Knopf and Jonas had supposed, but a porphyroblastic phase of the Baltimore Gneiss. At least part of it developed during post-Glenarm metamorphism. The hornblende gneiss, which occurs as thin layers in the Baltimore Gneiss, is not metagabbro, as Broedel implies, but a paragneiss probably derived from mafic volcanic sediments. Other than this, there is no pre-Glenarm gabbroic rock in the Baltimore area. Perhaps Knopf and Jonas referred to southeastern Pennsylvania, where Bascom (1905, p. 296) described gabbroic rocks intruding Baltimore Gneiss.

Regarding the 20-degree divergence in strike between the gneiss and the quartzite at the Phoenix dome Broedel (1937, p. 162–163) says:

"To the writer this appears to be insufficient evidence [for an unconformity], since the foliation of the

gneiss and the bedding of the quartzite often vary considerably in strike as a result of minor folding even within a short distance of the contact. Where contacts are exposed they are always conformable."

Broedel is correct. New road cuts opened in 1955 during construction of the Baltimore-Harrisburg Expressway provided continuous exposures across the gneiss–quartzite contact at the locality where the 20-degree divergence had been described. At the contact the foliation in the gneiss perfectly parallels the lowest Setters beds (Fig. 18). Beginning about 30 feet beneath the contact, however, the gneiss becomes increasingly contorted and folded on a small scale. Here and elsewhere my observations confirm those of Broedel, that wherever the actual contact can be seen the planar structures in the gneiss are parallel to the lowest quartzite bed. Nevertheless, this does not rule out the possibility that a strong angular discordance once existed. There was considerable plastic flowage in the gneiss during post-Glenarm metamorphism and doming, especially near the contact, so it would indeed be surprising if traces of the earlier structure were not erased. In the Green Mountain anticlinorium Brace (1958) has shown that beneath the unconformity, in a zone several thousand feet thick, the discordant Precambrian beds were dragged nearly parallel with the younger strata, or sheared out along surfaces parallel to the unconformity. In the Baltimore domes, where the folding was accompanied by migmatization and plastic flowage, the earlier structure would have been obliterated even more readily.

Everyone has called attention to the more intense deformation of the gneiss as compared with the Glenarm Series. The contrast is strong indeed, if one compares the "juicy" migmatitic gneiss with the Setters quartzite, but this reflects only their difference in competency at the time of deformation. The incompetent Cockeysville Marble displays flowage folds nearly as intricate as those in the gneiss.

Nevertheless, an unconformity between the gneiss and the Glenarm Series still seems to be the best interpretation. It is favored for the reasons listed below, and there seems to be no compelling evidence against it.

(1) The Glenarm rocks preserve their sedimentary structures and their original composition, whereas the original character of the Baltimore Gneiss has been almost completely effaced by recrystallization, migmatization, and flowage. Indeed, the gneiss is so thoroughly transformed that, except for the paragneiss in southwestern Baltimore, it is generally difficult even to say whether it was originally plutonic or supracrustal. It is hard to believe that two contiguous rock series would display such a vast difference in degree of transformation if they had both undergone the same amount of metamorphism.

(2) It has been shown by radiogenic dating (Tilton *et al.*, 1958) that the Baltimore Gneiss participated in two periods of crystallization: one about 1000–1100 m.y. ago (Precambrian) and the other during the Paleozoic. This is supported by new mineral dates obtained since. (*See* chapter on Radiogenic Dates.) Unless the Glenarm Series is older than about 1100 m.y.—which seems most unlikely—the present Baltimore Gneiss was already crystalline when the first Glenarm sediment was deposited.

(3) Ancient gneissic basement rocks, many of them evidencing a 1000 m.y. period of recrystallization (Tilton *et al.*, 1960), are recognized throughout the central and southern Appalachians. Along the Blue Ridge and in southeastern Pennsylvania the gneiss is overlain unconformably by Lower Cambrian or uppermost Precambrian strata. These relations do not prove an unconformity beneath the Glenarm Series, but they show that a 1-billion year-old basement is widespread in the region and is to be expected beneath the supracrustal formations.

SETTERS FORMATION

Name

Rocks of the Setters Formation were first described by G. H. Williams (1891, p. 308), who noted that they were "... best exhibited in the abrupt east-west hill, north of Baltimore, known as Setter's (Tyson: "Sater's") ridge." He subsequently named these rocks the *Setters quartz schist* (Williams and Darton, 1892) and gave the following brief description (quoted from Wilmarth, 1938, p. 1958):

"A peculiar schist composed mostly of quartz and divided into beds of varying thickness by parallel layers of muscovite. Underlies Cockeysville marble and overlies Baltimore gneiss. Most characteristic occurrence is in Setters Ridge."

The unit has since been widely recognized, not only in Maryland but in the Piedmont of Pennsylvania, and has often been described as Setters Quartzite. Knopf and Jonas (1929, p. 152–153) noted that feldspathic schist and mica gneiss were also important members of the quartzitic sequence, and so renamed the unit the Setters Formation.

Occurrence

The Setters Formation is best exposed in Baltimore County, but it extends southwestward through Howard County, where it crops out as narrow rims around the Woodstock, Mayfield, and Clarksville domes. Its southern limit is about 2 miles south of Highland, where it plunges southward beneath the younger formations. It does not reappear at the surface anywhere in Montgomery County.

The Setters forms low ridges, because its quartzitic rocks are more resistant to weathering than the calcareous and feldspathic formations on either side. The exposures in Howard County are very poor, and even along the ridges outcrops are scarce. The unit is best exposed along the western side of the Woodstock dome near Marriottsville, along the Patapsco River near Alberton and Davis, on the west limb of the Clarksville dome north of U. S. Route 40, and on the southern flank of the dome near Nichols.

Along the east and west flanks of the Clarksville dome the Setters Formation has been cut out by faulting, so that the Wissahickon Formation directly abuts the gneiss (Cloos and Broedel, 1940). Elsewhere the Setters appears to thin gradually and to disappear with no evidence of faulting. According to Knopf and Jonas (1923, p. 45):

"... the Setters is irregular in thickness and in some places is entirely lacking, so that the Cockeysville marble overlaps the Setters deposits and lies directly upon the Baltimore Gneiss. In the absence of positive evidence of erosion, it seems likely that the irregular thickness of the Setters Formation and of the Cockeysville marble is due to transgression rather than to erosion."

Bascom and Stose (1932, p. 10) came to a similar conclusion in Chester County, Pennsylvania:

"The absence of the Setters formation and the Cockeysville marble in such places may be due either to pre-Wissahickon erosion or to non-deposition of these formations, for there is undoubtedly considerable variation in the thickness of these sediments."

McKinstry (1961, p. 559–560) invoked a tectonic explanation, suggesting that their absence could be due to shearing out or extreme attenuation of the beds during folding. He points out that in Chester County those formations are most commonly missing along the steeply overturned limbs of the anticlines, where tectonic thinning would be greatest. In Howard and Baltimore Counties, however, the disappearance of the Setters

Formation shows no clear dependence on overturned fold limbs, and overlap may be the better explanation. The local thinning and disappearance of the lower and middle members of the Setters, described below, seems stratigraphic and not tectonic.

Thickness

There is no place in Howard County where the thickness of the formation can be reliably measured. The micaceous upper and lower members weather readily and are not completely exposed. The formation appears to be about 200 to 500 feet thick in most places.

In Baltimore County Knopf and Jonas (1929, p. 161) estimated a maximum thickness of 1000 feet. This figure is probably too large. At Oakleigh (Towson dome) they measured 700 feet of Setters section and estimated that an additional 300 feet is concealed above. New exposures reveal that this 300 feet is chiefly Cockeysville Marble. Choquette (1957, p. 5), who has mapped the area in greater detail, lists the Setters Formation as 0–750 feet thick.

Lithology

The rocks are mainly feldspathic mica schist, mica gneiss, feldspathic quartzite, and micaceous quartzite. Bedding is well preserved, especially in the quartzites. Schistosity parallels bedding, but in the mica schists a later, discordant fracture cleavage is also developed locally. The fracture cleavage crinkles the schistosity plane, producing a horizontal to gently dipping lineation.

In Baltimore County Knopf and Jonas (1929, p. 153–155) separated the formation into three members: a lower group of mica schists, a middle group of predominantly quartzitic rocks, and an upper group of mica gneiss and mica schist. This division holds also for Howard County. In some places the lower or the middle member is very thin or absent. For example, at Long Green Creek, Towson dome, the lower schist member is about 200 feet thick, but at the Phoenix dome, west of Verona, only about 5 feet of this schist separates the quartzite member from the Baltimore Gneiss.

The lower member is chiefly medium-grained feldspathic mica schist. The feldspar is predominantly or entirely microcline (Table 15, cols. 1–4). Locally the texture becomes quite coarse, and segregated pods and lenses of microcline are prominent. Beds of quartzite, 2–8 inches thick, are locally interstratified with the schist, especially toward the base, and one of these beds commonly marks the contact with the Baltimore Gneiss. Thus at Cromwell Bridge, Towson dome, a 3-inch quartzite bed separates the basal mica schist from pink granitic gneiss. Again at Phoenix dome, west of Verona, a thin quartzite bed separates coarse feldspathized muscovite schist from biotite-rich Baltimore Gneiss.

The lower schist locally shows evidence of granitization. Good examples are found near Hartley (Towson dome) and also 1 mile west of Verona (Phoenix dome) in Baltimore County. The schist at these places is replaced by strings of big microcline porphyroblasts, by swarms of partially coalesced porphyroblasts, and by small bodies of potassic granite that retain ghostlike traces of the earlier schistosity. Some of the replacement bodies parallel the bedding, but others cut discordantly across. Thus the granitization which affected the underlying migmatitic gneiss did not stop at the unconformity surface but extended into the base of the Setters Formation. It evidently occurred during the migmatization and doming of the gneissic basement.

The middle member consists of impure quartzitic rocks with thin beds of inter-

TABLE 15

COMPOSITION OF SCHIST AND GNEISS

Setters Formation

Modal Analyses

	1	2	3	4	5	6	7	8	9
Quartz	31.2	22.2	14.1	8.9	58.1	45.4	53.9	40.1	34.5
Microcline	21.0	15.5	27.7	18.1	19.3	31.5	15.2	24.5	19.4
Plagioclase	4.7		1.1		2.4	7.8	5.8	5.2	7.3
Muscovite	27.0	46.7	42.3	55.2	5.6	1.9	9.0	10.9	20.1
Biotite	15.3	15.5	13.0	17.6	12.2	11.2	14.9	17.8	18.4
Clay	0.3	—	1.1	—	—	0.2	0.2	0.3	—
Tourmaline	—	t	0.4	—	—	—	—	—	t
Apatite	—	0.1	—	t	0.5	—	0.2	—	0.2
Sphene	0.1	—	—	t	t	0.1	t	0.2	—
Monazite	t	—	0.1	0.1	0.2	—	0.2	0.1	0.1
Zircon	t	t	t	t	t	t	t	0.1	t
Magnetite	—	—	0.2	0.1	1.2	1.6	0.6	0.2	—
Hematite	0.4	—	—	—	0.5	0.3	—	0.6	—
Total	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0
Points	1224	1557	1279	1224	1642	1328	1251	1220	1653

Chemical Compositions

	1	2	3	4	5	6	7	8	9	10
SiO_2	64.8	58.1	57.0	51.4	77.3	74.3	76.0	69.7	66.8	62.66
TiO_2	0.9	1.0	0.9	1.1	0.5	0.5	0.7	0.9	0.9	1.55
Al_2O_3	17.6	23.1	23.9	26.9	8.1	9.7	9.9	12.6	15.7	15.76
Fe_2O_3	1.3	0.6	0.5	0.8	2.9	3.1	1.3	2.0	0.7	0.71
FeO	3.4	3.4	2.9	3.9	3.4	3.4	3.7	4.0	4.1	3.77
MgO	1.4	1.5	1.2	1.7	1.1	1.0	1.4	1.6	1.7	1.54
CaO	0.2	0.3	0.3	0.3	0.4	0.1	0.3	0.3	0.3	1.26
Na_2O	0.8	0.4	0.6	0.4	0.4	1.0	0.8	0.8	1.0	0.82
K_2O	7.7	8.7	10.1	10.1	5.0	6.4	4.9	6.9	7.1	9.97
H_2O	1.9	2.9	2.6	3.4	0.8	0.5	1.0	1.2	1.7	1.59
P_2O_5		t			0.1		t		t	0.19
MnO										0.03
BaO										0.07
Total	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	99.92

(1) Schist from basal member (H113-5). Clarksville dome, west of Doughoregon Manor.

(2) Schist from basal member (HG1-3). Towson dome, near Hartley.

(3) Schist from basal member (H106-1). Woodstock dome, at Marriottsville.

(4) Schist from basal member (H106-2). Woodstock dome, at Marriotsville.

(5) Mica gneiss from upper member (H36-1). Clarksville dome, 1½ miles N.W. of Pine Orchard.

(6) Mica gneiss from upper member (H106-7). Woodstock dome, at Marriotsville.

(7) Mica gneiss from upper member (H113-3). Clarksville dome, west of Doughoregon Manor.

(8) Schist from upper member (H113-4). Clarksville dome, west of Doughoregon Manor.

(9) Schist from upper member (H35-3). Clarksville dome, 1½ miles N.W. of Pine Orchard.

(10) Biotite gneiss from upper member. Towson dome, Cromwell Bridge Road, 3 miles N.E. of Towson, Maryland. Analysts, Penniman and Browne, Baltimore. Taken from Knopf and Jonas, 1929, p. 157.

Column 10 is a chemical analysis. The chemical compositions in columns 1–9 were calculated from the modal analyses. The biotite composition used for the calculations is from table 14, except that the FeO/MgO ratio is adjusted to correspond to the Setters gneiss shown here in column 10 after allowing for magnetite. The muscovite composition is from Barth (1936, Table 1, col. 1a).

stratified mica schist. The quartzites are mostly of two types: microcline quartzite (Table 16, cols. 1 and 2) and muscovite quartzite (Table 16, col. 4). The first type greatly predominates. Biotite quartzite (Table 16, col. 5) is rare. The quartzite beds are separated by thin muscovite-rich partings, which probably originally were shaly partings between sandstone beds. Large tourmaline crystals are commonly concentrated along these surfaces.

Many of the quartzite beds have a delicate internal lamination (Pl. 8, fig. 1), with alternating microcline-rich layers, muscovite-rich layers, and nearly pure quartz layers. Individual laminae an eighth to half an inch thick are straight and continuous and may be traced for many feet across the face of the outcrop. The regularity and continuity of this layering suggest that it is primary bedding and not the product of metamorphic differentiation. Adjacent beds that have about the same bulk compositions are unlaminated and show no evidence of metamorphic segregation.

The upper member consists of feldspathic mica schist and finer-grained, more quartzose rock that Knopf and Jonas (1929) called mica gneiss (Table 15, cols. 5–9). Thin quartzitic layers (Table 16, col. 3) are locally interbedded with the schist and gneiss. The bulk composition of the gneiss is intermediate between that of the quartzite and schist (Fig. 20). If the quartzite and mica schist were originally quartz sandstone and shale, then the mica gneiss probably came from siltstone with an intermediate quartz and clay content.

Petrography

The mineralogy of the Setters Formation rocks is given in the modal analyses (Tables 15, 16). The Setters mica schists and quartzites are readily distinguished from those of

TABLE 16

MODAL ANALYSES OF QUARTZITE

Setters Formation

	1	2	3	4	5
Quartz	76.0	82.0	72.8	74.2	88.6
Microcline	14.7	15.3	12.4		1.7
Plagioclase		t	7.2		0.3
Muscovite	6.5	1.7	1.1	25.5	0.9
Biotite	2.1	t	5.3		7.5
Montmorillonoid	0.4	0.5			0.1
Tourmaline	t	t			0.2
Garnet					0.1
Apatite			0.3	0.1	0.6
Sphene			t		
Monazite	t		t	0.2	t
Zircon	t	t	t	t	t
Magnetite	0.3	0.5	0.9		
Total	100.0	100.0	100.0	100.0	100.0
Points	789	1612	1629	1128	1153

(1) Microcline quartzite, from middle member (H64-1). Clarksville dome, near Nichols.

(2) Microcline quartzite, from middle member (H35-4). Clarksville dome, 1½ miles N.W. of Pine Orchard.

(3) Feldspathic quartzite, from upper member (H35-2). Clarksville dome, 1½ miles N.W. of Pine Orchard.

(4) Muscovite quartzite, from middle member (B-5). Phoenix dome, west of Verona.

(5) Biotite quartzite, from middle member (H67-1). Clarksville dome, 1 mile north of Browns Bridge.

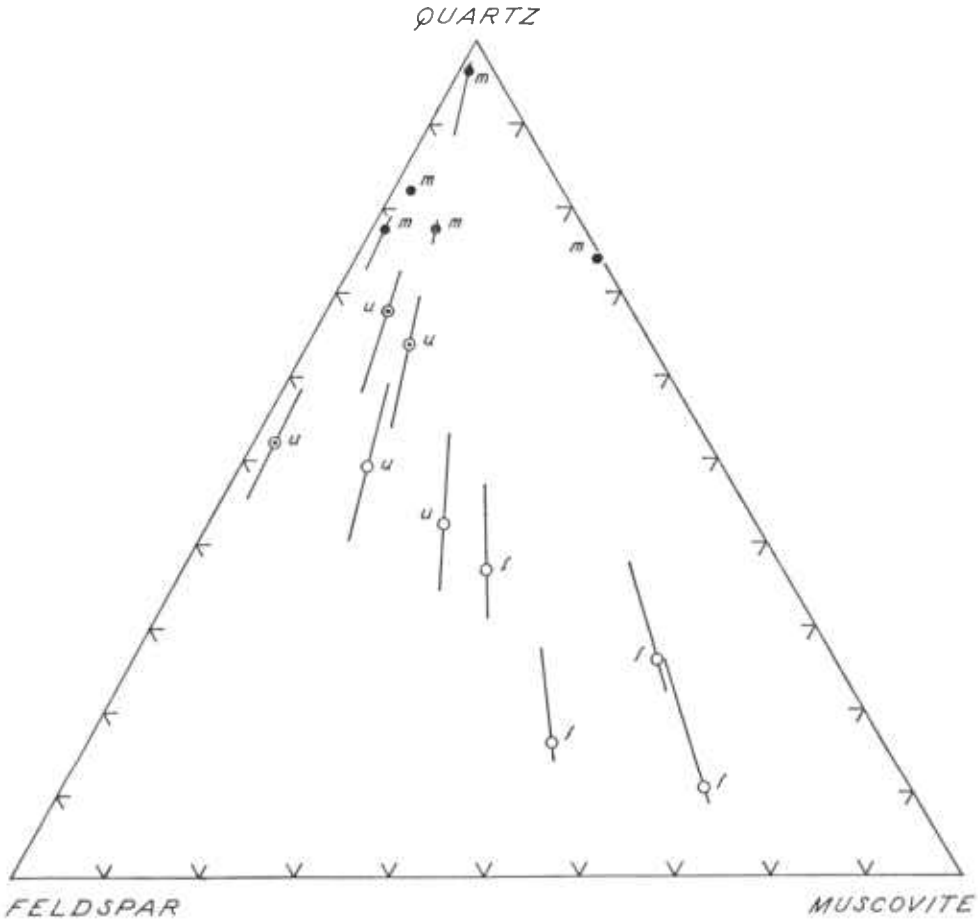


FIGURE 20. Modal composition of Setters Formation rocks. Solid circles are quartzite; circles with dots are mica gneisses; open circles are mica schists. The vertical distance between ends of the line through each point gives the percentage of biotite + minor accessories; top of the line marks percentage of feldspar + muscovite; bottom of line marks percentage of quartz. The feldspar is chiefly microcline. Rocks from the lower member (l) of the formation are argillaceous (rich in muscovite), those from the middle member (m) are quartzose, and those from the upper member (u) are about intermediate. Points are plotted from Tables 15 and 16.

the Wissahickon Formation by their mineralogy. The feldspar in the Setters is predominantly microcline, whereas in the Wissahickon it is sodic plagioclase. Garnet, staurolite, or kyanite are common in the Wissahickon mica schist but they are not found in the Setters schist. Knopf and Jonas (1929, p. 156) were wrong in stating that "... garnets are usually abundant ..." in the Setters. The upper gneiss member, in a quarry along the northern flank of the Towson dome, contains a layer about 4 inches thick consisting chiefly of very coarse garnet and mica, but otherwise garnet is rarely present, and then only as a minor accessory (*e.g.*, Table 16, col. 5).

The schist, gneiss, and quartzite are crystalloblastic, with no trace of relict primary textures. Thin blades of mica are strongly aligned in the schistose rocks and are paralleled by quartz and microcline in flat, lenticular grains and segregations. In the gneiss the micas are less strongly oriented, and the quartz and feldspars are more nearly equidimensional. The microcline is intergrown with blades of biotite and muscovite, and some

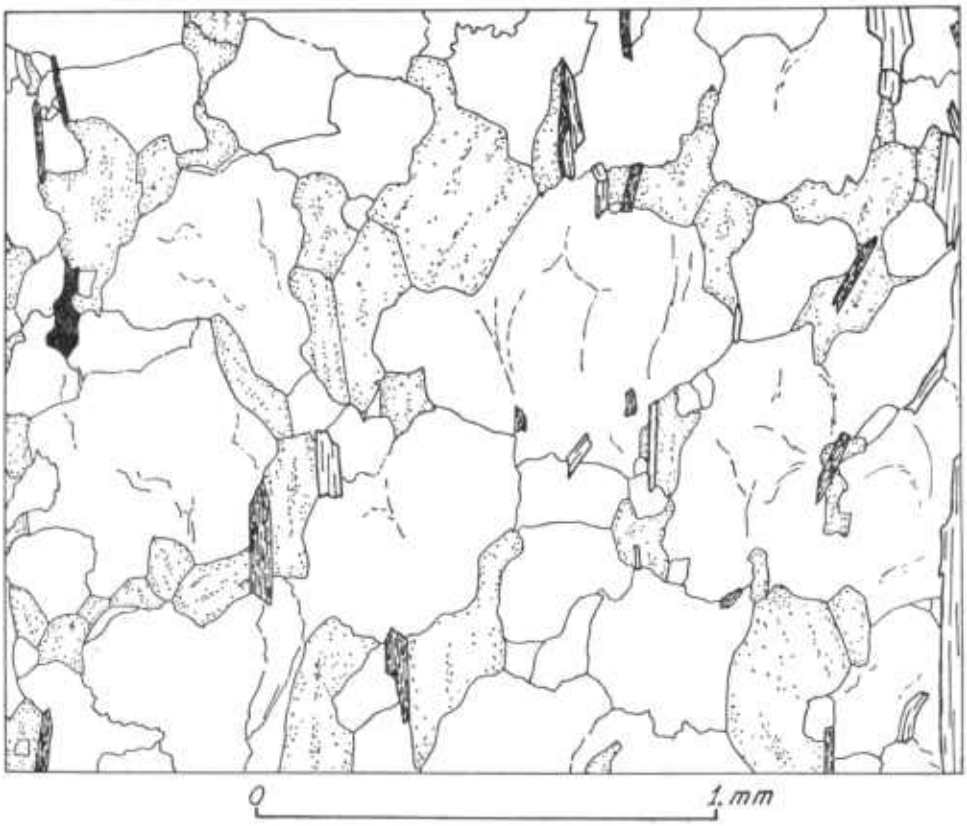


FIGURE 21. Textural relations in a feldspathic quartzite, Setters Formation. The minerals are microcline, stippled; quartz, clear; biotite, dark flakes; muscovite light flakes with heavy outlines; magnetite, solid black. The microcline and quartz are crystalloblastic, with irregular grain shapes and interlocking borders. The micas are aligned, defining a faint foliation. Drawn from thin section (H64-1).

of the larger microclines poikiloblastically enclose grains of quartz and mica. Some hand specimens of the quartzite show pink feldspars that resemble clastic grains, but in thin section they have irregular shapes and interlock intricately with the quartz (Fig. 21).

The plagioclase, which occurs sparsely in some of the gneiss and quartzite, is albite near An_0. Much of it has a "relict appearance" in thin section because of partial alteration to a brownish montmorillonite mineral. This is due to recent weathering and does not indicate that the plagioclases are relicit clastic grains. Its irregular shape and interlocking contacts with quartz and microcline show that it has recrystallized with the other minerals.

Origin

The quartzite, mica gneiss, and schist came from thinly stratified quartzose and aluminous sediments, probably sandstone, siltstone, and shale. No primary structures remain in the coarsely recrystallized and crumpled schists, but the more competent quartzitic rocks beautifully preserve the original thin sedimentary bedding. Metamorphic differentiation has produced lensoid segregations of coarse microcline in some of the schist but was not important in the quartzite and gneiss.

The quartzites came from thin-bedded quartz sandstone. The abundant muscovite

in many of the quartzite beds shows that the original sand was mixed with argillaceous material, which suggests that the sediment was not well sorted. The sandstones were probably marine or lacustrine rather than fluvial deposits because they lack such features as gravel lenses, channeling, cut-and-fill structures, and cross-bedding. The sand, with its thinly interlayered silt beds and shaly partings, was probably deposited in an environment that was protected from strong current action; this is suggested by the thin laminated bedding and by the lack of current structures. The high proportion of argillaceous material that was mixed with the sand also argues against a depositional environment in which the sand was thoroughly reworked and winnowed by currents. These sediments were not like the clean-washed, massive, or cross-bedded orthoquartzites in the Oriskany or Weverton Formations.

The feldspathic schist and mica gneiss probably came from argillaceous and silty sediments, but their unusual chemical composition poses a special problem. They have exceptionally high K_2O and low $Na_2O + CaO$ (Table 15), unlike most fine-grained clastic sediments or tuff. The compositions of average shale, arkose, and rhyolite are given for comparison (Table 17, cols. 1–3).

Three possible explanations are considered:

(1) Clastic sediments with high K/(Na + Ca) might have come from reworked saprolitic material that had developed by deep weathering of the gneissic basement. Such material would become enriched in K by leaching of Na, Ca, and Mg during prolonged chemical weathering (Goldich, 1938; Anderson and Hawkes 1958). Yet residual soils

TABLE 17

CHEMICAL ANALYSES OF SHALES, SANDSTONES, AND RHYOLITE

	1	2	3	4	5	6	7	8
SiO_2	58.10	76.37	73.66	56.73	54.66	68.40	56.29	78.73
TiO_2	0.65	0.41	0.22	0.88	1.01	0.96	0.64	0.35
Al_2O_3	15.40	10.63	13.45	19.27	20.14	14.44	19.22	10.36
Fe_2O_3	4.02	2.12	1.25	5.57	3.28	3.58	4.39	0.66
FeO	2.45	1.22	0.75	1.89	3.17	0.56		0.13
MnO		0.25	0.03		0.00	0.62		0.01
MgO	2.44	0.23	0.32	1.93	3.09	0.20	1.65	0.17
CaO	3.11	1.30	1.13	0.01	0.00	0.00	0.09	0.23
Na_2O	1.30	1.84	2.99	0.49	1.08	0.47	0.19	0.14
K_2O	3.24	4.99	5.35	8.85	9.39	7.77	10.85	8.60
H_2O	5.00	0.83	0.78	4.15	3.65	2.71	5.58	0.46
P_2O_5	0.17	0.21	0.07	0.00	0.00			0.19
CO_2	2.63	0.54		0.00	0.00			0.01
SO_3	0.64				0.10		0.72	
C	0.80							
Total	99.95	100.94	100.00	99.77	99.57	99.71	99.62	100.04

(1) Average shale (Pettijohn, 1957, p. 344, col. 1).
(2) Average arkose (Pettijohn, 1957, p. 324, col. F).
(3) Average calc-alkali rhyolite (Nockolds, 1954, p. 1012, col. II).
(4) Average of 6 slates, Cartersville Fm. (Shearer, 1918, p. 134).
(5) Potassic slate, Cartersville Fm. (Shearer, 1918, p. 142).
(6) Feldspathic sandstone, Cartersville Fm. (Shearer, 1918, p. 151).
(7) Glenwood Shale (Gruner and Thiel, 1937, p. 844).
(8) Feldspathized sandstone, Franconia Fm. (Berg, 1952, p. 222).

are rarely if ever enriched in K this much. Moreover, as K increases in such material Al increases even more; metamorphism would produce rocks with aluminum-excess minerals. It might be argued that the basement rocks were abnormally potassic to begin with; this seems unlikely, however, because the Baltimore Gneiss and other basement rocks in the central Appalachians have normal compositions, mostly corresponding to quartz diorite, granodiorite, and quartz monzonite (Fig. 14). On chemical grounds, therefore, saprolitic material is rejected as the original sediment. A similar conclusion may be extended to the quartzitic members of the Setters, some of which contain abundant K-feldspar but lack plagioclase or its breakdown products. Here it is unlikely that the original sand came from reworked weathered basement, where plagioclase is the most abundant feldspar (Fig. 14); chemical weathering would have had to destroy all plagioclase and leach away its breakdown products, while leaving abundant potassium feldspar. Although calcic plagioclase succumbs to weathering more readily than K-feldspar the basement plagioclases were relatively sodic, and it seems unlikely that even prolonged weathering would have brought about such a complete separation. The potassium feldspars in the Setters, therefore, are probably not original or recrystallized detrital grains from the weathered basement.

(2) Granitization during regional metamorphism is another possible, but rejected, explanation for the high K_2O content. Replacement dikes and other evidence for granitization are lacking, except locally in the base of the formation at the northeastern end of the Towson dome. But the formation is highly potassic nearly everywhere. It seems unlikely that the Setters could be so pervasively feldspathized without leaving widespread evidence of replacement, and without the overlying formations being affected. Thin sections lead to the same conclusion: potassium feldspar does not replace the other minerals but appears to be in equilibrium. The explanation for the high K_2O content must be sought in the original sediment.

(3) There are certain shales (Pettijohn, 1957, p. 369–370) siltstones, and sandstones that compare closely with the Setters, having exceptionally high K_2O and low Na_2O and CaO. In the shales and in the fine-grained fraction of the siltstone and sandstone K_2O may exceed 10 per cent, while combined $Na_2O + CaO$ total less than 0.5 per cent (Table 17, cols. 4–8). The clay-sized material in these sediments is chiefly authigenic orthoclase, according to Gruner and Thiel (1937) and Berg (1952), and the coarser grains are mainly quartz.

Such sediments generally occur in blanket-type deposits with thin, regular bedding, much like the Setters. For example, the Cambrian Cartersville Formation (Georgia) consists of shale and slate, feldspathic sandstone, and quartzite, in beds mostly less than a few feet thick (Shearer, 1918, p. 129–130). The Ordovician Glenwood Shale (Minnesota) consists of thin-bedded shales and sandstones (Gruner and Thiel, 1937, p. 843). The highly potassic part of the Cambrian Franconia Formation (Wisconsin, Minnesota) is ". . . a thin-bedded, shaly, micaceous sandstone . . ." (Berg, 1952, p. 221).

The potassium of these rocks is chiefly from very fine-grained orthoclase of authigenic origin. According to Berg (1952, p. 223):

"The feldspathized sandstone was deposited in a marine environment of quiet, shallow water. Muds and clayey to silty sands accumulated, and during lithification mineralogical changes took place. Authigenic orthoclase was formed in the sandstone and hydromuscovite (illite) developed in the shale. The original sediment contained a small amount of detrital feldspar which served as nuclei for precipitation of orthoclase overgrowths, but at the same time authigenic crystals formed in the pore spaces and filled the interstices between the detrital quartz grains. Clay minerals originally deposited with the clastic

quartz and feldspar probably provided a large part of the alumina and silica, and sea water supplied the potassium."

The Setters Formation probably came from potassic mud, silt and sand like that in the Cartersville, Glenwood, and Franconia Formations. Metamorphism that formed the present microcline-rich schist, gneiss, and quartzite was isochemical. The conditions which brought about K-enrichment in the original sediments are unknown.

COCKEYSVILLE MARBLE

Name

The Cockeysville Marble was named by G. H. Williams (1892a), after the small community of Cockeysville, 10 miles north of Baltimore. He described it as (Wilmarth, 1938, p. 478):

"Highly crystalline dol. with all its original impurities separated in the form of well-defined minerals Extensively quarried at Cockeysville, Md. Overlies Setters quartz schist and underlies gneiss [Wissahickon fm.]".

Since that time the marble has been studied by W. J. Miller (1905), Mathews and Miller (1905), Knopf and Jonas (1929), and Choquette (1957; 1960). Choquette studied both the structure and petrology of the Cockeysville in detail, and much of the information given here is taken from his 1960 paper.

Knopf and Jonas (1929, p. 166) noted the resemblance of the Cockeysville to the marble in the Doe Run-Avondale district in southeastern Pennsylvania and suggested that they might be equivalent. This correlation now seems well established (Bascom and Stose, 1932; Cloos and Hietanen, 1941; Swartz, 1948; McKinstry, 1961; Gray, 1960).

Occurrence

In Maryland the Cockeysville Marble is limited to Baltimore and Howard Counties, where it rims the gneiss domes. The marble weathers and erodes readily, so that it underlies the valleys, while the more resistant Setters and Wissahickon formations hold up the adjacent ridges. The marble is most extensive and best exposed in Baltimore County. In Howard County it appears to be thinner and is missing in many places on the flanks of the domes (Cloos and Broedel, 1940). The marble is best exposed in Howard County at Marriottsville (Woodstock dome), at Alberton (north end of Clarksville dome), in the valley 1 mile west of Carrolls Mill and 2 miles north of Clarksville (between the Clarksville and Mayfield domes), and at Nicols (southwest tip of Clarksville dome). It is also exposed at Browns Bridge Road near Lime Kiln Road south of Highland, in the stream valley 1 mile northeast of Alpha, and along the Patapsco River half a mile southeast of Davis.

The local thinning and disappearance of the Cockeysville around the flanks of the gneiss domes was attributed to sedimentary overlap onto irregular basement topography by Knopf and Jonas (1923, p. 45–46). This may be true for the Setters Formation, but probably not for the Cockeysville, because (1) the marble is most commonly missing where a continuous blanket of Setters already mantles the basement (*e.g.*, Mayfield and Woodstock domes), and (2) the marble does not contain intercalations of clastic detritus washed in from local islands of basement rock. Choquette (1960, p. 1032) suggests more plausibly that the marble was thinned and pinched off by deformation:

"The reason for these pinch outs is uncertain, but they occur within such short distances that they may be structural rather than stratigraphic, perhaps mainly because of flowage induced by differential move-

ment over the gneiss domes. Direct evidence that the carbonate rocks were extremely mobile lies in the flowage and drag folds found at almost every outcrop in the area."

The thinning and local disappearance of the Cockeysville might also be explained by uplift and erosion of the deposit before the Wissahickon was laid down. Choquette (1960, p. 1031; Fig. 2) notes a slight unconformity between the Cockeysville and Wissahickon, which would support this possibility. Perhaps both erosion and tectonic thinning were important factors.

Thickness

A related problem is the thickness of the marble. It ranges from 0 to 3000 feet or more in Baltimore County (Choquette, 1960, p. 1032), but this does not represent the original stratigraphic thickness. The original thickness can only be roughly estimated, because it has been modified to an unknown but probably large degree by flowage and minor folding. Also, part of the formation may have eroded away before it was covered by the Wissahickon. In Baltimore County Knopf and Jonas (1923, p. 46; 1929, p. 166) estimated the maximum stratigraphic thickness to be about 400 feet. Choquette (1960, p. 1032) put it at about 750 feet but cautioned that at least a 30 per cent error in the estimate was very possible.

In Howard County the marble appears to become thinner, as is evident from the greatly diminished average width of its outcrop belt. At Marriottsville, on the west flank of the Woodstock dome, the apparent stratigraphic thickness is about 250 feet. At Nichols, at the southwestern extremity of the Clarksville dome, the apparent thickness is only about 175 feet. These figures are approximate, but they may not be far from the true stratigraphic thickness, because at those places the beds dip 40° and 5°, respectively, and deformation was evidently not great.

Lithology

Metadolomite, calc-schist, and calcite marble are most abundant, while calc-gneiss and calc-silicate marble are widespread but minor. According to Choquette (1960, p. 1031) metadolomite and interlayered calcite marble make up about 80 per cent of the formation around the Phoenix and Texas domes, but in areas farther south and southwest calc-schist becomes progressively more abundant. This trend appears to hold through Howard County, where metadolomite and calc-silicate rocks predominate.

Choquette notes also that the carbonate rocks are most impure near the base and top of the formation. The basal Cockeysville, which conformably overlies Setters mica gneiss, is commonly a quartz-rich diopside-tremolite calc-gneiss and may represent a transition from noncarbonate to carbonate sedimentation. At the upper contact, where it is exposed 2 miles east of Pikesville in Baltimore County, Cockeysville calc-schist is succeeded upward by biotite-microcline paragneiss, tremolite metadolomite, and then by garnet-mica schists of the Wissahickon Formation (Choquette, 1960, p. 1032). The upper and lower contacts of the Cockeysville Marble were not observed in Howard County.

The metadolomite is composed of 80 per cent or more dolomite, 0–20 per cent calcite, plus minor phlogopite and other minerals. Two varieties are common: phlogopite-calcite metadolomite associated with calcite marble, and quartz metadolomite associated with calc-silicate marble. Much of the metadolomite occurs in relatively competent layers that may represent original bedding. The metadolomites were probably derived from calcareous and argillaceous dolostones. The chemical compositions of the metadolomites are given in Table 18.

TABLE 18

CHEMICAL COMPOSITION OF MARBLE AND CALC-SILICATE ROCKS

Cockeysville Marble

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16
SiO_2	0.44	5.57	3.88	2.2	0.8	2.2	0.7	—	17.0	39.9	43.7	53.6	54.1	64.6	67.5	77.9
Al_2O_3	1.22	0.40	1.24	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Fe_2O_3	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
FeO	tr.	n.d.	0.32	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
MgO	20.87	20.30	20.41	18.3	21.6	18.8	21.1	21.6	19.8	16.1	17.1	17.6	14.2	14.7	13.7	9.3
CaO	30.73	29.08	30.06	31.7	29.8	32.7	30.9	30.1	29.1	25.1	26.8	25.4	26.1	20.5	18.4	11.8
MnO	1.22	n.d.	0.02	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Na_2O	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
K_2O	n.d.	n.d.	n.d.	0.8	0.6	0.3	0.1	0.4	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
$H_2O^{\pm}$	n.d.	n.d.	n.d.	0.5	0.5	0.2	0.1	0.2	0.1	1.2	1.2	0.3	1.0	0.2	0.4	1.0
FeS_2	n.d.	n.d.	0.06	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
CO_2	45.85	44.26	43.93	44.7	46.1	45.1	47.1	46.8	33.8	15.1	11.2	3.1	4.6	—	—	—
Other	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Total	100.33	99.61	99.92	98.2	99.4	99.3	100.0	99.1	99.8	97.4	100.0	100.0	100.0	100.0	100.0	100.0
CaO:MgO	1.47	1.43	1.48	1.72	1.39	1.74	1.41	1.39	1.47	1.56	1.57	1.44	1.84	1.38	1.34	1.26

Chemical Analyses

(1) Metadolomite, locality unknown. T. E. Whitfield, analyst (*in* Knopf and Jonas, 1929a, p. 165).
(2) Metadolomite, locality unknown. E. A. Schneider, analyst (*in* Knopf and Jonas, 1929a, p. 165).
(3) Metadolomite, locality unknown. W. F. Hunt, analyst (*in* Knopf and Jonas, 1929a, p. 165).

Chemical Compositions, calculated from modal analyses.

(4) Metadolomite interlayered with calcite marble, 3½ miles southwest of Butler, Maryland, near State highway 128 (*in* Choquette, 1960, p. 1043).
(5) Metadolomite relic in calcite marble quarry at Texas, Maryland (*in* Choquette, 1960, p. 1043).
(6) Metadolomite interlayered with calcite marble, Ruxton, Maryland (*in* Choquette, 1960, p. 1043).
(7) Metadolomite, 2 miles southeast of Butler, Maryland (*in* Choquette, 1960, p. 1043).
(8) Metadolomite, 12 feet from contact with diabase dike, ¾ mile southwest of Butler, Maryland (*in* Choquette, 1960, p. 1043).
(9) Diopside metadolomite, 1¼ miles northeast of Towson, Maryland (*in* Choquette, 1960, p. 1043).
(10) Calcite-tremolite rock, 1½ mile north-northeast of Towson, Maryland (*in* Choquette, 1960, p. 1043).
(11) Dolomite calc-silicate rock, ½ mile north-northeast of Stevenson, Maryland, along State highway 130 (*in* Choquette, 1960, p. 1043).
(12) Tremolite-diopside calc-silicate rock, 1 mile north of Glenarm, Maryland (*in* Choquette, 1960, p. 1043).
(13) Calcitic quartz-diopside calc-silicate rock, 2⅓ miles east-northeast of Towson, Maryland (*in* Choquette, 1960, p. 1043).
(14) Quartz-diopside calc-silicate rock, 1¼ miles northeast of Stevenson, Maryland (*in* Choquette, 1960, p. 1043).
(15) Quartz-diopside calc-silicate rock, 1¼ miles north-northeast of Glenarm, Maryland (*in* Choquette, 1960, p. 1043).
(16) Diopside-quartz calc-silicate rock, 2 miles northwest of Ruxton, Maryland (*in* Choquette, 1960, p. 1043).

The metadolomite commonly contains a few scattered porphyroblasts of diopside and tremolite, but locally these are concentrated, forming layers, lenses, or pods of calc-silicate marble and granulite. Good examples are found near Marriottsville, especially in the hills north of the river. Choquette (1960, p. 1042) found that the calc-silicate marbles were restricted to the quartz metadolomite in the upper part of the Cockeysville Formation. He suggests that they formed by dedolomitization (chiefly loss of CO_2) of quartz dolostone or cherty dolostone. Chemical compositions are given in Table 18.

The calc-schist and calc-gneiss are fine- to medium-grained light-gray to bluish-gray marbles consisting of calcite plus 50 per cent or more of other minerals, mainly phlogopite, muscovite, quartz, microcline, plagioclase, and pyrite, but no dolomite. The calc-schist is micaceous and strongly foliated, whereas the calc-gneiss is rich in quartz and has a massive or weakly gneissic texture. Both are intimately associated with impure calcite marble, into which they interfinger, interlense, and intergrade (Choquette, 1960, p. 1038). Much of the small-scale layering and lensoid structure is probably due to metamorphic differentiation. The calc-schist and calc-gneiss were probably derived from argillaceous limestone and calcareous shale.

The calcite marble is white, gray, or bluish medium- to coarse-grained nondolomitic marble of which 85 per cent or more is calcite. In Baltimore County the Cockeysville Formation contains massive units of calcite marble 50 feet or more thick, but in Howard County it occurs mainly as thin layers, lenses, pods, and small discordant patches in the metadolomite and calc-schist. The fairly continuous marble layers a foot to several feet thick may be original beds, but the discontinuous layers and lenses a few inches or less thick are probably metamorphic segregation features. Choquette (1960, p. 1044–1049) gives convincing evidence that some of the discordant bodies of calcite marble formed by replacement.

Petrography

The common mineral assemblages in rocks of the Cockeysville Marble are as follows:

Calcite marble: Chiefly calcite, with minor phlogopite, muscovite, quartz, microcline, and calcic plagioclase. Pyrite and sphene are minor accessories.

Metadolomite: Chiefly dolomite, with minor calcite, phlogopite, and quartz. Pyrite is a common accessory.

Calc-schist: Calcite, phlogopite, muscovite, quartz, microcline, and sodic plagioclase. Small amounts of tremolite, diopside, and scapolite may be present. Minor accessories include pyrite, sphene, apatite, tourmaline, and graphite.

Calc-gneiss: Calcite and quartz, with lesser amounts of phlogopite, muscovite, microcline, and sodic plagioclase. Pyrite, graphite, tourmaline, sphene, apatite, scapolite, and clinozoisite are possible accessories.

Calc-silicate rocks: These span a wide composition range and may contain any of the minerals listed above. Diopside, tremolite, calcite, and quartz—occurring in various combinations—generally predominate. Microcline is very abundant in some.

The marbles are thoroughly recrystallized, and the minerals are commonly segregated into streaks or thin bands. In the calc-schists, for instance, thin calcite-quartz layers commonly alternate with phlogopite-muscovite-apatite-sphene-pyrite layers. The feldspar and quartz grains generally have interlocking contacts with the other minerals, and it is doubtful that any relict clastic grains survive.

In general the mineral assemblages appear to be in equilibrium, but in a few rocks

there is evidence of retrogression. Fibrous masses of tremolite replace diopside in a few rocks, although more commonly they appear to have grown stably together. Talc forms an alteration product on diopside and tremolite and occurs in microcrystalline felts along fractures and cleavages in the porphyroblasts (Choquette, 1960, p. 1042).

Origin

The original sedimentary rocks were dolostone, limestone, sandy to argillaceous dolostones and limestones, and calcareous shale. Some of the carbonate rocks may have been cherty. In Baltimore and Howard Counties the carbonates appear to become more argillaceous and arenaceous to the south or southwest, which may indicate a facies change to near-shore deposits in that direction. The formation also thins in that direction, but it is uncertain whether this represents thinning of the original deposit, pre-Wissahickon erosion of the deposit, or, less likely, tectonic thinning.

During Glenarm metamorphism the Cockeysville sediments were intensely deformed and completely recrystallized to new mineral assemblages. The carbonate rocks were incompetent, relative to the Setters and Wissahickon formations on either side, and were deformed to a large extent by plastic flowage. The formation was greatly thickened by flowage and minor folding in the synclines between the gneiss domes and was probably thinned along their flanks. Choquette (1960, p. 1049) notes that there is intense minor folding in the interior of the formation but that it decreases toward the upper and lower contacts, showing that the minor folding was largely intraformational.

The recrystallization was mainly syntectonic, for a strong foliation has developed. Locally the calc-silicate rocks record a minor retrogressive recrystallization, with the development of tremolite and talc; however, this has not been related to any new discordant structures.

WISSAHICKON FORMATION

Stratigraphic Relations

The Wissahickon Formation is a thick sequence of metamorphosed argillaceous and sandy sedimentary rocks that lie above the Cockeysville Marble. In Maryland the Wissahickon crops out in two belts, separated by the Sykesville Formation in Montgomery and Howard Counties and by the Peters Creek Formation in Baltimore and Harford Counties and adjacent parts of Pennsylvania. The eastern belt, which passes near Baltimore, is coarsely crystalline schist, whereas the much wider western belt is chiefly fine-grained schist and phyllite. Knopf and Jonas (1923; 1929) designated the two belts as the oligoclase-mica schist facies and the albite-chlorite schist facies, because of their difference in metamorphic grade. In this report, however, they are simply called the eastern and western sequences, because metamorphic grade does not always distinguish them. In southern Montgomery County, for instance, rocks of the "low-grade" western sequence locally attain the sillimanite zone.

Stratigraphic relations between the eastern and western sequences have been the subject of conflicting interpretations. The problem will be briefly reviewed.

Sixty years ago Bascom (1902) gave the name Wissahickon gneiss to the belt of mica schist and gneiss that extends from near Philadelphia, Pennsylvania, southwestward into Cecil County, Maryland. The type locality is along Wissahickon Creek, a tributary of the Schuylkill River. This belt of metasedimentary rocks is cut off in Cecil County by intrusive masses of granodiorite and gabbro, but it reappears along strike near the Susquehanna River and continues through the Baltimore area, where it was named the

Wissahickon formation by E. B. Mathews (1904; 1905). Mathews' Wissahickon formation, however, included not only the coarsely crystalline schists that form the continuation of the Philadelphia-Cecil County belt, but also all of the phyllitic rocks that extend for many miles across strike to the west. He states (1905, p. 335):

"The Wissahickon formation in Maryland occurs as a broad band wrapping about the older Baltimore gneiss and limestone areas and occupying practically all the territory between the Blue Ridge and the Coastal Plain deposits except those portions consisting of the sedimentary rocks already described [Setters and Cockeysville], the overlapping Triassic sandstones of Carroll, Frederick, and Montgomery Counties and the igneous rocks."

Mathews noted that ". . . the more crystalline Wissahickon schists pass by gradations [westward] into the less crystalline phyllites and sericite schists." Little was known, however, about their actual stratigraphic relations, for he says (1905, p. 335):

"The line of separation between these two divisions [schists and phyllites] is not marked stratigraphically by any pronounced change in lithologic character, and it is not entirely clear that the division is a stratigraphic one, although it seems probable that the phyllites in all instances are of as high or higher horizons than the more crystalline Wissahickon schists."

In Pennsylvania Bascom (1905) also expanded the Wissahickon to include the fine-grained mica schists which bordered the Paleozoic rocks flanking Mine Ridge and the Honeybrook Upland. These schists appeared to grade lithologically into the Wissahickon mica gneiss on the southeast, but here also the stratigraphic relationship was obscure.

Thus in 1905 nearly all the pelitic gneiss, schist, and phyllite above the basement rocks in the Pennsylvania and Maryland Piedmont were lumped into the Wissahickon. It was considered Ordovician, because the fine-grained schist of the South Valley Hills in Pennsylvania appeared to overlie conformably the Cambro-Ordovician limestones of Chester Valley (Bascom 1905). Also, in Maryland the coarsely crystalline schist near Baltimore rested upon Cockeysville Marble which Mathews (1905) correlated with the Cambro-Ordovician carbonates.

In 1909, however, Bascom became convinced that the Wissahickon mica gneiss near Philadelphia was Precambrian, chiefly because of its greater metamorphism and because it is cut by plutonic rocks which are absent from the phyllites and from near-by Paleozoic rocks in the Honeybrook Upland. The schist and phyllite of the South Valley Hills, which she still regarded as Ordovician, were separated from the mica gneiss and renamed the Octoraro schist (Bascom *et al.*, 1909). Because of their supposed age difference the Octoraro and the Wissahickon were inferred to be in fault contact, although they appeared to be gradational in the field:

"The mica gneiss [Wissahickon] on the northwest flank of Buck Ridge is adjacent to the Octoraro schist, from which it is distinguished with great difficulty. It becomes increasingly like the mica schist in the neighborhood of that formation, losing completely the character which characterizes the rock elsewhere." (Bascom *et al.*, 1909, p. 4).

In the Doe Run-Avondale district Bliss and Jonas (1916) followed Bascom and separated fine-grained "Octoraro" schist from the coarser-grained Wissahickon schist and mica gneiss by a fault contact, although here again the rocks appeared to be gradational.

These conclusions were soon changed. Along the Susquehanna River at the Peach Bottom syncline Knopf and Jonas (1923) found that the schists underlying the Cardiff Conglomerate contained a higher proportion of quartzitic beds than they did elsewhere.

These more arenaceous schists, which overlapped the boundary formerly drawn between the Wissahickon and Octoraro schists, were renamed the Peters Creek schist. Its relation to the Wissahickon was described as follows (Knopf and Jonas, 1923, p. 48):

> "Since the Peters Creek schist on the southeast side of the Peach Bottom syncline grades downward into an oligoclase-mica schist and since the Peters Creek schist on the northwest side of the syncline grades downward into an albite-chlorite schist it is considered that the albite-chlorite schist is the stratigraphic equivalent of the oligoclase-mica schist and that they are different mineralogic facies of the same formation, the Wissahickon."

The term Octoraro Schist, which had included rocks that Knopf and Jonas now placed partly in the Wissahickon albite-chlorite facies and partly in the Peters Creek Schist, was dropped.

This interpretation won wide acceptance and has become so firmly entrenched that Swartz (1948, p. 1509), reviewing the literature, concluded that it ". . . may be accepted as a basic feature of the geology of the schists."

The structural and stratigraphic relations in southern Maryland, nevertheless, do not fit this picture. No trace of the Peach Bottom syncline—which had been postulated as a major structure extending through Maryland and deep into Virginia (Stose and Stose, 1948)—has been found. The eastern Wissahickon, dipping westward off the gneiss domes, grades upward into the boulder-bearing Sykesville Formation which in turn grades into the western Wissahickon sequence. This section appears to be homoclinical, and the strata become progressively younger to the west. This is shown by graded bedding in the western Wissahickon which indicates, despite local reversals due to small folds, that the tops of the beds lie predominantly to the west (G. W. Fisher, 1963, and personal observation). Other features further argue that the eastern and western Wissahickon are not the same rocks, repeated on opposite limbs of a syncline: for instance, the western sequence is considerably thicker than the eastern and contains a higher proportion of sandy beds. Some of these beds, in the western sequence, were medium- to coarse-grained sand, whereas those in the eastern sequence nearly all appear to have been fine-grained sand and silt. Moreover, graded bedding is common in the western Wissahickon but has not yet been found in the eastern. The eastern and western Wissahickon do not, therefore, appear to have been the same sequence of sediments, unless there was an abrupt change in thickness and facies.

These points, especially the younging direction of the graded bedding, indicate that the succession: eastern Wissahickon-Sykesville-western Wissahickon is homoclinal rather than synclinal.

A problem of nomenclature then arises. The top of the Wissahickon Formation was never defined by the early workers, but since 1923, following Knopf and Jonas, it has been placed at the base of the Peters Creek Formation (in southern Maryland this would correspond approximately to the base of the Sykesville). There may be strong justification for continuing to follow this precedent, but it would mean excluding the western sequence (the albite-chlorite schist facies of Knopf and Jonas) from the Wissahickon Formation and giving it a new name, for it now appears to be stratigraphically higher. For many years, however, the western sequence of schists have been called Wissahickon by the Maryland Geological Survey (Mathews and Johannsen, 1904; Knopf and Jonas, 1925; Jonas, 1928; Jonas and Stose, 1938; Cloos and Broedel, 1940; Cloos and Cooke, 1953). Moreover, there are areas in which the Peters Creek Formation is not distinctive, as in Baltimore County, and where the Sykesville Formation pinches out, as in southern Montgomery County, leaving an unbroken section of Wissahickon-type rocks. If the

top of the Wissahickon were placed in this interval it would be an artificial boundary in the midst of a continuous mica schist-metagraywacke sequence. For the present, therefore, the western sequence will be included in the Wissahickon Formation, and its top tentatively placed at the base of the Ijamsville Phyllite.

WISSAHICKON FORMATION, EASTERN SEQUENCE

Occurrence

The eastern Wissahickon sequence crops out over a wide area in Howard and Baltimore counties where it surrounds the mantled gneiss domes, but to the southwest and northeast it narrows and plunges beneath younger formations. In Howard and Montgomery Counties the eastern Wissahickon is bounded on the west and east by the Sykesville and Laurel formations respectively, both of which lie stratigraphically above it. Along the Patuxent River the eastern Wissahickon forms a band 10 miles wide, but it narrows to the south and disappears at Washington, D. C., where the Sykesville and Laurel formations converge at the nose of the southward-plunging Baltimore anticlinorium.

West of Kensington the eastern and western Wissahickon sequences come directly in contact for a short distance, where the Sykesville Formation is interrupted. They are similar lithologically and evidently gradational. Abundant small lenses of Sykesville rock pervade both sequences in this area. Farther south the eastern and western Wissahickon are separated by mafic intrusive rocks of the Georgetown Complex.

In southeastern Howard County the Laurel Formation is cut off by intrusive rocks of the Baltimore Gabbro Complex. From there northward through Howard and Baltimore Counties the gabbro complex bounds the Wissahickon on the southeast.

The eastern Wissahickon, which consists largely of micaceous schists and semischists, weathers readily and is very poorly exposed. It is easily mapped, nevertheless, because the schist disintegrates to saprolite and soil made distinctive by glistening flakes of muscovite. Fresh exposures of the schist are mostly found in the larger stream valleys, especially along the Patapsco, Patuxent, and Potomac Rivers and their tributaries. Occasional good exposures are also found in the uplands in Howard County and adjoining parts of Baltimore County.

The contact between the eastern Wissahickon and the overlying Sykesville and Laurel Formations is gradational and conformable. Its basal contact with the Cockeysville and lower formations may be disconformable (Choquette, 1960), but the evidence is inconclusive. An older interpretation, which still persists, is that the Wissahickon is in thrust contact with the lower formations. This idea was first advanced by Bliss and Jonas (1916) to explain the superposition of the Wissahickon, which they then supposed was Precambrian, upon marble [Cockeysville] in the Doe Run-Avondale district of Pennsylvania which they correlated with the Cambro-Ordovician limestone of Chester Valley. Miller and others (1917, p. 3, 11) accepted this interpretation and placed a thrust between the Wissahickon schist and the lower formations northeast of Baltimore. In both areas, however, the thrust contact was not based on field evidence but was postulated to explain the supposed geologic age difference of the schist and marble. Soon afterward, however, Knopf and Jonas (1923) repudiated the thrust hypothesis, and for many years it remained in discard. Then it was once more revived (Stose and Stose, 1946, p. 83), although for a different reason:

"In New England the Taconic sequence belongs to a different basin of sedimentation from that in which the carbonate sequence was deposited. If the same conditions of sedimentation obtained in Pennsylvania and Maryland, the Wissahickon formation and associated rocks of the crystalline schist series

have come to lie on the Cockeysville marble, Setters formation, and Baltimore gneiss in the Maryland region by overthrusting which took place in post-Cambrian time."

As supporting evidence they (1946, p. 83) note that:

"In many places in Baltimore County the Wissahickon formation overlaps the Cockeysville marble onto the Setters formation and in places onto the Baltimore gneiss."

This apparent overlap, however, can be explained stratigraphically or by tectonic movement involving only small displacement. The Cockeysville and Setters (1) may never have been deposited over those places where they are now missing; (2) may have been partly eroded before the Wissahickon was deposited; or (3) may have been thinned and pinched off tectonically during rise of the gneiss domes. Where the contact is exposed (Choquette, 1960, p. 1031–1032) there is no evidence of thrusting. Moreover, the bedding in the Wissahickon and the underlying formations is approximately concordant wherever they crop out close together. The Glenarm Series is now known to be older and unrelated to the Taconic sequence, so that the structural and stratigraphic relations of one cannot be deduced from the other. In the Baltimore-Washington area, at least, the Wissahickon thrusting hypothesis is rejected.

Thickness

The original thickness of the Wissahickon is difficult to estimate because of small-scale folding and other tectonic complications. Bascom *et al.* (1909, p. 4) estimated the thickness at about 1000–2000 feet in the Philadelphia area, but the top of the sequence there is not exposed. In the Baltimore region the thickness of the eastern Wissahickon can be estimated along the west flank of the Woodstock and Mayfield gneiss domes, where the sequence dips steeply to the west and its bottom and top (taken at the belt of serpentinite that bounds the Sykesville Formation) are well defined. The apparent stratigraphic thickness along the South Branch of the Patapsco River, where the Wissahickon dips off the Woodstock dome, is approximately 5500 feet. Farther south, where it dips off the Mayfield dome opposite Glenelg, the apparent thickness is about 6500 feet. These estimates indicate only the order of magnitude of the stratigraphic thickness, because the amount of tectonic distortion is unknown. They may be too small, since the measurements were made at the steeply dipping limbs of the gneiss domes, where the incompetent Wissahickon strata were more likely stretched and tectonically thinned than repeated by close folding. Most of the individual beds in this area, however, do not appear to be greatly distorted.

Lithology

Thinly interstratified pelitic schist, semipelitic schist, and finely crystalline granulose psammitic rocks rich in quartz and feldspar compose the bulk of the eastern Wissahickon sequence. Amphibolite and calc-silicate rocks, also in thin beds, total considerably less than 1 per cent. Concordant sheetlike bodies of serpentinite, chlorite-actinolite schist, and talc-carbonate schist are remnants of ultramafic igneous rocks that intruded the Wissahickon and were metamorphosed with it.

The pelitic schists are lustrous, medium- to coarse-grained rocks rich in muscovite, biotite, and sometimes chlorite. The micas are generally well aligned, forming the pronounced schistosity. Garnet porphyroblasts, commonly up to 6 mm in diameter, are conspicuous in many of the schists, and large stubby staurolite prisms may also be present. Near the gneiss domes these minerals may be accompanied by kyanite, some-

times occurring in bladed porphyroblasts up to 3 inches long but more often in tiny grains visible only under the microscope.

The semipelitic schists have more quartz and feldspar and less mica than do the pelites, but there are all gradations between them. Some contain garnet but lack staurolite and kyanite. Texturally the semipelites are finer-grained and show less segregation of quartz and mica. As the schistosity and mineral segregation become even weaker, and the proportion of mica decreases still more, the semipelitic schists pass into granulose quartzo-feldspathic and quartzitic rocks. Thus the semipelites are rocks of intermediate composition and texture which link the pelitic and psammitic beds.

The psammitic rocks are fine-grained, granulose to weakly schistose rocks rich in quartz and plagioclase. Some of them are quartzites, with quartz as the chief mineral, but more commonly oligoclase, biotite, and muscovite are also important constituents, and quartz may total less than 50 per cent. Previous authors have described the latter rocks as mica gneiss; however they lack the conspicuous segregation of light and dark minerals and the medium- to coarse-grained texture generally implied by the term gneiss. Harker's (1939, p. 246–248) term psammitic granulite is more appropriate.

Original bedding is generally well preserved and easily recognized where psammitic and pelitic rocks crop out together (Pl. 8, fig. 2), although it is rarely apparent in the pelitic schist alone. The psammitic and semipelitic rocks occur in beds that range from half an inch to 4 feet or more thick, but they are normally less than 8 inches thick. These beds are regular and uniform except where tightly folded; even the very thin beds may be traced across the outcrop with little change in thickness. The pelitic schists that are interstratified with psammitic beds range from thin partings to beds tens of feet thick.

The ratio of pelitic to psammitic rocks varies greatly from place to place, generally in the range from 2:1 to 10:1. In general the proportion of psammitic beds appears to increase higher in the stratigraphic section. The basal part of the eastern Wissahickon sequence is almost entirely pelite. This is best seen in a long road cut along the Jones Falls Expressway just west of Bare Hills (Baltimore County), where a continuous 500-foot stratigraphic section through the schist, just above its contact with the Cockeysville Marble, reveals a monotonous succession of garnet and staurolite schists with less than a dozen thin psammitic beds. The top of this pelitic section is not exposed, so that its total thickness may have been much greater. In contrast, the upper part of the eastern Wissahickon sequence, where it borders the Sykesville and Laurel Formations, displays a fairly high proportion of psammitic beds.

Schistosity is well developed in the pelitic and semipelitic rocks but is weak or lacking in many of the psammitic layers. Where only one generation of schistosity is visible it almost invariably parallels the bedding, even around the crests of folds. Where the beds are strongly folded, however, a discordant axial-plane cleavage is usually superimposed; this cleavage crosscuts and crinkles the earlier bedding schistosity. This cleavage first appears only in the pelitic layers, as widely spaced shear planes near the crests of folds. With more intense deformation the cleavage planes become more closely spaced and have newly recrystallized micas aligned along them. At this stage the discordant cleavage also appears in the interstratified psammitic beds, although here it is more widely spaced and less pervasive. With still further deformation and recrystallization a through-going axial-plane schistosity pervades all beds and nearly obliterates the original bedding schistosity.

The bedding schistosity is commonly the dominant metamorphic structure, but there

is little to indicate that deformation was necessary for its development. It appears to have preceded the strong deformation, for it is best shown in flat or weakly folded strata. Where the folds become tight a newer axial-plane cleavage has overprinted the bedding schistosity. This structure is probably mimetic after bedding, not tectonic.

Folds of an entirely different character, unrelated to an axial-plane cleavage, are also found in the Wissahickon. Some psammitic beds have been pulled apart and the segments then rumpled up like a rug, producing small-scale folds of great complexity (Fig. 22). Yet the enclosing strata above and below these chaotic zones—which are generally only a foot or a few feet thick—are straight and relatively undeformed. This fact, together with the wild, irregular style of folding and the lack of any related axial-plane cleavage, suggests that this deformation was not tectonic but due to slumping of soft sediments. Slumping is further suggested by the blurred contacts between the psammitic and pelitic beds in the strongly rumpled zones, as if the material had mixed together before the beds were firmly consolidated.

Slump folds and other phenomena associated with soft-sediment deformation are uncommon in the eastern sequence, but in the western sequence slumping and sliding were widespread.

Petrography

The pelitic schists consist chiefly of quartz, muscovite, plagioclase, biotite, chlorite, and garnet, with apatite, magnetite, tourmaline, and occasionally epidote, allanite, monazite, zircon, and xenotime as minor accessories. The more aluminous schists contain porphyroblastic staurolite in addition to garnet, and near the gneiss domes kyanite may be present as well. Modal analyses of some typical pelites are given in Table 19. Sillimanite was found at only one locality—in strongly pegmatized schist flanking the western edge of the Clarksville gneiss dome, near Carrolls Mill in Howard County. There the sillimanite occurs as tiny stout prisms and as fibrolite, associated with but possibly later than, kyanite and garnet. Biotite appears to be the normal brown variety but has a rather high magnesium to iron ratio (see Table 23) for pelitic schist (Foster, 1960, Fig. 12).

The psammitic rocks consist chiefly of quartz and plagioclase, with lesser biotite and muscovite. Garnet and epidote or clinozoisite may be sparingly present. The common accessories are magnetite, apatite, and tourmaline. Trace amounts of allanite, zircon, monazite, xenotime, rutile, and other unidentified high-relief minerals may also be present. Modal analyses of some typical psammitic rocks are given in Tables 20 and 21.

The feldspar in both the pelitic and psammitic rocks is chiefly oligoclase, chiefly An_{18-28}. In the most Ca-poor schists the plagioclase is more albitic. Potassium feldspar is extremely rare, except as minute granules that accompany the alteration of biotite to chlorite. Approximately 60 stained thin sections of eastern Wissahickon rocks were examined, and potassium feldspar (microcline) was present as a primary constituent in only 2 of them.

The pelitic schists and most of the psammitic granulites are crystalloblastic, with no trace of relict primary textures. Quartz and feldspar form a mosaic of irregular interlocking grains, interspersed with flakes or wispy segregations of mica. The scarcity of relict detrital grains in the psammitic rocks cannot be entirely ascribed to the severity of metamorphism, for psammitic rocks in the western sequence that are of equal or even higher metamorphic grade (sillimanite zone), and just as strongly deformed, preserve

FIGURE 22. Folds in Wissahickon schist produced by slumping of soft sediments. The strata above and below the folds are straight and continuous. Stippling denotes sandy beds (psammitic granulite), and the lined pattern pelite (garnet-staurolite schist). Patuxent River, near Snell Bridge.

TABLE 19

COMPOSITION OF PELITIC SCHIST

Wissahickon Formation (Eastern Sequence)

	1	2	3	4	5	6	7
Modal Analyses							
Quartz	46.2	38.1	35.6	31.6	28.2	28.7	21.1
Plagioclase	10.1	18.6	9.5	19.1	22.2	5.2	
K feldspar				2.2	t		
Muscovite	25.9	22.4	37.4	24.9	20.1	37.8	63.6
Biotite	5.5	13.6	6.3	9.2	10.3	6.2	0.2
Chlorite	3.3	1.8	1.4	7.6	2.8	4.8	5.1
Garnet	2.2	1.0	6.9	0.3	8.0	9.1	8.6
Staurolite	3.6	t	1.6		1.8	5.7	0.7
Kyanite	0.4		0.1		2.0		
Epidote	0.2			0.6			
Allanite		0.2		0.2	0.3		
Apatite	0.5	0.2	0.3	t	0.5	0.2	t
Tourmaline	0.2		0.1	0.2	0.1	0.7	0.1
Monazite	t	t			t		
Zircon		t	t	t	t		t
Magnetite	1.8	4.1	0.8	4.1	3.7	1.6	0.6
Hematite	0.1						t
Total	100.0	100.0	100.0	100.0	100.0	100.0	100.0
Points	1,316	1,643	1,501	1,553	1,692	1,258	1,458
Chemical Composition, Calculated from the Modes							
SiO_2	66.3	62.1	61.9	58.8	56.2	54.9	53.7
TiO_2	0.3	0.4	0.4	0.4	0.4	0.4	0.2
Al_2O_3	18.0	16.3	21.1	18.6	20.2	24.3	28.1
Fe_2O_3	2.4	5.3	1.1	5.3	4.5	2.0	0.7
FeO	4.5	5.5	5.6	5.7	8.5	8.3	5.7
MgO	1.3	2.1	1.3	2.2	1.9	1.8	1.0
CaO	1.1	1.3	1.0	1.2	1.7	0.8	0.5
Na_2O	1.0	1.6	1.0	1.8	1.9	0.7	0.5
K_2O	3.0	3.4	4.2	3.6	2.8	4.1	6.1
H_2O	1.9	1.9	2.3	2.4	1.7	2.6	3.5
P_2O_5	0.2	0.1	0.1	t	0.2	0.1	
Total	100.0	100.0	100.0	100.0	100.0	100.0	100.0

(1) Kyanite-garnet-staurolite-mica schist (H27-5). Wards Chapel Road near Marriottsville Road, Baltimore County.

(2) Garnet-biotite-muscovite schist (M459-1). Patuxent River below Browns Bridge, Montgomery County.

(3) Kyanite-staurolite-garnet-mica schist (H139-4). Three-tenths mile north of Browns Bridge, Howard County.

(4) Garnet-biotite-muscovite schist (H27-4). Wards Chapel Road near Marriottsville Road, Baltimore County.

(5) Staurolite-kyanite-garnet-mica schist (H27-6). Wards Chapel Road near Marriottsville Road, Baltimore County.

(6) Staurolite-garnet-biotite-muscovite schist (M-258f). Patuxent River below Snell Bridge, Howard County.

(7) Staurolite-garnet-muscovite schist (H23-2). Tunnel Road 0.7 miles south of Henryton, Howard County.

TABLE 20

COMPOSITION OF METAGRAYWACKE

Wissahickon Formation (Eastern Sequence)

	1	2	3	4	5	6	7	8
Modal Analyses								
Quartz	48.3	47.8	46.8	44.3	44.4	42.2	30.7	25.4
Plagioclase	30.3	33.1	28.8	29.0	36.6	42.7	35.3	32.0
K feldspar							t	
Muscovite	6.8	1.7	10.2	12.9	6.7	0.1	15.2	27.2
Biotite	11.6	16.1	10.8	9.6	8.8	11.1	15.2	11.7
Chlorite	t			0.1			t	t
Garnet			0.7	2.7	t		t	1.1
Epidote	0.4	0.2			1.5	0.3	t	
Allanite					t	0.1	t	
Apatite	0.2	0.1	0.5	0.4	0.1	0.3	0.7	0.4
Tourmaline		t			t			
Monazite	t	t	t	t	0.1		t	t
Zircon	t	t	t	t	t	t	t	t
Magnetite	2.4	1.0	2.2	0.9	1.8	3.2	2.9	2.2
Hematite		t		0.1				t
Total	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0
Points	1,601	1,191	1,662	1,585	1,366	1,496	1,767	1,546
Chemical Compositions, Calculated from the Modes								
SiO_2	72.2	73.4	70.9	70.8	72.0	70.3	63.0	60.6
TiO_2	0.2	0.3	0.3	0.3	0.2	0.2	0.3	0.4
Al_2O_3	12.0	11.8	13.1	14.5	13.3	12.2	16.5	19.9
Fe_2O_3	3.3	1.6	2.9	1.5	2.7	4.3	3.9	2.9
FeO	3.5	3.5	3.6	3.7	2.7	3.8	4.3	3.9
MgO	1.6	2.2	1.5	1.4	1.2	1.5	2.1	1.6
CaO	1.8	1.9	2.0	1.8	2.4	2.4	1.9	1.7
Na_2O	2.6	2.8	2.4	2.5	3.2	3.6	3.3	3.1
K_2O	1.8	1.7	2.0	2.2	1.5	1.0	2.9	3.8
H_2O	0.9	0.8	1.0	1.1	0.8	0.5	1.5	1.9
P_2O_5	0.1	t	0.3	0.2	t	0.2	0.3	0.2
Total	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0

(1) Metagraywacke (H27-2). Wards Chapel Road near Marriottsville Road, Baltimore County.
(2) Metagraywacke (H72-1). Mackenzie Road, 1.8 miles west of Rockland, Howard County.
(3) Metagraywacke (M459-2). Patuxent River below Browns Bridge, Montgomery County.
(4) Metagraywacke (H107-3). Patapsco River above Marriottsville, Baltimore County.
(5) Metagraywacke (H75-1). Near West Friendship, Howard County.
(6) Metagraywacke (H72-3). Mackenzie Road, 1.8 miles west of Rockland, Howard County.
(7) Volcanic graywacke (H27-3). Wards Chapel Road near Marriottsville Road, Baltimore County.
(8) Volcanic graywacke (M-258C). Patuxent River below Snell Bridge, Howard County.

abundant relict grains with clastic outlines. However, most of the clastic grains that survived there are larger than about $\frac{1}{4}$ mm; this suggests that the rocks of the eastern sequence were originally finer-grained—possibly silts and very fine-grained sands.

The effects of metamorphic differentiation are plainly visible in the more coarsely crystalline schists. Quartz has segregated into thin lenses and irregular knots up to several inches long. Micas and chlorite have segregated into thin laminae. These and

TABLE 21

COMPOSITION OF METASUBGRAYWACKE AND QUARTZITE

Wissahickon Formation (Eastern Sequence)

	1	2	3	4	5	6	7	8
Modal Analyses								
Quartz	54.4	60.5	59.4	61.5	76.5	81.4	84.5	87.0
Plagioclase	31.6	23.4	22.3	17.6	8.2	13.3	5.4	0.2
K feldspar							3.8	
Muscovite	1.7		8.4	14.0	12.4	3.1	0.6	10.6
Biotite	10.7	9.9	6.3	4.8	1.4	1.2	4.6	1.8
Chlorite			t		t			
Garnet		0.3	3.0	0.5	0.3		t	
Epidote		3.6				0.2	0.4	
Allanite	t			t	t		t	
Apatite	0.1	0.3	0.1	t	0.5	0.1		0.2
Monazite		0.2	t	t	t	t	t	t
Zircon	t	t	t	t	t	t	t	t
Magnetite	1.5	1.8	0.5	1.6	0.7	0.7	0.7	0.2
Hematite					t			
Rutile			t					
Total	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0
Points	1,500	1,459	1,529	1,503	1,637	1,527	903	1,703
Chemical Compositions, Calculated from the Modes								
SiO_2	77.0	77.8	78.8	78.9	86.7	90.8	91.3	91.9
TiO_2	0.2	0.2	0.2	0.2	0.1	t	0.1	0.1
Al_2O_3	10.2	8.6	10.8	10.6	7.3	4.7	3.4	4.6
Fe_2O_3	2.1	3.0	0.8	2.1	0.9	1.0	1.0	0.3
FeO	2.8	3.0	3.1	2.1	0.8	0.6	1.3	0.5
MgO	1.5	1.4	0.9	0.7	0.2	0.2	0.6	0.3
CaO	1.6	2.5	1.3	0.9	0.8	0.8	0.5	0.2
Na_2O	2.8	2.0	2.0	1.7	0.8	1.2	0.5	0.1
K_2O	1.2	0.9	1.4	1.9	1.4	0.4	1.1	1.3
H_2O	0.6	0.5	0.7	0.9	0.7	0.2	0.2	0.6
P_2O_5	t	0.1	t		0.3	0.1		0.1
Total	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0

(1) Metasubgraywacke (H107-1). Patapsco River above Marriottsville, Baltimore County.

(2) Metasubgraywacke (H72-5). Mackenzie Road, 1.8 miles west of Rockland, Howard County.

(3) Metasubgraywacke (H139-1). Three-tenths mile north of Browns Bridge (Patuxent River), Howard County.

(4) Metasubgraywacke (M-258b). Patuxent River below Snell Bridge, Howard County.

(5) Micaceous quartzite (M-258a). Patuxent River below Snell Bridge, Howard County.

(6) Feldspathic quartzite (H27-1). Wards Chapel Road at Marriottsville Road, Baltimore County.

(7) Feldspathic quartzite (H23-3a). Tunnel Road, 0.7 miles north of Henryton, Howard County.

(8) Micaceous quartzite (H23-1). Tunnel Road, 0.7 miles south of Henryton, Howard County.

other metamorphic segregation phenomena are usually quite distinct from the relict bedding, which is on a larger scale and appears as distinct layers of alternating pelitic and psammitic bulk composition (Pl. 8, fig. 2).

The thin calcareous beds consist chiefly of granoblastic quartz, clinozoisite, and minor

biotite (Table 22). The thin layers of amphibolite consist mainly of green hornblende and plagioclase (sodic andesine), with minor epidote and other accessories (Table 22). No relict textures or structures in these rocks can be discerned.

The planar structures are closely related to the microscopic textures of the bedded rocks. Where only bedding schistosity is present the textures suggest that the metamorphic minerals grew under fairly static conditions. Two lines of evidence are as follows:

(1) Garnet porphyroblasts occur as spongy skeletons (Pl. 9, fig. 1) and as well-formed idioblasts and include traces of quartz, mica, and fine opaques which preserve the trend of the enclosing matrix. These internal trend lines are useful for establishing the time relations between the growth of the porphyroblasts and movement within the rock. In those rocks with only bedding schistosity the garnets have straight trend lines,

TABLE 22

COMPOSITION OF AMPHIBOLITE AND CALC-SILICATE ROCKS

Wissahickon Formation (Eastern Sequence)

	1	2	3
Modal Analyses			
Quartz	2.2	68.9	86.4
Plagioclase	31.4		
Hornblende	57.8		
Biotite			3.0
Chlorite	0.3		
Epidote	6.6		
Clinozoisite		30.0	10.0
Apatite	0.2	0.1	
Sphene	1.5	1.0	0.6
Total	100.0	100.0	100.0
Points	1,433	696	170
Chemical Composition, Calculated from the Modes			
SiO_2	50.7	77.9	90.0
TiO_2	1.1	0.5	0.4
Al_2O_3	14.2	11.0	4.4
Fe_2O_3	3.0	0.9	0.4
FeO	5.8		0.6
MgO	8.8		0.4
CaO	12.2	8.9	3.2
Na_2O	2.5		
K_2O	0.2		0.3
H_2O	1.4	0.7	0.3
P_2O_5	0.1	0.1	
Total	100.0	100.0	100.0

(1) Epidote-andesine amphibolite (M459-3). Patuxent River below Browns Bridge, Montgomery County.

(2) Clinozoisite-quartz granulite (H23-36). Tunnel Road, 0.7 miles south of Henryton, Howard County.

(3) Biotite-clinozoisite quartzite (H23-3C). Tunnel Road, 0.7 miles south of Henryton, Howard County.

aligned parallel to the enclosing schistosity (Pl. 9, fig. 1), showing that they grew in place without rotation by penetrative movement. The delicate skeletal form of some of the garnets (Pl. 9, fig. 1) also suggests lack of movement during growth. Where the bedding schistosity has been strongly crinkled or cut by later cleavage the garnet trend lines are still nearly straight, showing that the garnets grew before the crinkling, but they are rotated out of alignment with the enclosing schistosity (Pl. 9, fig. 2). This indicates that the garnets were forcibly twisted *after* their growth was complete, presumably by the forces that caused the crinkling or crosscutting cleavage. Some of the staurolite porphyroblasts show similar relationships, although their internal trend lines are rarely well developed. Garnets or other porphyroblasts with strongly curving or spiral trend lines, which indicate growth *during* movement, have not been found.

(2) The rocks with only bedding schistosity generally lack lineation. Elongate minerals, such as staurolite or hornblende, are randomly oriented when the rock is viewed perpendicular to the schistosity plane. This suggests that these minerals grew statically, or when penetrative movement was too weak to cause grain alignment. Where a later cleavage cuts the bedding schistosity, however, a prominent lineation is formed by the intersection of the two structures or by the axes of the small crinkles. In some cases recrystallization has also caused new mineral alignment.

Thus it appears that the bedding schistosity is early and formed under fairly static conditions, by recrystallization parallel to bedding. Shearing and pervasive internal movement later superimposed lineations and discordant cleavages, and mechanically twisted the garnets. The deformation is described in detail by Cloos in another section of this report.

In many of the schists the discordant cleavages formed while the grade of metamorphism was still high: new biotites have grown along the slip planes; aluminum-excess minerals remain unaltered; granulated quartz and feldspar have thoroughly recrystallized; and bent and broken micas have reannealed, leaving no trace of strain shadows. In other schists, however, the cleavage and other shear structures are accompanied by cataclasis without recrystallization, or by retrogressive mineral alterations. Here the deformation evidently occurred much later, when temperatures were lower.

Mineralogical evidence of retrogressive metamorphism is most evident in the pelitic schists. The chief effects are chloritization of biotite, alteration of garnet to aggregates of sericite, chlorite, and iron oxide (Pl. 10, fig. 1), and conversion of staurolite and kyanite to shimmer aggregates consisting mainly of sericite (Pl. 10, fig. 2). Sericitization of feldspar is also common. Where there has been strong shearing the original coarse flakes of muscovite are milled down to streaky aggregates of fine sericite. The most strongly retrograded rocks are completely reduced to quartz-chlorite-sericite schist. More commonly, however, retrogression has not gone so far, and relics of the earlier minerals survive (Pl. 10).

The retrogressive metamorphism is widespread but not pervasive in the eastern Wissahickon sequence. It was localized along certain zones, especially where movement and shearing were most intense.

Chemical composition.—The approximate chemical composition of typical eastern Wissahickon metasediments, calculated from the modal analyses, are given in Tables 19–22. The mineral compositions used in the calculations are as follows: chemically analyzed eastern Wissahickon biotite (Table 23, no. 1); muscovite from pelitic schist in Duchess County, N. Y. (Barth, 1936, Table 1, no. 1a); average Wissahickon chlorite [$Mg_2 \cdot Fe_{2.2}Al_{2.6}Si_{2.7}O_{10}(OH)_8$], determined by Fisher (1963, p. 128). All plagioclases

TABLE 23

CHEMICAL ANALYSES OF BIOTITE

Wissahickon Formation

	1	2	Specimen data
SiO_2	35.96	36.09	(1) Kyanite - staurolite - almandine schist (H143-1), Wissahickon eastern sequence. Wards Chapel Road at Marriottsville Road, Baltimore County. O. von Knorring, analyst.
TiO_2	1.40	2.40	
Al_2O_3	18.54	17.40	
Fe_2O_3	1.58	4.47	
FeO	16.73	16.39	
MnO	0.06	0.36	
MgO	12.53	9.23	
CaO	tr	0.54	(2) Garnet - chlorite - biotite - albite - quartz granulite (meta-subgraywacke) (M421-1), Wissahickon western sequence. Hawkins Creamery Road, 0.6 mile southwest of Etchison, Montgomery County. O. von Knorring, analyst.
Na_2O	0.20	0.24	
K_2O	8.49	8.10	
H_2O^+	3.94	4.29	
H_2O^-	0.06	0.08	
F	0.38	0.34	
	100.29	99.93	
$-O \equiv F$	0.02	0.14	
Total	100.27	99.79	

Structural Formulae

$$(1)\quad \underbrace{[(Al_{0.33}Ti_{0.08}\overset{+0.47}{Fe^{'''}_{0.09}Fe^{''}_{1.05}Mg_{1.40}Mn_{0.01}})}_{2.96}\ \overset{-1.31}{(Si_{2.69}Al_{1.30})O_{10}}\ (OH)_2]^{-0.84}\ \overset{+0.84}{(K_{0.81}Na_{0.03})^{+0.84}}$$

$$(2)\quad \underbrace{[(Al_{0.30}Ti_{0.14}\overset{+0.41}{Fe^{'''}_{0.26}Fe^{''}_{1.04}Mg_{1.04}Mn_{0.02}})}_{2.80}\ \overset{-1.26}{(Si_{2.74}Al_{1.26})O_{10}}\ (OH)_2]^{-0.85}\ \overset{+0.84}{(K_{0.78}Na_{0.04}Ca_{0.02})^{+0.84}}$$

were determined optically on the universal stage. The standard formulae of the other minerals were used.

Engel and Engel (1958, p. 1380–1382) have shown that chemical compositions of rocks calculated from reliable modes may approach the accuracy of standard chemical analyses. The rock compositions calculated here may be at least as good as some of the older chemical analyses of Wissahickon rocks cited in the literature, and the proportions of K_2O to Na_2O are probably more accurate. In any event, they are sufficiently accurate for constructing Figures 23 and 24, which is their main purpose.

Origin

The mica schists, garnet schists, and staurolite and kyanite schists, which together constitute the bulk of the eastern Wissahickon sequence, were derived chiefly from shales. This is shown by their bedded structure and chemical composition (Table 19), especially their high alumina/alkali and potassium/sodium ratios, and relatively low silica content.

The granulites are also of sedimentary origin, as evidenced by their thin interstratification with the pelitic metasediments. Previous authors believed they came from sandstones, and it was inferred that many of them were arkosic because of their high feldspar content (Knopf and Jonas, 1929, p. 173–194; Weiss, 1949, p. 1697–1698). Yet graywackes and tuffaceous sediments also may be quite feldspathic (Pettijohn, 1957).

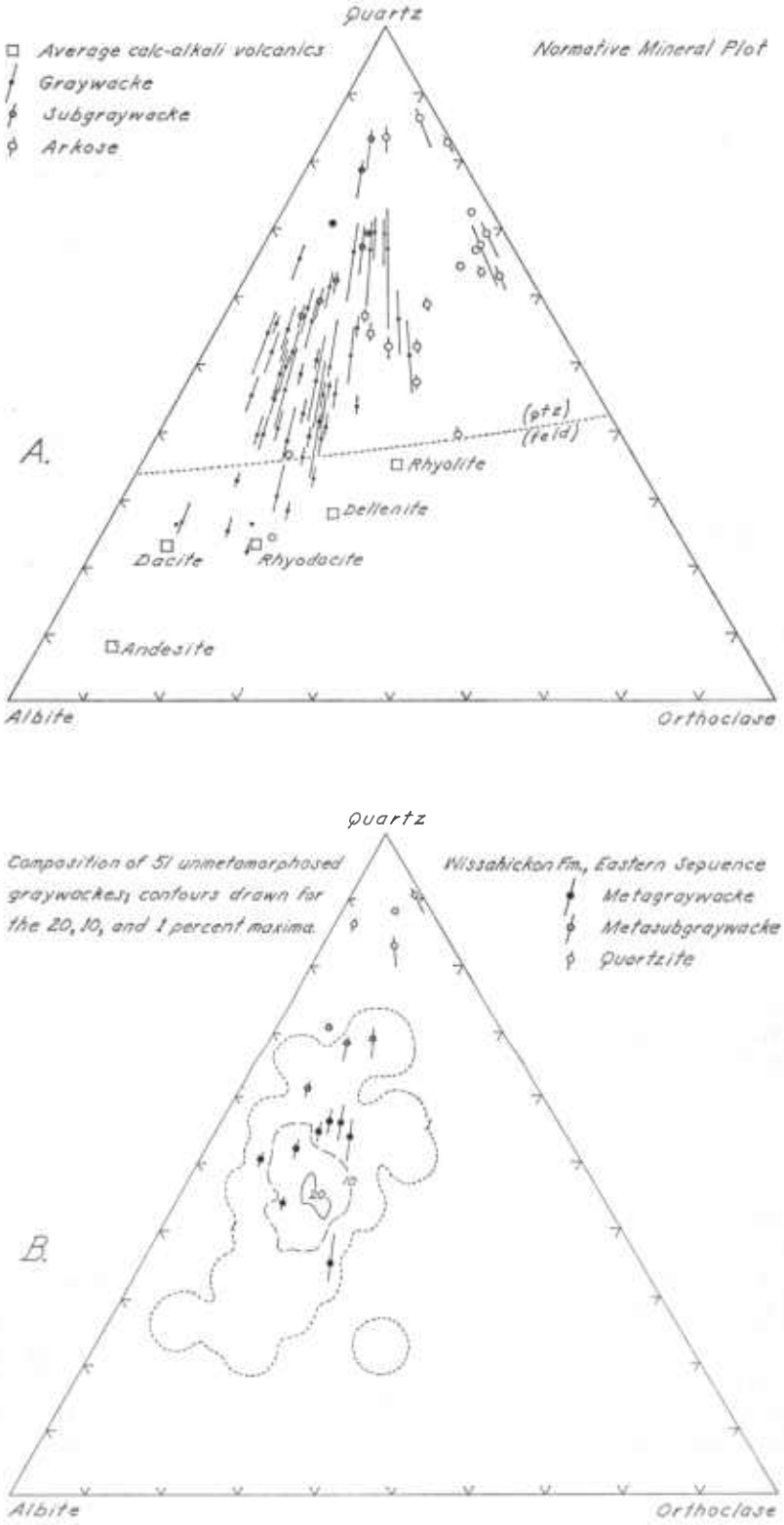

Quartz
Average calc-alkali volcanics
Graywacke
Subgraywacke
Arkose
Normative Mineral Plot
A.
(qtz)
(feld)
Rhyolite
Dellenite
Dacite
Rhyodacite
Andesite
Albite
Orthoclase
Quartz
Composition of 51 unmetamorphosed graywackes; contours drawn for the 20, 10, and 1 percent maxima.
Wissahickon Fm., Eastern Sequence
Metagraywacke
Metasubgraywacke
Quartzite
B.
Albite
Orthoclase

FIG. 23

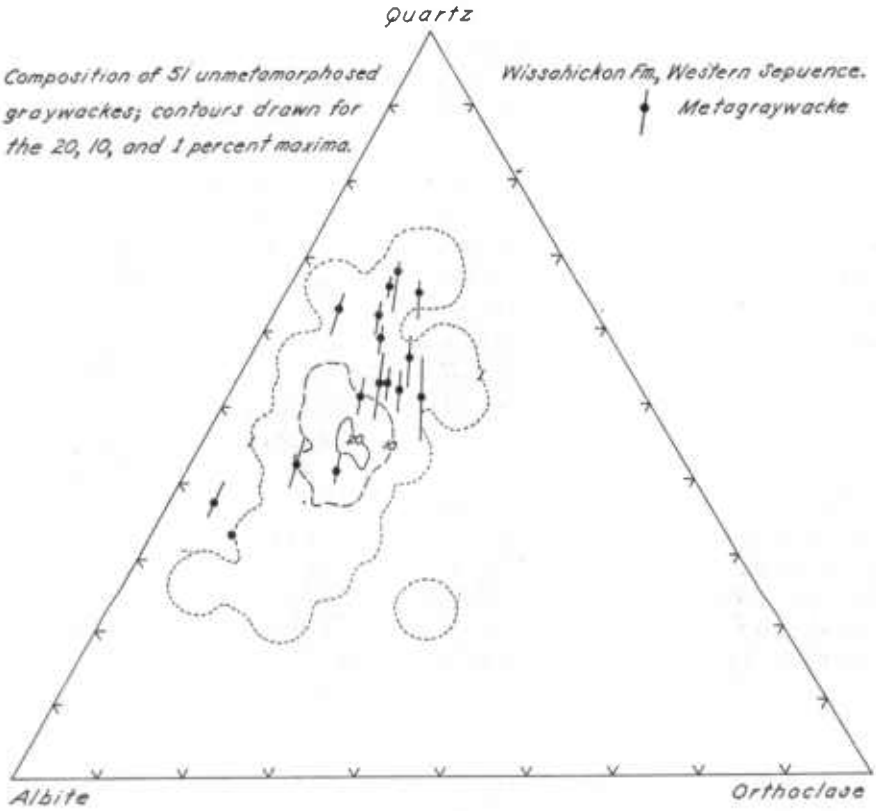


FIGURE 24. Composition of typical psammitic rocks from the Wissahickon Formation, western sequence. Normative quartz, albite, and orthoclase at the corners; normative corundum given by the steep line through each point. Shown also is the contoured graywacke field, taken from Figure 23.

Arkoses, graywackes, and bedded rocks of purely volcanic derivation may be to a large extent separated from one another by differences in chemical composition. Especially critical are their relative proportions of K, Na, Al, and Si. These differences are brought out by plotting their normative (CIPW) orthoclase, albite, quartz, and corundum on a triangular diagram (Fig. 23A). With normative orthoclase and albite at the base of the triangle the K/Na ratio (molecular) for any plotted point can be estimated by how far it lies to the left or right. With normative quartz at the top of the triangle the relative amount of excess silica can be estimated from the distance of each point above

FIGURE 23. Composition of eastern Wissahickon psammitic rocks compared with various possible parents. Normative minerals (CIPW) are plotted at the corners. Normative corundum is shown by the steep line through each point; percentage is read from the vertical distance between the ends of each line, measured on the scale at the sides.

A. Normative compositions of average calc-alkali volcanic rocks (Nockolds, 1954), 51 graywackes, 8 subgraywackes, and 19 arkoses (from F. J. Pettijohn, personal communication, 1963). Dashed line shows position of the thermal valley, projected from the liquidus surface, in the dry system albite-orthoclase-silica (Bowen, 1937). Volcanic rocks plot mostly below this line, sandstones mostly above it. Graywackes and subgraywackes lie mostly left of the center (sodic side), arkoses to the right of center (potassic side).

B. Normative composition of typical psammitic rocks of the Wissahickon Formation, eastern sequence. The curved lines denote a graywacke field, made by contouring the 51 graywackes from 23A. Contours drawn for the 20, 10, and 1 per cent maxima.

the base. Normative corundum—a measure of a rock's excess alumina—is shown as a steep line through each point, following the plotting method of Johannsen (1931, Fig. 114). The amount of normative corundum is proportional to the vertical distance between the ends of the line and may be read from the scale on the sides of the triangle (tick marks at 10 per cent intervals).

Plotted in Figure 23A are average calc-alkali volcanic rocks from Nockolds (1954) and unmetamorphosed sandstones from analyses loaned to me by Prof. F. J. Pettijohn. The sedimentary rocks consist of 51 graywackes, 8 subgraywackes, and 19 arkoses. Despite some overlapping it is evident that the various sandstones and volcanic rocks plot in different parts of the field. The volcanic rocks are separated from most of the sandstones by the dashed line, which is a projection from the liquidus surface of the thermal valley that separates the crystallization fields of quartz and feldspar, in the dry synthetic system albite-orthoclase-silica (Bowen, 1937, Fig. 7). Volcanic rocks normally plot below this line, whereas most of the sandstones plot above it (the few sandstones that plot below this line in Figure 23A are mostly volcanic graywackes from New Zealand). The graywackes and subgraywackes lie on the left side of the triangle, whereas most of the arkoses fall to the right. This follows from the well-known fact that most graywackes have Na/K ratios greater than 1, whereas this ratio is less than 1 for most arkoses (Pettijohn, 1963, Fig. 2). The graywackes, with abundant argillaceous matrix, are further distinguished from arkoses by their abundant excess alumina, shown on the diagram by their long "corundum tails."

In Figure 23B the 51 graywackes shown in 23A have been contoured like a petrofabric diagram. The outer contour encloses all points and defines a "graywacke field"; the inner contours show the area of greatest graywacke density. The psammitic granulites from the eastern Wissahickon, plotted as points on Figure 23B, fall mostly inside this field. The only way in which many of them differ from typical graywacke appears to be their slightly lower corundum. This could reflect a lower proportion of original argillaceous matrix, and hence slightly better sorting.

As already shown, the Wissahickon granulites were evidently derived from clastic sediments with the following characteristics: (1) composition of graywacke and subgraywacke; (2) fine original grain size; (3) thin but continuous beds, intimately interstratified with greater amounts of shale; (4) local slump structures. On the other hand, the granulites lack primary structures indicative of fluvial or shallow-water origin, such as discontinuous lensing beds, cross-bedding, cut-and-fill structures, or conglomerate lenses—features that could not have been entirely obliterated by metamorphism. Thus, the granulites were probably marine silts and fine-grained but poorly sorted sands, deposited intermittently with muds in fairly deep water.

The impure quartzites (Fig. 23B; Table 21, nos. 6, 7, 8), which are also thinly interstratified with the pelitic schists, correspond chemically to highly siliceous shales (*see* Pettijohn, 1957, Table 71) or impure chert (Pettijohn, 1957, Table 82). Their occurrence in thin, laterally continuous beds, and the absence of relict detrital sand grains favor such an origin. Some other thin siliceous beds rich in clinozoisite (Table 22, nos. 2, 3) may have come from calcareous cherty sediments.

The origin of the sparse amphibolite beds in the eastern Wissahickon is not clear. They are too thin for lava flows, but their high Na/K and Fe/Mg ratios argue against shaly dolomitic sediment. They were probably derived from thin sheets of basaltic aquagene tuff (Carlisle, 1963).

Thus, the eastern Wissahickon sequence appears to have been predominantly shales

with thin but abundant interbeds of silt and fine graywacke sand. Subordinate thin beds of impure chert or siliceous shale were also present. Shale greatly predominated toward the base of the sequence, but stratigraphically higher the proportion of sandy beds increased. The entire sequence was probably marine sediments, deposited in fairly deep water, judged by the overall lithologic character of the sequence, the thin but even bedding, the presence of slump structures, and the absence of any features suggesting current action.

WISSAHICKON FORMATION, WESTERN SEQUENCE

Occurrence

The western sequence forms a broad northeast-trending belt of metasedimentary rocks that extends from Fairfax County, Virginia, through Maryland and into southeastern Pennsylvania. These rocks occupy many square miles in western Howard County and central Montgomery County (Pl. 7). They are exceptionally well exposed along the gorge of the Potomac River, which affords a natural cross section through the sequence. A valuable study of this section has recently been made by G. W. Fisher (1963). In Howard County good exposures of the schists are found along the South Branch of the Patapsco River. The uplands between the Potomac and Patapsco Rivers, however, are deeply weathered, and exposures of fresh rock there are small and widely scattered.

In Howard County and in Montgomery County as far south as Rockville the eastern border of the sequence is in gradational contact with the Sykesville Formation, and the succession is inferred to lie stratigraphically above the Sykesville. The contact marks a change from massive granitic-appearing rock (Sykesville) to pelitic schist with thin psammitic interbeds. Southeast of Rockville the Sykesville grades laterally into rocks of Wissahickon lithology, so that locally the western and eastern Wissahickon sequences come directly in contact. Here there appears to be a gradation from one to the other. From Beane southward to the Potomac River the mafic rocks of the Georgetown Complex intervene and form a convenient although arbitrary eastern boundary for the western sequence.

Along its western margin the Wissahickon western sequence is in contact with rocks shown as Ijamsville Phyllite on the Geologic Map of Montgomery County (Cloos and Cooke, 1953; *see also* Plate 7 of this report). This contact, from Annapolis Rock southward to Brownstown in Montgomery County, marks a change from pelitic schist with thin, regular beds of fine-grained metasandstone and metasiltstone to distinctive silvery and greenish-blue phyllites. The contact was well chosen, not only because it is a readily mapped break between rocks of contrasting lithologies, but because it also marks the westernmost extent of metasedimentary rocks having purely Flysch characteristics. This contact can be followed northward through Howard County, passing just west of Florence and Woodbine. It is poorly exposed and not so distinct there, however, and was not originally shown on the Geologic Map of Howard County. The contact can also be traced farther south. At Brownstown it does not double back to the northwest, as shown on the Montgomery County map, but continues southward to the Potomac River, crossing just west of Blockhouse Point. Graded bedding and cleavage-bedding relationships at Blockhouse Point, according to Fisher (1963, Pl. 1), indicate that the Wissahickon beds are overturned and become younger to the west. This suggests that the Wissahickon lies stratigraphically beneath the adjacent phyllite (Ijamsville), provided the two are in normal contact.

West of this contact a large northeast-trending re-entrant of Wissahickon Formation, overlain by Triassic rocks in the vicinity of Boyds, is shown on the Montgomery County geologic map (also Pl. 7, this report). The map pattern suggests that this re-entrant is a plunging fold, bringing to the surface Wissahickon rocks belonging to the main belt on the east. This is not the case, however. The rocks of the re-entrant resemble Wissahickon because they have abundant metasandstone beds. However, the metasandstones are unlike the thin-bedded, fine-grained Flysch-type psammites that characterize the main Wissahickon belt; many of them are coarser-grained and thicker-bedded, and occasionally one finds a bed of well-sorted, cleanly washed orthoquartzite—a lithology common in the molassic rocks farther west, but unknown in the Wissahickon to the east. The Boyds reentrant, therefore, appears to represent a different sedimentary facies and should perhaps be excluded from the Wissahickon Formation.

Thickness

Poor exposures and structural complications make it difficult to even approximate the stratigraphic thickness of the western Wissahickon. The great width of the belt—5 to 9 miles in Montgomery County—and the prevalence of steeply dipping beds suggest that even with a good deal of structural repetition the sequence is many thousands of feet thick. The careful mapping of the section along the Potomac River by George Fisher, who used graded bedding and other top-and-bottom criteria to work out the folds, provides the best estimate of stratigraphic thickness yet made. He concludes (1963, p. 21):

> "The top of the formation is not exposed, and therefore its total thickness is unknown; but at least 14,000 feet of rocks belonging to the Wissahickon Formation [all western sequence] are now exposed along the Potomac, and the original thickness may have been much greater."

Lithology and Field Relations

Pelitic schist with interbedded psammitic rocks constitute the bulk of the western sequence. The pelitic rocks dominate most of the section, but locally they are subordinate. According to Fisher (1963, p. 21) roughly 3000 feet of the section is predominately metasandstones, and about 11,000 feet is mainly pelitic. Very minor amounts of calc-silicate rock—chiefly as relict concretions—and a few mafic interbeds occur in various parts of the sequence. Amphibolite, magnesian schists, and soapstone, mostly in small concordant bodies, are remnants of intrusive mafic and ultramafic rocks that were metamorphosed with the sediments. Thus the western and eastern sequences were alike in many respects. However the western sequence appears to have a higher proportion of metasandstone, and it lacks the thin, very siliceous beds that were probably derived from chert.

Prior to metamorphism the thick western sequence must have been a monotonous repetition of lithologically similar rocks, but the rocks now vary greatly, owing to differences in degree of metamorphism. In the western part of the belt the metamorphic grade is fairly low, and the pelitic rocks appear as fine-grained sericitic schists and phyllites. The schists coarsen to the east, and biotite and garnet appear. The rocks exposed along the Potomac River downstream from Great Falls were the most strongly metamorphosed. Here the pelitic rocks form coarse-grained mica schists, with quartzo-feldspathic segregation lenticles and remnants of large aluminum silicate porphyroblasts. In a few places the schists have been feldspathized and injected by pegmatitic granite, and they locally look like migmatitic gneiss.

Despite strong metamorphism the metasedimentary rocks retain many primary structures: normal bedding, graded bedding, parallel bedding laminations, ripple cross-laminations, convolute bedding, sandstone dikes, slump structures, and chaotic structures resulting from sliding of soft sediments.

The most obvious is bedding, which is well preserved wherever psammitic rocks are present, even at the highest metamorphic grade. Alternating metasandstone and pelitic schist, in thin but laterally continuous layers, is the characteristic bedding style (Pl. 11, fig. 1). The psammitic beds average only about 8 inches thick but are remarkably straight and uniform. Only rarely are individual psammite beds thicker than about 4 feet. Thicker metasandstone layers generally consist of several individual beds without intervening pelite. The pelitic beds, on the other hand, range from very thin partings between the metasandstones to unbroken sequences several hundred feet thick.

Graded bedding is common in some parts of the sequence and is remarkably well preserved. The best examples are along the gorge of the Potomac River, from the lower end of Bear Island upstream to Great Falls. Dozens of graded beds can easily be found in this area, and a diligent search would probably reveal hundreds. The metamorphism throughout much of this area was unusually severe; the sediments were progressively metamorphosed through the sillimanite zone, injected by granite and locally feldspathized, and then partly retrograded to chloritic schists. Yet despite this activity the graded beds and other primary structures preserve their original character with remarkable clarity.

The most obvious graded beds are those that were originally coarse-grained at the base and graded upward to silty tops. In these many of the grains retain their original clastic outlines, and the upward decrease in grain size is still readily apparent (Pl. 12, figs. 1, 2). Beds this coarse are comparatively rare, but good examples are found at Rocky Island below Great Falls, and at the upper end of Bear Island. One graded bed about 4 feet thick on upper Bear Island contains clastic grains up to 6 mm in diameter at its base.

Most of the Wissahickon graded beds, however, were originally much finer-grained. Even their basal part was no coarser than fine-grained sand, and perhaps many were silty. Recrystallization has blurred or obliterated most of the detrital grain outlines. The grading is manifested instead by an upward change in composition; the original sediment became progressively more argillaceous toward the top. In the metamorphosed graded beds this is reflected by an upward increase in mica and decrease in quartz (Table 26, cols 1–2, 4–5). When viewed in outcrop these beds become darker toward the top (Pl. 11, fig. 2; Pl. 13, figs. 1, 2).

The base of the graded beds is always in knife-sharp contact with the underlying pelitic schist or metasiltstone (Pls. 11, 12, 13), but the upper contacts are generally indistinct (Pl. 13, fig. 1) or gradational (Pl. 11, fig. 2; Pl. 13, fig. 2). This provides another criterion for recognizing beds that were graded, where the gradation in grain size or composition is no longer evident. In a few places, however, the graded beds have a sharp upper as well as lower contact; this is most common for relatively coarse-grained beds (Pl. 12, fig. 1).

A characteristic of graded turbidite beds in unmetamorphosed terranes is a distinctive succession of internal structures. Ideally they are as follows, going from the base upward (Kuenen, 1953; Ksiazkiewicz, 1954; Bouma, 1962): (a) massive psammite, (b) thin parallel laminations, (c) ripple cross-laminations, (d) very fine or indistinct parallel laminations, (e) pelite. Some of the zones may be missing, but the order remains the

same. Many of the metamorphosed graded beds along the Potomac show the same sequence of structures. Most commonly they grade from a massive base to a finely laminated upper portion (Pl. 12, fig. 2; Pl. 13, fig. 1) and thence into pelite. In a few beds very fine, wispy current cross-laminations lie above the parallel laminations. In addition to giving valuable information on the origin of the beds these structures provide a further criterion for distinguishing between top and bottom, where other evidence of grading is poorly shown.

Another distinctive type of psammitic bed are those having sharp upper and lower contacts and thin parallel laminations throughout (Pl. 14, fig. 2; Pl. 18, fig. 1). A single bed a few inches thick may contain dozens or even hundreds of these very fine laminations. The laminated beds themselves alternate rhythmically with pelitic beds. Some of the uniformly laminated psammites have sharp lower but indistinct upper contacts (Pl. 14, fig. 2) like the graded beds. Others, in which the laminated structure still predominates, show an obvious size grading at the very base (Pl. 14, fig. 1). Thus there are all gradations from the purely laminated beds to those that are graded and mainly massive.

Laminated beds just like these are well known in unmetamorphosed terranes, where they also occur in thin-bedded sandstone–shale sequences and are associated with graded beds. An excellent summary of these rocks—called laminites—and their relation to graded turbidites, has recently been given by Lombard (1963).

It might be asked whether the lamination in the Wissahickon psammites could be the result of metamorphic differentiation, instead of an inherited sedimentary structure. This seems unlikely, because it compares so closely with the laminations in similar rocks that are unmetamorphosed. Moreover, the parallel laminations in each bed consistantly occurs just above massive zone but below the cross-laminations, something not readily explained by purely metamorphic processes.

Features suggestive of convolute bedding are common in some of the laminites and in the laminated upper parts of some graded beds. The fine laminae are distorted and folded in a wild, irregular way, suggesting lack of tectonic control. The convolutions are restricted to the interior portions of individual beds, and they die out toward the upper and lower contacts. The edges of the beds themselves are straight and undeformed (Pl. 14, fig. 2).

This structure closely resembles the type of convolute bedding described by Williams (1960) and by Rich (1950, p. 729–731). It is believed to have formed by liquefaction and flowage of water-logged fine-grained sands and silts, in beds already covered by other sediments. This structure is obviously not a tectonic or metamorphic phenomenon: deformation of solidified rocks would have also deformed the edges of the beds, not only their internal laminations. Moreover, deformation after the rocks were solidified would have been most intense through the incompetent micaceous (pelitic) beds, and had least effect on the interior of the brittle quartzitic layers. Furthermore, the axes of the convolutions are highly irregular and unrelated to the larger-scale tectonic fold axes in the area.

Other features also point toward soft-sediment deformation. Many of the thinly layered psammite-pelite sequences are intricately contorted and the beds disrupted. Most commonly the pelitic (mica schist) beds were pulled apart, and their fragments were rotated and engulfed by the psammitic material (Pl. 15, fig. 1). At first sight this appears to be normal metamorphic boudinage. However, it is the reverse of the usual metamorphic case, in which the more brittle quartzitic layers are boudinaged. Here

the quartzose psammitic material was clearly the less competent, for it engulfs the pelitic fragments and swirls around them, showing every indication of having been very soft.

Similar effects in metamorphosed psammite-pelite sequences have been described by Sutton and Watson (1955) and by Naha (1961), who also point out that the competency relationships are explicable only if deformation occurred when the sand or silt was a very soupy quicksand.

Some intricately deformed zones, in which the pelitic beds are crumpled, dismembered, and engulfed by psammitic rock, are bounded both above and below by straight, undeformed strata. One such zone (Pl. 15, fig. 2) is only 1–2 feet thick but can be followed laterally for about 40 feet, with the disturbed strata sandwiched between the same two undeformed beds. Throughout this zone the pelitic (mica schist) layers were pulled apart, and the psammitic material flowed in between. Had tectonic deformation acted on hard rocks the enclosing beds could not have escaped deformation.

Relict clastic dikes are another feature in the area that indicate soft-sediment disturbance. These occur in a variety of forms. The simplest are narrow necks that join two adjacent psammite beds, cutting through the pelitic layer between them (Pl. 16, fig. 1). They appear to have been soft sand fillings, formed where the more cohesive pelitic beds were pulled apart. They are extremely common at Bear Island. Another variety, more like a true dike, rises from one psammitic layer and cuts through the overlying strata for several feet or more. Most of these dikes are 1–6 inches thick and rise from a bed whose contorted laminations and other features suggest considerable internal disturbance. Such dikes cut sharply through the pelitic layers but fade into the transected psammitic beds, as if the material in the dike and beds had mixed together while both were soft. Commonly the strata on each side of the dike are offset, showing that the mobilized sand was injected along faults in the unconsolidated beds. A few clastic dikes are quite large and carry with them fragments of disrupted pelite beds (Pl. 16, fig. 1).

A closely related feature is the irregular or bulbous projections that punch upward from the top of some psammitic beds. These were diapir intrusions of soft sand into the overlying strata (Pl. 16, fig. 2). The fine bedding laminations in the psammite become intricately swirled when traced up into the diapir, giving evidence of its mobility.

In most cases the clastic dike material appears to have been forcibly injected upward. This is the rule for clastic dikes in unmetamorphosed Flysch sequences. According to Dzulynski and Radomski (1956, p. 258).

"... it becomes evident that the dikes of the Carpathian Flysch do not result from the entrance of sand into cracks or fissures open to the bottom. They were formed by substratal intrusion of quicksand." "The clastic material was squeezed upwards at the points of minimum resistance." ... "The chief force involved in the formation of the clastic dikes was probably the downward pressure of the load of overlying strata."

These authors also call attention to the close relationship between these dikes and submarine slumping (1956, p. 258–259):

"Most clastic dikes in the Carpathian Flysch are associated with the submarine slumps. It was pointed out by many authors that the slides involved a local tension in the subjacent strata which favours the quicksand injections (Brown, Ch. B. (1938), Anderson, J. L. (1944), Fairbridge, R. W. (1946) and others). This seems to be clearly demonstrated by the occurrences of the dikes penetrating beds pulled apart or thinned out. The slide movement involves also some fracturing which may be followed by quicksand intrusions."

The relict clastic dikes in the Wissahickon also appear to be restricted to zones in which there was slumping and sliding of soft sediments.

Some evidence for slump movements—the disruption of thin pelitic beds—has already been noted, but this marks only an early stage of the process. Where movement was greater the beds are more thoroughly disrupted, and the fragments are rotated out of alignment and rafted apart. Some were curled over, forming hook-shaped fragments. All gradations can be traced from mildly disturbed beds to chaotic zones in which all semblance of stratification is lost. Here randomly oriented fragments of dark mica schist—remnants of dismembered pelitic beds—are engulfed in massive or swirled metasiltstone.

The sandy material was not always the less competent phase, for in some places the psammitic beds were pulled apart and engulfed by pelite. Where the process was arrested at an early stage the psammitic bed is greatly thickened at some places and thinned or necked out at others (Pl. 17, fig. 1). Followed laterally the bed may become increasingly distorted and then dismembered into a string of disconnected, rotated fragments. These fragments were not boudins of brittle metamorphic beds, however, but incompetent gobs of sandy material whose margins have swirled and mixed with the enclosing pelite (Pl. 17, fig. 2). The bedding laminations within these fragments are intensely crumpled and contorted, giving further evidence that the material was very soft. In other places psammitic beds end abruptly in a wavy slurry (Pl. 18, fig. 1), then pass into a chaotic jumble in which randomly oriented gobs and ill-defined smears of psammitic material, amid mica schist, are all that remain of the beds (Pl. 18, fig. 2). It is evident that both the pelitic and psammitic materials were extremely soft when movement occurred.

It may seem strange that in some places the pelitic beds appear to have been the more competent, but that in others they were the less competent; however this is consistent with soft-sediment deformation. Mud would normally flow more readily than the interbedded sand, leading to the kind of relations shown in Plates 17 and 18. Under certain conditions, however, liquefaction of the sand might occur, so that momentarily it behaves as a soupy liquid, having less strength than the somewhat cohesive mud. Once liquified, however, the sand expels water and achieves a more stable packing that resists further movement (Williams, 1960, p. 208–210). Thus the competency relations between pelitic and psammitic sediment could be reversed, even before firm compaction. Such appears to have been the case with the Wissahickon strata.

The most advanced stage of soft-sediment deformation in the Wissahickon is marked by very large chaotic zones, in which fragments and tiny chips of pelitic schist, interspersed with slabs and gobs of psammitic rock up to several feet long, lie in random disarray amid massive or swirled semipelitic matrix. Such zones may range from several feet to tens of feet thick and are underlain and overlain by less deformed or undeformed beds. At Bear Island some of these zones may be followed for several hundred feet along strike. They are sheetlike or lenticular bodies in the otherwise well-stratified metasediments.

These chaotic deposits are believed to have formed by slumping and sliding of poorly consolidated sediments. They are remarkably similar to many of the submarine slide sheets in unmetamorphosed terranes, such as those described by Jones (1937; 1939), Brown (1938), Crowell (1957), and others. Some of the slide structures in the Carpathian Flysch, described and illustrated by Ksiazkiewicz (1950; 1958) are especially similar to the Wissahickon.

The features ascribed here to soft-sediment deformation are best displayed along the Potomac River at Bear Island, where they are easily confused with metamorphic phenomena. Metamorphism has reached the sillimanite zone, and the schists are locally permeated with albite-quartz exudation lenticles and veinlets. Dikelike masses of pegmatitic granite—some intrusive but others clearly formed by replacement—also cut the schists. Thus the evidence of disrupted bedding and flowage in the metasediments, found with the metamorphic exudation and replacement features, leads easily to the conclusion that they all occurred together and that they reflect migmatization and anatexis. However, the same deformation features—the distorted, disrupted, and chaotic bedding—are widespread in fine-grained rocks of the Wissahickon and also in the Peters Creek Formation, where the metamorphic grade is much lower and migmatization is clearly not involved. Conversely, if one were to view the small scale and very localized exudation and replacement features at Bear Island without the background of slump-deformed and chaotic rocks, it is doubtful that many geologists would call it migmatite.

Superimposed on the stratified rocks and on the soft-sediment structures in the Wissahickon are a variety of structures resulting from later deformation and metamorphism —folds, lineations, and one or more generations of schistosity or cleavage. These are described in detail by Cloos (this report), Fisher (1963), and by Reed and Jolly (1963).

Petrography

The western Wissahickon rocks have diversified textures and mineralogy, owing not only to the variety of original rocks but to their present differences in metamorphic grade.

In the central and western parts of the belt, where metamorphic grade is low, the phyllites and fine-grained pelitic schists are composed chiefly of quartz, muscovite, and chlorite. Paragonite occurs sparingly in some of the more aluminous schists, and albite is found in those that are richer in quartz. Magnetite is invariably present and commonly forms conspicuous porphyroblasts. Tiny grains of hematite may also be present, but it is difficult to tell whether it is primary or secondary. Tourmaline is the chief accessory.

The associated psammitic rocks consist chiefly of quartz, plagioclase, muscovite, and chlorite, with various minor accessory minerals (Tables 24 and 25). The newly recrystallized plagioclase in the matrix is albite, averaging about An_3, but large relict grains of more calcic plagioclase (chiefly oligoclase and albite) are generally also present (Pl. 19, fig. 2). Sparse, tiny brown flakes of stilpnomelane are common in the metagraywacke along the western side of the belt and in the Boyd salient. Farther east its place is taken by biotite.

The psammitic rocks preserve their original clastic texture (Pl. 19, figs. 1, 2), leaving no doubt about the nature of the parent rock. They were mostly graywackes, with poorly sorted, subrounded to angular clastic grains in a micaceous-chloritic matrix. Quartz is the most abundant clastic mineral (Pl. 19, fig. 1), but plagioclase (Pl. 19, fig. 2) and rock particles were also common. Texturally, mineralogically, and chemically (Tables 24 and 25) these rocks are typical of unmetamorphosed graywackes, as defined by Pettijohn (1957).

The metagraywackes were originally rather fine-grained; even the largest relict grains rarely exceed about 2 mm. With decreasing grain size the metagraywackes grade into metasiltstones. Many of the metasiltstones are thinly laminated and contain sparse sand-sized clastic quartz and plagioclase grains in a fine, recrystallized matrix (Pl. 21, fig. 1).

TABLE 24

COMPOSITION OF PSAMMITIC ROCKS

Wissahickon Formation, Western Sequence (main belt)

	1	2	3	4	5	6
Quartz	33.5	36.5	31.0	45.5	42.0	53.6
Plagioclase	33.6	29.6	43.9	18.0	21.8	24.7
Muscovite	12.8	11.1	—	15.2	21.8	11.1
Biotite	—	12.9	13.9	9.4	.1	7.2
Chlorite	8.0	.8	tr	4.2	7.7	.1
Stilpnomelane	3.0	—	—	—	—	—
Epidote	3.7	6.6	9.6	3.1	4.6	1.3
Tourmaline	tr	.1	.1	.3	.1	tr
Sphene	.1	—	—	.2	tr	—
Apatite	.2	.2	.1	.3	.3	.3
Heavies*	.1	.1	.1	tr	tr	.1
Magnetite	4.8	2.1	1.2	3.8	1.6	1.6
Hematite	tr	—	.1	—	—	—
Calcite	.2	—	—	—	—	—
Total	100.0	100.0	100.0	100.0	100.0	100.0
Points	3,545	7,858	4,576	1,365	3,218	4,709
Chemical Composition, Calculated from the Modes						
SiO_2	63.2	66.5	67.2	67.1	68.6	75.1
TiO_2	0.3	0.4	0.4	0.4	0.1	0.2
Al_2O_3	15.3	14.4	13.8	13.2	16.5	11.9
Fe_2O_3	6.7	4.2	3.9	5.6	2.8	2.6
FeO	4.8	3.7	3.3	4.6	2.5	2.4
MgO	1.3	1.4	1.4	1.5	1.0	0.9
CaO	1.5	2.2	2.9	1.4	1.7	1.7
Na_2O	3.7	3.5	5.0	2.2	2.6	2.3
K_2O	1.5	2.2	1.2	2.3	2.2	1.7
H_2O	1.6	1.4	0.9	1.6	1.9	1.0
P_2O_5	0.1	0.1	tr	0.1	0.1	0.2
Total	100.0	100.0	100.0	100.0	100.0	100.0

(1) Metagraywacke (H144-1). Ellicott Road half a mile north of Annapolis Rock, Howard County.
(2) Metagraywacke (M456-1). One mile northeast of Clopper, Montgomery County.
(3) Metagraywacke (M455-2). Two-thirds of a mile northeast of Clopper, Montgomery County.
(4) Metagraywacke (M314-1). Piney Branch, half a mile west of Glen, Montgomery County.
(5) Metagraywacke (H164-2). South Branch of Patapsco River near Morgan, Howard County.
(6) Metasubgraywacke (H163-2). South Branch of Patapsco River near Hoods Mills, Howard County.

* Zircon, monazite, xenotime, and other unidentified high-index minerals.

The metamorphic grade increases to the east and southeast and the rocks become slightly coarser-grained, but on the whole they do not change greatly in appearance. Along the eastern side of the belt, adjacent to the Sykesville Formation, the typical pelitic rocks are fine-grained chlorite-biotite-muscovite-quartz schists, some with albite or garnet. The garnet has a high Mn/Fe ratio and is nearer spessartite than almandite (Fisher, 1963). The psammitic rocks are more strongly recrystallized than they are farther west, but their graywacke texture is usually still evident. They consist chiefly of quartz, plagioclase, muscovite, biotite, and chlorite (Table 24).

TABLE 25

COMPOSITION OF PSAMMITIC ROCKS

Wissahickon Formation, Western Sequence

(far western portion)

	1	2	3	4	5	6
Quartz	41.4	31.4	55.6	55.5	69.9	24.2
Plagioclase	17.7	46.5	16.2	25.6	3.8	31.3
Muscovite	26.5	6.9	15.8	10.3	16.7	19.6
Chlorite	8.7	10.0	5.4	6.0	1.2	21.2
Stilpnomelane	.1	—	—	.5	—	—
Epidote	—	.7	—	—	—	.1
Tourmaline	.1	.1	tr	tr	—	.1
Sphene	—	2.4	—	—	—	—
Apatite	tr	.2	—	—	—	.1
Heavies*	.1	tr	.3	tr	.1	tr
Magnetite	3.2	1.8	1.0	1.1	3.7	1.3
Hematite	.1	—	2.3	—	—	.1
Limonite	tr	—	—	.1	.2	—
Leucoxene	1.6	—	3.4	.1	—	2.0
Clay	.5	—	—	.8	—	—
Zeolite	—	—	—	—	4.4	—
Total	100.0	100.0	100.0	100.0	100.0	100.0
Points	5,060	2,875	2,704	1,509	2,613	2,993

Chemical Composition, Calculated from the Modes

	1	2	3	4	5
SiO_2	65.8	67.8	70.9	78.2	80.1
TiO_2	0.9	1.3	2.9	tr	0.3
Al_2O_3	16.5	15.2	10.8	11.5	8.4
Fe_2O_3	4.2	2.5	5.5	1.8	5.0
FeO	3.6	3.1	4.3	2.0	2.4
MgO	1.2	1.3	0.7	0.9	0.2
CaO	0.9	1.5	0.3	0.4	0.1
Na_2O	2.2	5.2	1.8	3.0	1.0
K_2O	2.6	0.7	1.5	1.1	1.6
H_2O	2.1	1.3	1.3	1.1	0.9
P_2O_5	tr	0.1	—	tr	—
Total	100.0	100.0	100.0	100.0	100.0

(1) Metasiltstone (M458-1). One mile southeast of Clarksburg, Montgomery County.
(2) Metagraywacke (M288-1). One mile north-northeast of Seneca, Montgomery County.
(3) Metagraywacke (M458-2). One mile southeast of Clarksburg, Montgomery County.
(4) Metagraywacke (H132-3). Two miles west of Florence, Howard County.
(5) Micaceous quartzite (M452-5). U. S. Route 240, 1 mile west of Neelsville, Montgomery County.
(6) Metasiltstone (M452-4). U. S. Route 240, 1 mile west of Neelsville, Montgomery County.
* Zircon, monazite, xenotime, and other unidentified high-index minerals.

A node of high-grade metamorphism is centered around Bear Island. Approaching this area from the northwest, the pelitic schists become increasingly coarse-grained, and pseudomorphs after staurolite and possibly andalusite appear (Fisher, 1963, p. 172–173). At Bear Island proper the schists contain conspicuous quartz-albite segrega-

tion lenses, fine fibrolitic sillimanite, and large sericitic pseudomorphs after other aluminum silicate minerals. Although the psammitic rocks at Bear Island are more strongly recrystallized than they are elsewhere many of them still preserve quartz and feldspar grains with rounded clastic outlines. Some of the best-preserved relict clastic textures are in a large metasandstone dike near the upper end of Bear Island. The chief minerals in the psammitic rocks are quartz, plagioclase (albite to Na-oligoclase), muscovite, biotite, and retrogressive chlorite (Table 26).

The pelitic schists at Bear Island have had a complex paragenesis, and several im-

TABLE 26

COMPOSITION OF PSAMMITIC ROCKS

Wissahickon Formation, Western Sequence, Potomac Gorge

	1	2	3	4	5	6	7
Quartz	60.1	18.2	39.2	53.7	38.7	46.1	58.9
Plagioclase	20.9	32.1	26.5	22.6	31.7	27.2	15.7
Muscovite	8.3	26.9	17.6	10.3	14.6	12.4	18.8
Biotite	.3	13.5	6.9	6.1	4.1	5.1	4.7
Chlorite	3.0	.4	1.7	.9	3.2	2.1	0.9
Epidote	5.8	4.9	5.4	3.1	3.9	3.5	—
Tourmaline	tr	—	tr	—	—	—	tr
Sphene	.4	.7	.5	.8	1.1	1.0	—
Apatite	.1	.4	.2	.3	.4	.3	tr
Heavies*	tr	tr	tr	.1	tr	.1	tr
Magnetite	1.1	2.9	2.0	2.1	2.3	2.2	.1
Calcite	—	—	—	—	—	—	.9
Total	100.0	100.0	100.0	100.0	100.0	100.0	100.0
Points	2,997	3,192	6,189	3,119	3,073	6,192	1,285
Chemical Composition, Calculated from the Modes							
SiO_2	78.0	55.0	66.5	73.7	66.3	70.0	79.4
TiO_2	0.2	0.7	0.4	0.6	0.7	0.6	0.2
Al_2O_3	11.1	21.4	16.3	11.7	16.1	13.9	11.8
Fe_2O_3	2.3	4.8	3.6	3.4	3.8	3.6	0.4
FeO	1.3	4.1	2.7	2.5	2.7	2.6	1.1
MgO	0.4	1.4	0.9	0.8	0.8	0.8	0.7
CaO	3.2	3.9	3.5	2.8	3.8	3.3	0.8
Na_2O	1.7	2.7	2.2	1.9	2.6	2.3	2.0
K_2O	0.9	3.7	2.3	1.6	1.8	1.7	2.3
H_2O	0.9	2.1	1.5	0.9	1.3	1.1	1.3
P_2O_5	tr	0.2	0.1	0.1	0.1	0.1	—
Total	100.0	100.0	100.0	100.0	100.0	100.0	100.0

(1) Base of 10 inch graded metagraywacke bed (M479-1).
(2) Top of 10 inch graded metagraywacke bed (M479-2).
(3) Metagraywacke (average of 1 and 2).
} Potomac River at head of Rocky Island, Montgomery County.

(4) Base of 30 inch graded metagraywacke bed (M478-1).
(5) Top of 30 inch graded metagraywacke bed (M478-2).
(6) Metagraywacke (average of 4 and 5).
} Potomac River at head of Rocky Island, Montgomery County.

(7) Metasubgraywacke (M429-3). Potomac River at Bear Island, Montgomery County.

* Zircon, monazite, xenotime, and other unidentified high-index minerals.

portant mineralogical relationships have been obscured by strong retrograde metamorphism that has altered the high-grade minerals to fine aggregates of sericite, chlorite, iron oxides, and other products. Quartz, sodic plagioclase, muscovite, and biotite were the original minerals. There were also several aluminum silicate minerals in coarse porphyroblasts, now almost completely altered to sericitic shimmer aggregates. Chief among these was andalusite, which formed large porphyroblasts with square outlines that are easily recognized in outcrop (Pl. 20, fig. 1). Another was a slim, prismatic mineral up to 2 cm long (Pl. 20, fig. 2) that may have been kyanite (Fisher, 1963, Fig. 48; Reed and Jolly, 1963, p. H2, H4) but in my opinion was more likely sillimanite, chiefly because of its square to stout pseudodiamond-shaped cross sections. It is completely altered to shimmer aggregate. A third porphyroblastic mineral appears in outcrop as spots of soft, waxy green material up to 8 mm in diameter. In thin section these appear as shimmer aggregates containing chlorite and sericite. Some of them display relict sieve texture, enclosing grains of quartz. Originally this material was probably cordierite. Reed and Jolly (1963, p. H4, Fig. 6) also report cordierite from Bear Island. Rock from the east end of Bear Island contains still another kind of shimmer aggregate: one composed chiefly of sericite but with magnetite and minor very fine-grained chloritoid; the chloritoid here is clearly one of the alteration products. Perhaps this type of shimmer aggregate was derived from staurolite. Thus, prior to retrogression, the pelite may have been a (staurolite?)-cordierite-andalusite-sillimanite schist.

TABLE 27

CHEMICAL ANALYSES OF METASEDIMENTARY ROCKS, MARYLAND PIEDMONT

	1	2	3	4	5
SiO_2	72.25	70.39	66.43	71.45	53.13
TiO_2	0.83	0.46	0.80	—	1.13
Al_2O_3	11.98	14.00	16.78	14.36	23.58
Fe_2O_3	3.12	2.16	2.56	2.07	5.21
FeO	2.23	2.40	3.67	2.78	4.68
MnO	0.05	0.12	0.13	—	0.17
MgO	1.41	1.55	1.84	1.17	2.07
CaO	0.51	0.63	1.17	1.58	0.26
Na_2O	2.94	2.70	1.55	1.95	1.88
K_2O	2.58	3.70	2.61	3.28	3.41
H_2O^+	2.41	1.65	1.95	—	4.29
H_2O^-	0.15	0.04	0.05	—	0.10
CO_2	0.00	0.05	0.40	—	0.00
P_2O_5	0.17	0.13	0.06	—	0.28
N_2O	—	—	—	1.30	—
Total	100.63	99.98	100.00	99.94	100.19

(1) Metagraywacke (M280-1), Wissahickon Formation, western sequence, Maryland Route 117, 1.2 miles southeast of Boyds, Montgomery Co. H. B. Wiik, analyst.

(2) Meta-arenite (M-251), Sykesville Formation. Maryland Route 97, 0.3 mile south of Glenwood, Howard County. O. von Knorring, analyst.

(3) Meta-arenite (M-133), Sykesville Formation. Seven Locks Road at Bells Mill Road (Batista quarry), Montgomery County. Penniman and Browne, analysts.

(4) Meta-arenite, Sykesville Formation (called biotite granite by Williams). Sykesville, Carroll County. W. F. Hillebrand, analyst. Analysis from G. H. Williams (1895, p. 672).

(5) Phyllite (M281-1), Ijamsville Phyllite. Burnt Hill Road, 1 mile northeast of Clarksburg, Montgomery County. H. B. Wiik, analyst.

Fresh sillimanite in the form of very fine fibrolite is also present but is distinctly later than the porphyroblastic aluminum silicates. It forms felted masses of minute fibers that ramify through the rock, generally parallel to the schistosity but sometimes cutting across it. Fine sprays of sillimanite fibers are concentrated in muscovite, radiating in all directions. Minute fibrolite needles spear through quartz and feldspar. Fisher (1963, p. 185–186) finds that

"... relics of andalusite are partly altered to sillimanite (Figure 51) and partly to muscovite which is itself altered to sillimanite; the muscovite occurs as coarse plates enclosing scattered, optically continuous relics of andalusite, and is partly replaced by a mat of sillimanite needles."

The relationship of the fibrolitic sillimanite to the main period of regional metamorphism is important because it bears strongly on interpretation of the metamorphic facies. It has been proposed that sillimanite, chiefly as fibrolite, marks the peak of regional metamorphism in the area (Fisher, 1963, p. 186; Reed and Jolly, 1963, p. H6). Andalusite and kyanite, according to Fisher (1963, p. 186, 208–210), grew during an earlier stage in the progressive metamorphism but became unstable and gave way to sillimanite (fibrolite) at the highest metamorphic temperatures. Reed and Jolly (1963, p. H6) suggest that andalusite and cordierite may have formed during an early stage of regional metamorphism, but more likely stem from still earlier contact-metamorphism around the small gabbraic intrusions at Bear Island.

The following relationships suggest that the fibrolite is later than the main metamorphism, however, and that it grew during the retrogressive phase:

(1) The fibrolite occurs chiefly in schists in which the coarse-grained minerals were being broken down, recrystallized, and altered. Fibrolite replaces some of these earlier coarse minerals, especially muscovite but also andalusite, quartz, and feldspar. Also, patchily distributed mats and trains of fibrolite anastomose irregularly through some of the rocks, locally wandering across the schistosity formed by the coarse minerals, or following late shear zones.

(2) Fibrolite appears to be later than some of the sericitic shimmer aggregates which are pseudomorphs after the large aluminum silicates porphyroblasts. Sprays of fibrolite seem to penetrate and grow within these sericitic clusters, although the minerals are so fine-grained that it is difficult to be sure. However, *if* the sericitization (formation of shimmer aggregates) came later than all of the sillimanite then why did it so thoroughly replace all the coarse porphyroblastic aluminum silicates but spare the tiny, delicate fibrolite needles? The simpler explanation is that the fibrolite grew later.

(3) Fibrolite is known from only two localities in the Maryland Piedmont—Bear Island, and near Carrolls Mills in Howard County. Its occurrence in both places is restricted to small areas, near where the schist is injected by pegmatite and locally feldspathized. The absence of fibrolite throughout such broad expanses of metamorphic terrane, except in a few small areas where pegmatitic fluids have permeated the schist, strongly suggests that this mineral is related to the pegmatization, not to the regional metamorphism.

These relations recall Watson's (1948, p. 149) conclusions on the formation of fibrolite in the Moine schists at the migmatite complex in Central Sutherland:

"The field and microscopic evidence show that this mineral was formed as a result of metasomatic activity at a late stage in the history of the injection complex when the general metamorphism was already on the wane. The sillimanite seems to have no direct connection with the conditions of regional metamorphism. It was formed under the influence of pegmatitic solutions."

This, I think, was also the case in the Wissahickon.

The regional metamorphism in this area climaxed with the formation of coarsely crystalline schists containing porphyroblastic andalusite, cordierite, sillimanite, and perhaps staurolite. The injection of pegmatitic granite, accompanied by feldspathization, came after this climax, perhaps interrupting a period of shearing and retrogressive hydrothermal alteration. The rise in temperature around the granitized zone, and perhaps the solutions that spread locally through the schists, were probably responsible for the fibrolite.

It seems unlikely that the andalusite and cordierite in the Bear Island schists had anything to do with contact metamorphism around the small gabbro bodies. The distribution of these minerals does not coincide with the gabbro. Moreover, had the gabbro formed contact aureoles with these minerals in them, then such aureoles should be even more pronounced around the Baltimore Gabbro Complex and other large gabbro masses that cut pelitic rocks elsewhere in the Maryland Piedmont. However, andalusite and cordierite have never been found in their vicinity. On the other hand, andalusite-cordierite schists are part of the regional metamorphic sequence in the Virginia Piedmont (Redden, 1963, p. 84–85) and are well known in other regional metamorphic terranes (Read, 1952; Miyoshiro, 1961).

Origin

Prior to metamorphism the western Wissahickon sequence was a thick pile of shales with interstratified beds of fine-grained sandstone and siltstone. The sandy and silty layers were thin but continuous and alternated rhythmically with the shale. Some parts of the section were more sandy, and others more shaly, but rarely were there thick accumulations of one without thin interbeds of the other. These sediments formed a monotonous sequence of strata more than 10,000 feet thick.

The sandstones were chiefly feldspathic graywacke. This is shown by their mineralogy and well-preserved textures, in areas where the metamorphism was weak. The graywackes were fine-grained and consisted of unsorted, poorly rounded to angular grains of quartz, plagioclase, and subordinate rock particles, embedded in matrix of silt and mud. The metasandstones can still be recognized as graywacke where metamorphism was much stronger, partly from relict textures but also from their chemical composition. The compositions of 16 typical metasandstones, taken from Tables 24, 25, 26, and 27, are plotted in Figure 24. They fall well inside the field of unmetamorphosed graywackes (Fig. 23A), although most of them are more siliceous than the average. The plot also shows that these rocks are high in normative corundum—another feature of graywackes, which reflects their abundant argillaceous matrix.

The rhythmic alternation of shale with fine graywacke sands and silts, together with the distinctive assemblage of primary structures which includes graded bedding, thin laminated bedding, convolute bedding, slump structures, and clastic dikes, clearly marks these strata as belonging to the Flysch facies and provides strong evidence for a deep-water marine origin.

Most of the graded beds are of the type believed to have formed by deposition from turbidity currents. The distinctive characteristic is their upward change from massive to laminated sediment, and in some cases on upward to ripple-laminated and then to recurrent (but finer and less distinct) parallel lamination. This sequence of structures is typical of turbidite deposits (Kuenen, 1953; Ksiazkiewicz, 1954; Dzulynski, Ksiazkiewicz, and Kuenen, 1959; Bouma, 1962) and is believed to record deposition from a turbid cloud under conditions of continuously decreasing current velocity (Menard, 1950; Bouma, 1962).

A distinctive feature of turbidites which seems to be scarce in the Wissahickon is sole markings. The difficulty is that the metamorphic rocks break across the bedding rather than along it, thus providing few opportunities to see these structures. At one place on Bear Island, where the base of a psammitic bed is well exposed, there are features that strikingly resemble groove casts. Elsewhere, where the beds are seen only in cross section, the sharp base of some psammitic beds make well-defined bulges, 2–4 inches long and 1–2 inches deep, into the underlying pelite. These appear to be load casts, although some may be flute casts.

Laminites are even more abundant in the Wissahickon than are the graded turbidite beds. There is a gradation between them, however, which suggests a genetic relation. According to Lombard (1963, p. 21):

> "The transition from turbidites to laminites is a fact that can be easily observed. . . . The association of the basic factors is common in both formations: terrigenous material, periodic transport under the force of gravity along a slope, pulsations of current velocity (Sanders, 1960, p. 416), deposition at a relatively great depth."

The conditions of laminite deposition are still not well known, but the consensus is that laminar currents carrying fine, more dispersed debris play the important role. Lombard (1963, p. 21) envisions the process as follows:

> "The transport is fractioned into fine pulses showing that the effects of gravity are combined with the movements of the water. It is not a turbid mass of sediments that slides quickly along the floor but the "cloud" of muddy suspensoids that settles afterwards, either on top of the turbidite or around and even in front of it. . . . This aureole of laminites may directly surround the turbidites but may also spread far away from the sources, retreating or advancing during the sedimentation of the basin."

Perhaps these were the conditions under which many of the Wissahickon beds were laid down.

There is no evidence bearing on the deposition of the pelitic beds, except that individual beds have wide lateral extent. Some may represent slowly accumulated pelagic muds, like those in the Martinsburg Formation (McBride, 1962, p. 70–71). Others may have accumulated more rapidly, during the type of process described above.

Submarine slumping and sliding were evidently quite active during Wissahickon time, judging from the widespread occurrence of the chaotic metasediments. Doubtless some of these slides developed into turbidity flows. The slides themselves may have begun simply by spontaneous movement of unstable, water-logged sediments on oversteepened slopes, but others clearly record later tectonic disturbance, for they involve turbidite and laminite beds—deposits that must have accumulated where the ocean floor was flat or very gently sloping. Evidently orogenic movements were already underway while sedimentation was still in progress.

Volcanic activity was unimportant during accumulation of the thick Wissahickon sequence. Its only record comes from the few beds of amphibolite and greenschist, probably derived from mafic aquagene tuffs. The graywackes do not contain particles of volcanic rock or other evidence of volcanic detritus. The clastic rock particles appear to have been mainly bits of fine-grained metamorphic rock. Also, the relict plagioclases in the graywackes conspicuously lack the characteristics of volcanic plagioclase: pronounced zoning, complex twinning, and a composition generally more calcic than about An_{25}. They appear instead to have come mostly from a low- or medium-grade metamorphic source, perhaps similar to the present Piedmont. The striking lack of relict clastic potassium feldspar in the graywackes apparently indicates that gneisses and granites were not important in the Wissahickon source area.

SYKESVILLE FORMATION

Name

Granitic-appearing rock in the vicinity of Sykesville, Carroll County (pl. 7), was named the Sykesville granite by C. R. Keyes (1895), who believed that it was part of an intrusive igneous mass. This name was retained by A. I. Jonas (1928) and by A. I. Stose and G. W. Stose (1946), who mapped and described the rock in Carroll County, and by Cloos and Broedel (1940) who mapped it southward through Howard County. While subsequently mapping Montgomery County, however, Ernst Cloos became convinced that this rock was metasedimentary rather than igneous. Accordingly, he designated it as the Sykesville formation on the Geologic Map of Montgomery County (Cloos and Cooke, 1953). In this report the name Sykesville Formation is retained, and evidence is given for its sedimentary origin.

Occurrence

The Sykesville Formation extends in a broad belt from the southeastern corner of Carroll County into east-central Montgomery County, where it narrows and fingers out just east of Rockville (Pl. 7). In southernmost Montgomery County the Sykesville rocks reappear, a mile east of Glen Echo, and continue southward into Virginia. They are well exposed along the gorge of the Potomac River in the area around Chain Bridge. This southern extension of the Sykesville Formation, partly isolated by intrusive rocks, lies directly on strike with the main belt and is linked to it by numerous small lenticular masses of Sykesville-like rock, interspersed through the intervening Wissahickon schist.

Another body of Sykesville rocks, much thinner than the main belt and partly interstratified with mica schist, crops out along the Potomac River near Cabin John. From there it may be traced northward for about 5 miles to Tuckermans Lane. Beyond that point isolated slivers of Sykesville-like rock in the Wissahickon schist, mostly too small to show on the map, continue northward past Rockville and join with the main Sykesville belt.

Thus the Sykesville Formation in southeastern Montgomery County is a semicontinuous mass all the way to the Potomac River, rather than 3 unrelated bodies. South of Rockville, however, the Sykesville intertongues with Wissahickon schist, which for several miles becomes the dominant rock. Also, the Sykesville in this area has been split apart and widely separated by mafic and quartz dioritic intrusive rocks. The general relations are shown in Figure 25.

Still another body of Sykesville crops out along the Potomac River farther upstream, at Offut and Hermit Islands. Fisher (1963, Pl. 1) maps this as an anticline, and suggests that the Sykesville here may correlate with the Cabin John belt 3 miles to the east.

Thickness

The Sykesville Formation is thought to be a thick pile of submarine slide material. It is not a sheet-like deposit of uniform thickness but an elongate lenticular mass that intertongues with the Wissahickon Formation along its top, bottom, and front. It may be roughly likened to a broad apron of coalescing alluvial fans along the front of a mountain range (Fig. 26). The main Sykesville belt provides a longitudinal cross section through the mass, which is tipped on end.

In Howard County and in the northern part of Montgomery County the Sykesville belt is approximately 3 miles wide and is bounded on both sides by stratified schists dipping 60°–90° west. Its apparent stratigraphic thickness here is about 15,000 feet,

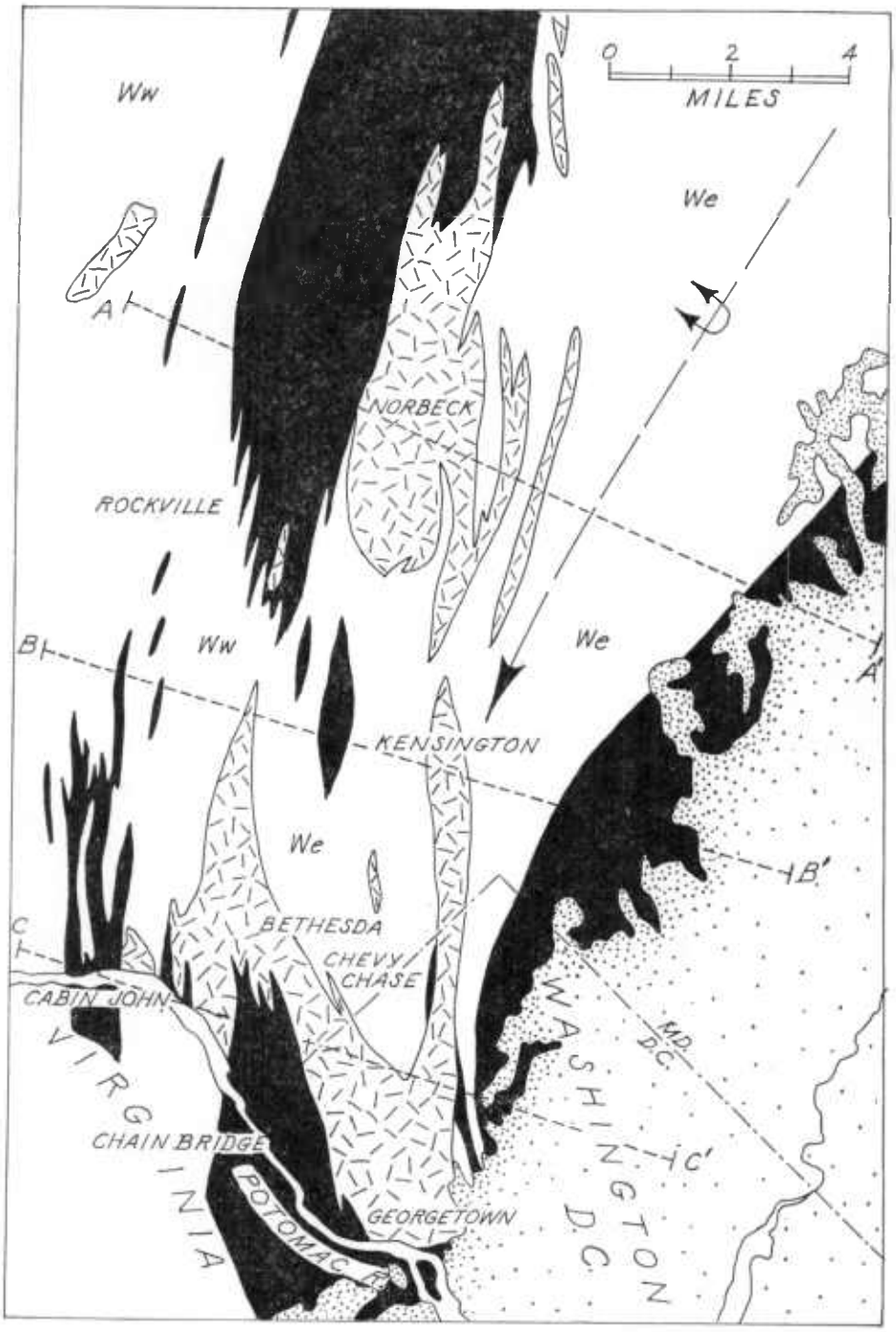

0 2 4
MILES
Ww
We
A
NORBECK
ROCKVILLE
B
Ww
We
A'
KENSINGTON
We
B'
BETHESDA
C
CHEVY CHASE
CABIN JOHN
VIRGINIA
MD.
D.C.
WASHINGTON D.C.
C'
CHAIN BRIDGE
POTOMAC R.
GEORGETOWN

FIG. 25

but the amount of tectonic distortion is unknown. If the mass was distorted by regional folding it was probably stretched and thinned, since it lies along a steep limb of the Baltimore anticlinorium, and was probably dragged upward by the gneiss domes rising at the core. The steeply plunging *a* lineation in the rock (*see* Cloos, this report) appears to reflect this stretching. It therefore seems unlikely that the original body was much thinner than the apparent 15,000 feet, here at its thickest point.

Lithology and Field Relations

The Sykesville Formation consists of a heterogeneous group of pebble- and boulder-bearing arenaceous to pelitic metamorphic rocks. Its appearance varies from place to place. Typically it resembles medium-grained, weakly gneissic granite (Pl. 22, fig. 1), but locally the foliation is stronger and there are all gradations to intensely foliated gneiss or even schist. Another variation is a fine-grained, nearly massive rock that resembles dark, impure quartzite. In the field these variants are broadly gradational into one another; only rarely are they found in sharp contact. Indeed, the massive, unstratified appearance of the Sykesville is one of its most distinctive characteristics.

Another important feature is the "inclusions" found throughout the formation. These rock fragments range from boulders and slabs up to 15 feet long to particles so tiny that they are difficult to distinguish from grains in the enclosing matrix. In some outcrops the fragments are so abundant that the rock is best described as breccia, but elsewhere they are scarce or almost lacking. The inclusions are chiefly of three kinds: (1) rounded quartz lumps, (2) tiny chips and blebs of mica schist, and (3) large blocks and slabs of various metamorphic rocks. Types (2) and (3) look much like autoliths and xenoliths in granitic rocks.

(1) The quartz lumps are spherical to ellipsoidal pieces of recrystallized massive vein quartz, and they look very much like pebbles (Pl. 23, fig. 1). Quartz lumps up to an inch in diameter are common, and some reach 2 inches. The largest found—a perfectly round cobble—was 6 inches across. Pea-sized lumps are common, and they range down to grains a few millimeters in diameter that cannot be distinguished from quartz grains in the host rock. These objects are not concentrated in beds or pockets like normal conglomerate but are interspersed randomly and often are widely scattered. A quartz lump may be several feet or even tens of feet from its nearest neighbor. Locally there are also gash fillings and irregular segregations of quartz. These are distinct from the rounded lumps.

(2) Tiny chips of dark mica schist are most prolific (Pl. 23, fig. 2). Most of them range from 5 mm to 3 cm long, but there is no well-defined size limit. On one hand some are so tiny that it is hard to tell them from clots of segregated mica in the matrix, while on the other hand there is no sharp separation from slabs of schist up to several feet long.

FIGURE 25. Generalized geologic map of part of southern Montgomery Country and Washington, D. C., showing the relationship of the Sykesville and Laurel Formations (in black) to the other rocks. The Laurel Formation borders the Coastal Plain north of Georgetown; all other rocks shown in black are Sykesville Formation. Wissahickon Formation shown in white (We, eastern sequence; Ww, western sequence); intrusive rocks, hatched; Coastal Plain sediments, stippled. Shown also is the axial trace of the Baltimore anticlinorium, overturned slightly to the east and plunging to the south. The Sykesville and Laurel Formations outline the limbs of the anticlinorium, but are split apart by mafic and quartz dioritic intrusive rocks at its nose. A-A′, B-B′, and C-C′ show traces of cross sections shown in Figure 26. Geology taken from Cloos and Cooke (1953), Fisher (1963), and unpublished mapping in Virginia by Martin Tischler.

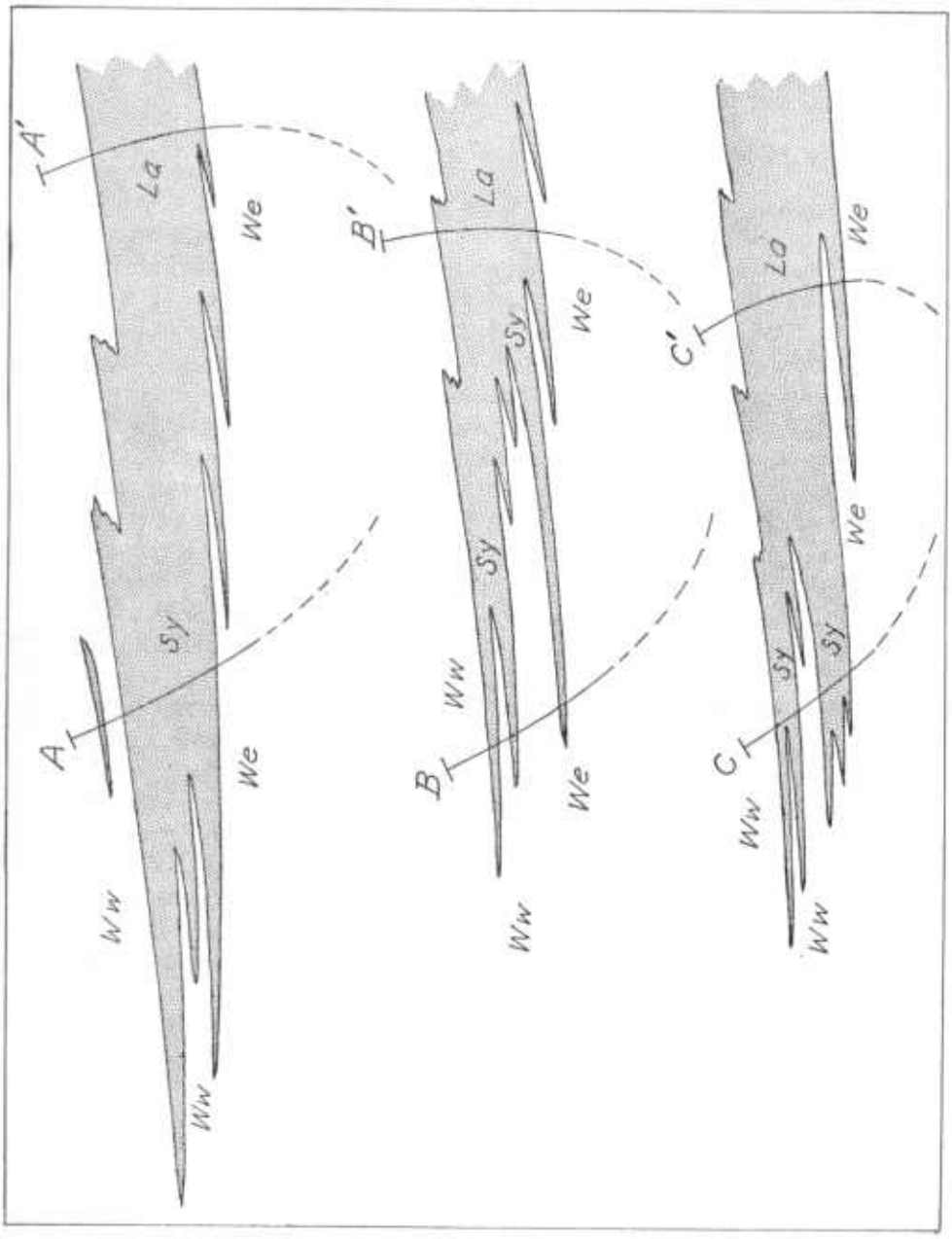


FIGURE 26. Hypothetical cross sections showing the Sykesville (Sy), Laurel (La), and Wissahickon (We, eastern sequence; Ww, western sequence) Formations as they might have appeared before folding and igneous intrusion. The Sykesville and Laurel are interpreted as coalescing fans of submarine slide debris, which thin and lense out to the west. The cross sections are taken at A-A', B-B', and C-C' in Figure 25. If the Sykesville-Laurel wedge were folded into an anticline, so that the curving lines A-A', B-B', and C-C' were bent to straight horizontal lines, the formations would show approximately their present dip. Diagram not drawn closely to scale.

(3) Most striking are the large blocks and slabs of metamorphic rock (Pl. 24, figs. 1, 2). Blocks 2 feet across are quite common. An 8-foot block of amphibolite occurs in the rock near Sykesville, and Fisher (1963, p. 38) reports schist fragments up to 15 feet long. The blocks and slabs are randomly oriented at most places, standing at all angles to one another (Pl. 24, fig. 2) and to the foliation of the host. Even where the blocks are roughly aligned with the foliation, their internal structures are discordant (Pl. 24, fig. 1). The lamination inside the fragments may be intensely contorted and crumpled, as if deformed while soft—yet the fragments have straight edges (Pl. 22, fig. 2). In a few cases the blocks themselves appear to have been soft: some have long wavy tails like a tadpole, and others have irregular "slurried" outlines and grade into the matrix.

The large inclusions consist of the following rock types:

(1) Metagraywacke, and dark micaeous quartzite (Pl. 24)

(2) Mica schist (Pl..23, fig. 2); commonly garnetiferous

(3) Amphibolite

(4) Calc-silicate rock; usually thin bodies in a sheath of contorted psammitic rock (Pl. 22, fig. 2)

(5) Ultramafic rock

(6) Gneiss, granulite and granite; apparently basement rock.

Types (1) and (2) overwhelmingly predominate, and (5) and (6) are rare.

West of Brookville, near Chain Bridge, and elsewhere the bouldery inclusions seem to be concentrated along ill-defined zones tens of feet thick. Fragments of all kinds may be more abundant along some zones than others, or a certain type of block, such as amphibolite, may predominate along one zone. This very coarse, poorly developed "layering" is the only hint of bedding throughout most of the mass.

Contacts between the Sykesville and the Wissahickon are gradational. Toward a contact the Sykesville rocks gradually become less granitic in appearance, and the inclusions decrease and then disappear. Layers of mica schist appear in the massive Sykesville rock and gradually predominate. At the same time the massive rock becomes finer-grained. This soon passes into normal Wissahickon lithology: mica schist with thin beds of psammitic granulite. The gradational contact zone may be tens or hundreds of feet wide.

Petrography

Mineralogically the Sykesville rocks are remarkably uniform, despite their wide variations in texture. Quartz, plagioclase, and muscovite generally make up about 85 to 90 per cent of the rock, with biotite, chlorite, epidote, magnetite, and commonly garnet as the prominent dark minerals. Minor accessories include apatite, zircon, tourmaline, sphene, allanite, calcite, and pyrite. The major minerals occur in varying proportions; some rocks are more quartzose, others more feldspathic or more micaceous, but the range is not great (Fig. 27). Modal analyses of typical Sykesville rocks are listed in Table 28.

The rocks have a relict clastic texture, modified to varying degrees by deformation and recrystallization. Some of the larger grains of quartz and feldspar have obvious rounded clastic outlines. These are enclosed in a seriate granoblastic to schistose matrix of fine-grained quartz and feldspar, and fine- to coarse-grained mica and chloride (Pl. 25). Some of the rocks show strong segregation of micas and finely granular quartz into streaks and lenses, emphasizing the schistosity; in others these minerals are randomly interspersed and have only weak orientation. These differences in the character of the matrix are due mainly to differences in the intensity of deformation and recrystallization.

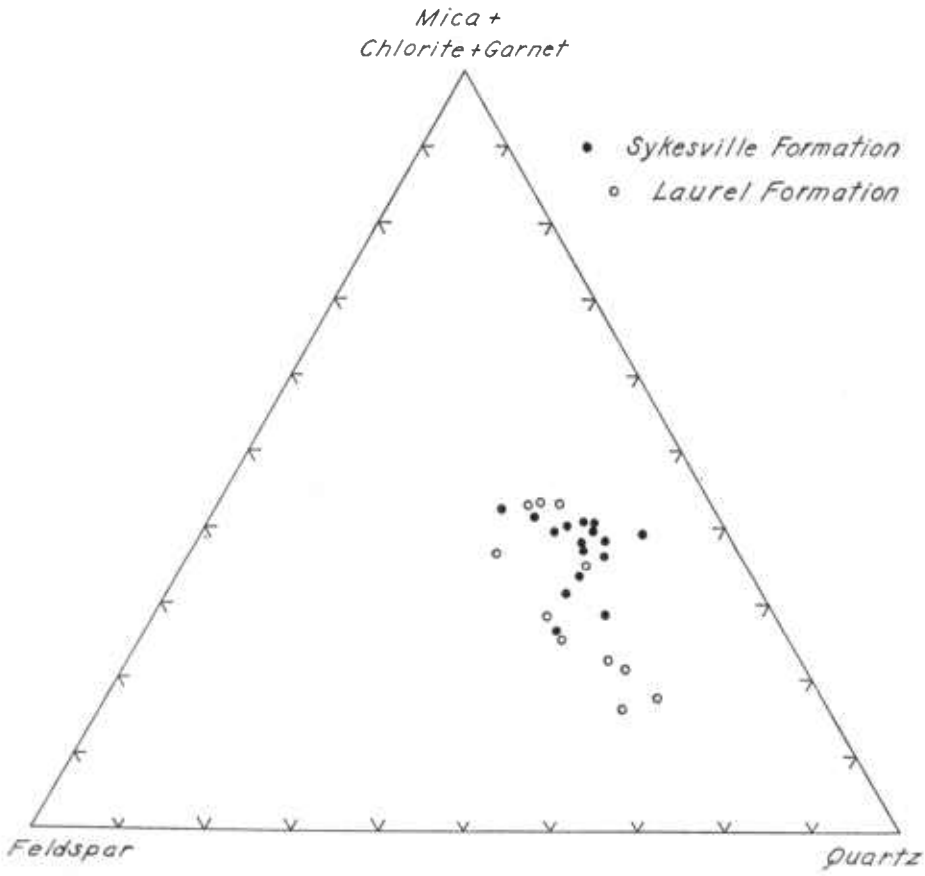


FIGURE 27. Mineralogical compositions of Sykesville and Laurel Formations. Plotted from modal analyses in Tables 28 and 29. The two Laurel rocks richest in quartz are from local remnants of psammitic beds and do not represent the Laurel's average composition.

Quartz is the most abundant mineral; it constitutes 35–50 per cent of most rocks. It occurs as relict clastic grains and as tiny irregular grains in the recrystallized matrix. Clastic grains smaller than about $\frac{1}{2}$ mm are rarely preserved. In the least-deformed rocks the relict clasts are spherical to egg-shaped (Pl. 25), but with increasing deformation they become almond-shaped, with tails of recrystallized mortar at the ends. In the most intensely deformed rocks even the largest clastic grains are pulled out into thin pods and lenticles, and their original outlines are destroyed. But even the least-deformed clasts are recrystallized, so that under crossed nicols they appear as sutured mosaics of smaller grains (Pl. 25, fig. 1).

There is no difference, other than size, between the clastic quartz grains seen in thin section and the quartz "lumps" so conspicuous in outcrop (Pl. 23, fig. 1). Moreover, there are all gradations in size between them. It is evident that the quartz lumps are relict pebbles and granules, in a partly sandy matrix.

The feldspar is plagioclase, which occurs as rounded clastic grains that have survived metamorphism and as newly formed porphyroblasts and fine granoblastic grains in the matrix. The porphyroblastic plagioclase is readily distinguished from the relict clastic plagioclase as the latter has rounded outlines, sharp margins, few inclusions, relict albite or pericline twinning, and is commonly clouded by sericite and tiny epidotes or

TABLE 28

MODAL ANALYSES, SYKESVILLE FORMATION

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16
Quartz	44.1	46.5	42.9	32.4	43.0	50.2	45.6	48.1	42.4	46.2	45.7	38.9	36.3	42.9	40.0	43.4
Plagioclase	17.3	22.7	15.5	24.1	14.6	18.6	14.1	9.3	13.9	20.4	15.1	19.1	20.5	17.1	17.2	15.6
K feldspar	—	3.0	—	—	—	—	—	—	—	—	—	—	—	—	—	—
Muscovite	26.4	19.7	29.9	33.4	26.8	20.8	25.2	25.4	28.1	19.6	22.0	26.8	28.4	24.0	26.9	27.8
Biotite	8.2	6.2	9.6	8.3	7.2	4.0	10.4	11.5	4.7	8.1	12.6	11.5	11.9	12.6	7.7	7.3
Chlorite	1.8	—	0.7	t	4.1	3.0	1.9		5.9	1.8	—	—	0.2	0.2	3.9	2.7
Garnet	0.2	—	0.2	0.2	0.5	—	0.3	1.2	—	—	—	—	—	—	0.4	—
Epidote	0.2	1.0	—	0.5	2.1	1.9	2.0	2.4	3.1	2.9	2.2	1.8	1.8	1.7	0.4	0.2
Allanite	—	t	—	0.1	—	0.1	—	—	—	—	—	—	0.2	—	—	—
Tourmaline	—	—	t	—	—	—	0.3	—	—	t	0.1	—	—	—	—	—
Calcite	—	0.2	—	—	—	—	—	—	—	0.2	0.1	0.2	—	0.4	—	0.3
Sphene	—	0.4	—	—	t	—	—	0.1	—	—	—	—	t	0.3	—	0.1
Apatite	t	0.3	0.1	0.2	t	0.1	t	0.2	0.3	0.2	—	0.2	t	0.3	0.4	0.2
Zircon	t	t	t	t	t	t	t	t	—	t	t	t	t	—	—	0.2
Opaques*	1.8		1.1	0.8	1.7	1.3	0.2	1.8	1.6	0.6	2.2	1.5	0.7	0.5	3.1	2.2
Total	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0
Points	1278	1686	1286	1255	1158	1286	1237	1143	1159	1328	1148	1117	1250	1272	1351	1290

(1) H15-1. Piney Run at Sykesville Road, 1.6 miles north of Sykesville, Carroll County.
(2) H16-1. South Branch Patapsco River, 0.7 miles below Sykesville, in Howard County.
(3) H14-1. U. S. 40, 2 miles east of Hoods Mill Road, Howard County.
(4) M-251. Roxbury Mills Road, 0.3 mile south of Glenwood, Howard County.
(5) M-252. Near Roxbury Mills Road, 0.2 mile east of Roxbury Mills, Howard County.
(6) H99-1. Near Roxbury Mills Road, 0.2 mile east of Roxbury Mills, Howard County.
(7) M-253. Patuxent River at Roxbury Mills Road, Howard County.
(8) M-254. Boat landing, near head of Tridelphia Reservoir, Montgomery County.
(9) M-255. Mount Zion Road, 0.5 mile west of Brookville, Montgomery County.
(10) M-256b. Quarry along Mount Zion Road, about 1 mile west of Brookville, Montgomery County.
(11) M-133. Batista quarry, Cabin John Creek at Bells Mill Road, Montgomery County.
(12) M-260. Batista quarry, Cabin John Creek at Bells Mill Road, Montgomery County.
(13) M409-1. North shore of Potomac River at Chain Bridge, Montgomery County.
(14) M409-2. North shore of Potomac River at Chain Bridge, Montgomery County.
(15) M328-1. South shore of Potomac River at Chain Bridge, Fairfax County, Virginia.
(16) M-42. Potomac River, north of the hydro plant, Dalecarlia, Montgomery County.
* Chiefly magnetite, but includes some hematite and pyrite.

clinozoisite. The porphyroblasts are strongly sieved and have irregular to amoebaform shapes with delicate arms that project into the matrix (Pl. 21, fig. 2); they are also untwinned and unclouded by alteration products. The porphyroblastic and fine-matrix plagioclase are sodic oligoclase to albite (An_{8-15}). The relict plagioclases are generally about the same composition, but some are more calcic.

Not all the clastic grains were single crystals. Some consist of two feldspars or of several interlocking feldspar and quartz crystals (Pl. 25, fig. 1); these were particles of fine-grained metamorphic rock.

Potassium feldspar is not a normal constituent of the Sykesville rocks. All thin sections were stained for this mineral, but it was found in only one (Table 28, H16-1). Here it was microcline, which appeared to replace plagioclase and quartz along zones of crushing and granulation. The rock was collected near a pegmatite dike and had probably been permeated by potassic fluids.

Muscovite is inconspicuous in hand specimen but normally constitutes 20–30 per cent of the rock. It occurs in fine flakes and in concentrations of very fine sericite. In a few rocks it forms large skeletal porphyroblasts, lying athwart the schistosity. The white mica was X-rayed and none was paragonite.

The chief dark minerals are biotite and chlorite, which together total about 9–12 per cent of the rock. Biotite predominates, except where is has been partly replaced by later chlorite. Not all the chlorite is secondary, however; generally it is in well-formed flakes sharply bounded against the biotite, and apparently in equilibrium with it. The biotite and chlorite are commonly segregated together into clusters, making the rock appear much coarser-grained than it really is (for example, see Pl. 22, fig. 2). The biotite is pleochroic from dark brown to straw yellow. Chlorite is pleochroic from grass green to colorless; according to Fisher (1963, p. 128) it is repidolite.

Most of the Sykesville rocks contain very sparse garnets, that form small skeletal crystals and large well-formed porphyroblasts. Many are sieved, but none show the curving trends that indicate rotation during growth. The garnets are pyralspites, according to Fisher (1963, p. 129; Table 15), with variable but often high spessartite content.

Magnetite is the chief opaque mineral. Some tiny opaque patches, now consisting chiefly of leucoxene, may have been ilmenite. In addition, some of the rocks contain conspicuous cubes of pyrite. No primary hematite was observed.

Origin

The origin of the Sykesville rocks was first considered by C. R. Keyes (1895) and G. H. Williams (1895). Keyes, working in the area around Sykesville, concluded that the rock was an intrusive granite, but that it had been "considerably squeezed", so that in many places it might be mistaken for a gneiss. The inference that it was igneous was based chiefly on the presence of abundant blocks of foreign rock, which Keyes (1895, p. 726) believed were xenoliths:

> "During the progress of the work large numbers of inclusions of foreign rock were brought to light in the granite, furnishing good proof of its eruptive origin."

Also (1895, p. 732),

> "Perhaps one of the most conclusive proofs of the eruptive nature of some of the Maryland granites is the occurrence in the mass of large numbers of inclusions—fragments of foreign rocks, both sedimentary and eruptive. . . . At Sykesville, where they occur so abundantly, the irregular angular fragments and blocks of all sizes are identical with rocks of the neighborhood."

For the origin of other granites in the Maryland Piedmont Keyes placed much reliance on intrusive relations visible along their contacts—the crosscutting, diking, and metamorphism of the wall rocks. He was unable to obtain such evidence for the Sykesville mass, however, partly because the deep soil cover obscured its contacts. The inference of its igneous origin, therefore, rested chiefly on the inclusions.

A similar conclusion was reached by G. H. Williams (1895, p. 665, 669) for rocks along the Potomac River near Chain Bridge, now recognized as part of the Sykesville Formation. He observed that:

"... extreme foliation has obliterated most of the granitic features, and it is only because of the abundance of indistinct masses resembling inclusions which are scattered through it that we assume for this rock an igneous origin."

Stose and Stose (1946, p. 91–93) believed that the rock was intrusive near Sykesville and adjacent parts of Carroll County, chiefly because of relationships along its margin:

"The intrusive relation of the Sykesville granite to the Peters Creek formation is well shown at many places along the contact. The bordering country rock for some distance from the massive granite contains stringers and lenses of granite which have been twisted and kneaded with the sedimentary rock. The granite has penetrated the Peters Creek formation in the form of lit-par-lit injection and in stringers and dikes. Near the contact the granite contains inclusions of country rock, both large and small, the largest observed being 6 feet across. The inclusions are garnetiferous biotite schist and quartzite. The contact rock contains abundant large garnets which are chloritized in many places. Knots of quartz up to 1½ inches in diameter are abundant in the granite near the contact. The quartz knots are elongated bodies that resemble pebbles."

This does not accord with the observations of Cloos (Cloos and Cooke, 1953; and personal communication), Fisher (1963, p. 41–42), or myself; all of us have found the contact to be gradational. The Sykesville rock becomes more micaceous and less "granitic" in appearance toward the contact and passes gradually into lustrous mica schists of the Wissahickon Formation. Moreover, the inclusions in the Sykesville diminish, rather than increase toward the contact. Internal structures in the Sykesville are concordant with those in the wall rocks, never crosscutting. At some contacts layers in the schist vaguely resemble granite, but their relict clastic textures show that these are arenaceous interbeds, not *lit-par-lit* injections. The contact relations of the Sykesville, in our opinion, show sedimentary gradation into the Wissahickon metasediments, not intrusion.

Petrographic features of the Sykesville rock also disprove an igneous origin. The rock's texture does not show evidence of magmatic crystallization but preserves relict sand grains and pebbles. Nor does the Sykesville correspond mineralogically to magmatic granites, in which feldspars equal or exceed quartz. In the Sykesville rocks feldspar is only one-half to one-third as abundant as quartz. The average feldspar content of 16 Sykesville rocks is 17 per cent, the average quartz content 42 per cent.

It might be argued that the granite was metamorphosed, with much of the original feldspar being converted to muscovite and zoisite (epidote). The chemical composition, therefore, might be more relevant than the mineralogy. Chemical analyses of three Sykesville rocks are given in Table 27, cols. 2–4. Their composition does not at first appear to differ greatly from that of quartz monzonite or granite. An important difference, however, is that the Sykesville rocks are lower in alkalis, relative to alumina and silica, than igneous granites are. This is reflected by their low normative feldspar and relatively high normative corundum and quartz.

These chemical characteristics are best brought out graphically, by plotting normative orthoclase, albite, quartz, and corundum on a triangular diagram, as was done for the Wissahickon. Plotted in Figure 28 are three Sykesville rocks, taken from Table 27. Also plotted, for comparison, are average granodiorites, quartz monzonites, and granites and average shale. Shown also on the diagram is the contoured "graywacke field" and a projection of the synthetic granite minimum. The granitic rocks plot on the feldspar side of the granite minimum, and the sedimentary rocks plot mostly on the SiO_2 side. The Sykesville rocks plot with the sediments. Moreover, the Sykesville rocks show high normative corundum—a characteristic of sediments rich in argillaceous material—whereas the granitic rocks have little or none.

In conclusion, the field relations, textures, mineralogy, and chemical composition of the Sykesville rocks point unmistakably to a sedimentary origin.

In Figure 28 it can be seen that the Sykesville rocks—ignoring for a moment the inclusions and pebbles—are intermediate in composition between typical graywacke and shale. They are more potassic and have more excess alumina than average graywacke does, which probably reflects a higher proportion of argillaceous matrix. This is borne out by the texture of the rock itself (Pl. 25, figs. 1, 2) which shows that relict sand grains form an unsupported framework amid an excess of micaceous-chloritic (argillaceous) matrix. The rock, therefore, was not a normal sandstone but closer to a sandy mudstone.

The origin of the Sykesville must take into account the following characteristics: (1)

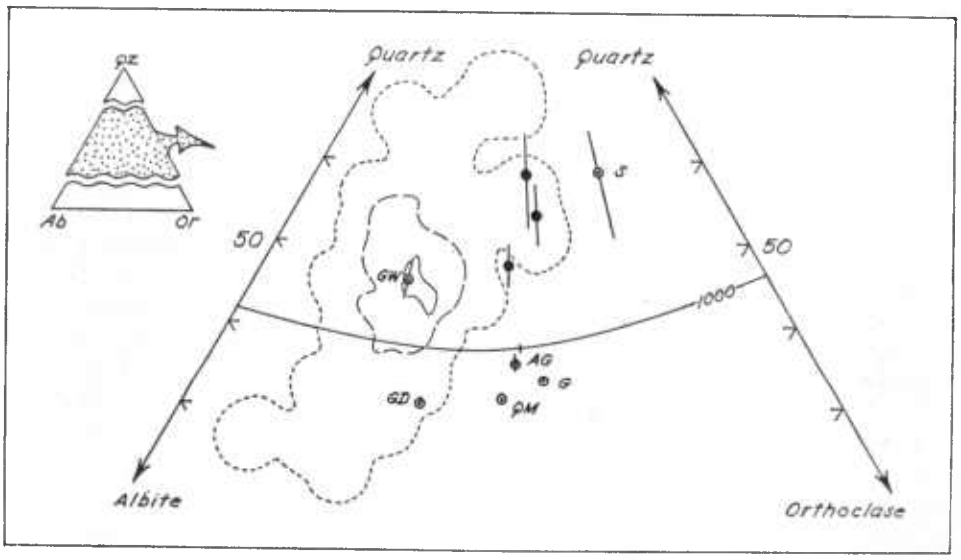


FIGURE 28. Composition of rocks from the Sykesville Formation, and their relation to possible parent rocks. The diagram is a portion of a triangle with normative (CIPW) quartz, albite, and orthoclase at the corners. The steep line through the points is normative corundum. Solid black circles are chemically analyzed Sykesville rocks from Table 27. Circles with dot are the following average sedimentary and plutonic rocks: S, shale (Clarke, 1924, p. 24); GW, graywacke (Pettijohn, 1957, p. 307, col. A); GD, granodiorite; QM, quartz monzonite (adamellite); G, granite; AG, alkali granite (Nockolds, 1954, Tables 1 and 2). The curved dashed lines are contours outlining the graywacke field (see Figure 23). The solid curved line marks the projection of the synthetic granite minimum at 1,000 bars P_{H_2O} (Tuttle and Bowen, 1958, p. 75). The igneous rocks plot below this minimum, and the clastic sediments mostly above. Graywackes and shale have considerable normative corundum (excess alumina), in contrast to the granitic rocks. Composition of the Sykesville rocks is intermediate between average graywacke and shale.

It was an unsorted clastic sediment; the matrix was a mixture of mud, silt, sand, and it carried pebbles, cobbles, and boulders. (2) The mass lacks bedding or stratification. (3) The slabby inclusions stand at all angles to one another (Pl. 24, fig. 2), showing that the sediment did not accumulate gradually but that masses of it were "dumped" as a unit.

Several types of deposits have these characteristics: epiclastic volcanic breccia, glacial till, terrestrial mudflows and landslides, and the products of submarine slumping and sliding. Volcanic breccia can be eliminated, because neither the fragments nor the matrix material is metavolcanic. Pettijohn (1957), Crowell (1957), Dott (1961), and others have discussed criteria for distinguishing glacial tills from till-like deposits originating by mass flowage. Some of these criteria cannot be applied to the Sykesville because of its metamorphism. Nevertheless, the great thickness of the Sykesville, and its close association with the Wissahickon marine turbidites—instead of with sediments normally associated with tills, such as varved argillites with rafted pebbles, and channeled, cross-bedded stream deposits—argue strongly against a glacial origin. Other origins on land, such as mudflows or landslides, are unlikely for the same reasons.

The features of submarine slide deposits are well known from the studies of Bailey *et al.* (1928), Hadding (1931), Jones (1937; 1939), Brown (1938), Rich (1950), Dorreen (1951), Kugler (1953), Renz *et al.* (1955), Crowell (1957), Hawley (1957), Ksiazkiewicz (1958), and others. Most of the examples involving large-scale sliding are restricted to Flysch environments, where even the well-bedded strata reflect gravity transport, by turbidity currents and minor slumping. A typical example may be cited (Dzulynski, *et al.*, 1959, p. 1095):

> "There are also numerous indications of slumping in the Carpathian flysch, and nearly every stratigraphic unit contains beds distorted or disrupted by submarine sliding. In other cases the reworking has gone much farther, and the resulting deposits consist of clay or silt mixed with pebbles and blocks. . . . The thorough mixing of pebbles and boulders with clay and their chaotic arrangement signify that these beds were formed by true subaqueous mud flows. By the regional occurrence of such deposits within the Babica clays these slumps can be traced at least 30 kilometers from the presumed source. This indicates extreme mobility for such mud slides, and depths far exceeding neritic must have been attained at the far end."

The Sykesville and the Wissahickon Formations, taken together, duplicate most of the features shown by unmetamorphosed sediments that were formed by submarine slumping and sliding. The Wissahickon preserves the early stages of the process, in which well-bedded sediments were distorted and disrupted by slumping, and then chaotically mixed by sliding on a fairly small scale. The Sykesville shows the same chaotic mixture of fragments, but developed on an enormous scale, and with all trace of bedding obliterated.

Some of the fragments in the Sykesville have characteristics that are unique to those found in subaqueous slide deposits, described by Crowell (1957), Ksiazkiewicz (1958) and others:

(1) Hook-shaped, curled, and rolled-up fragments, evidently remnants of disrupted slump overfolds

(2) Quartz pebbles embedded in pelitic schist and metasiltstone, suggesting that the pebbles were pressed into the fragments while they were still soft

(3) Long, thin, angular slabs, fragments of competent beds that were pulled apart

(4) Fragments with intensely contorted internal laminations, suggesting deformation of soft beds, before they were pulled apart.

(5) Indistinct, blurred contacts between fragments and host, showing that the fragment was soft and had partly mixed with the matrix.

Many of the features of the Sykesville could be accounted for by sliding and thorough mixing of Wissahickon sediments alone. The Wissahickon consisted chiefly of graywacke and shale, which, if mixed together while still soft, might have yielded a sandy mudstone like the Sykesville sediment. Moreover, the vast majority of fragments found in the Sykesville could have come from disrupted Wissahickon beds. The three most abundant kinds of inclusions—metagraywacke and metasiltstone (Pl. 24), pelitic schist chips (Pl. 23, fig. 2), and calc-silicate slabs (Pl. 22, fig. 2)—are closely matched in the Wissahickon by disrupted psammitic beds (Pl. 17), disrupted pelite beds (Pl. 15), and calcareous concretions. Some of the amphibolite fragments might also have come from ophiolitic beds or sills in the Wissahickon.

Yet the Wissahickon could not have been the sole source of the fragmental material. The Sykesville also carries exotic inclusions—the quartz pebbles and fragments of gneissic basement rock—as well as abundant relict clastic quartz and feldspar grains that are much coarser than those found in the fine-grained Wissahickon sands. Furthermore, a thorough mixing of the Wissahickon—in which shale was 3 or 4 times more abundant than sandstone—would have led to a sediment considerably less sandy than the Sykesville.

The Sykesville submarine sliding, therefore, appears to have involved not only the Wissahickon Flysch sediments but also large amounts of extraneous coarse sand, quartz gravel, and coarse fragmental debris from the basement. Such material was probably derived from a near-shore, shallow-water environment.

The Sykesville slide mass is unique in many ways and raises some puzzling questions. One outstanding feature is its immense size. The mass appears to be up to 15,000 feet thick in Maryland, and a rough reconnaissance finds similar rocks along strike from as far north as Lancaster County, Pennsylvania, to as far south as near Quantico, Virginia (F. J. Pettijohn, personal communication), a distance of at least 100 miles. The only other submarine slide mass of comparable size, known to me, is the argille scagliose of the Apennine Mountains, Italy (Merla, 1951; Maxwell, 1953; 1959).

Another peculiar feature is that the mass appears to be a thoroughly homogenized mixture of deep-water Flysch sediments with coarser clastic material derived from nearer shore. It is hard to inagine how material belonging to the two different facies came to be so thoroughly mixed in any single epoch of sliding. It seems, rather, to call for repeated sliding and remixing, perhaps in a manner analogous to the formation of the argille scagliose (Merla, 1951; Maxwell, 1959). A tentative hypothesis for the formation of the Sykesville is shown in Figure 29.

LAUREL FORMATION

Name

Rocks now assigned to the Laurel Formation were first studied in reconnaissance by G. H. Williams (1895), but many years went by before they were systematically mapped (Cloos and Broedel, 1940; Cloos and Cooke, 1953). During this mapping the rocks were named Laurel migmatite, as they were best exposed along the Patuxent River above the town of Laurel. Subsequently the name was modified to Laurel gneiss (Chapman, 1942; Cloos and Cooke, 1953). In this report it is called the Laurel Formation, as it is a mappable metasedimentary unit which includes much schist in addition to gneiss.

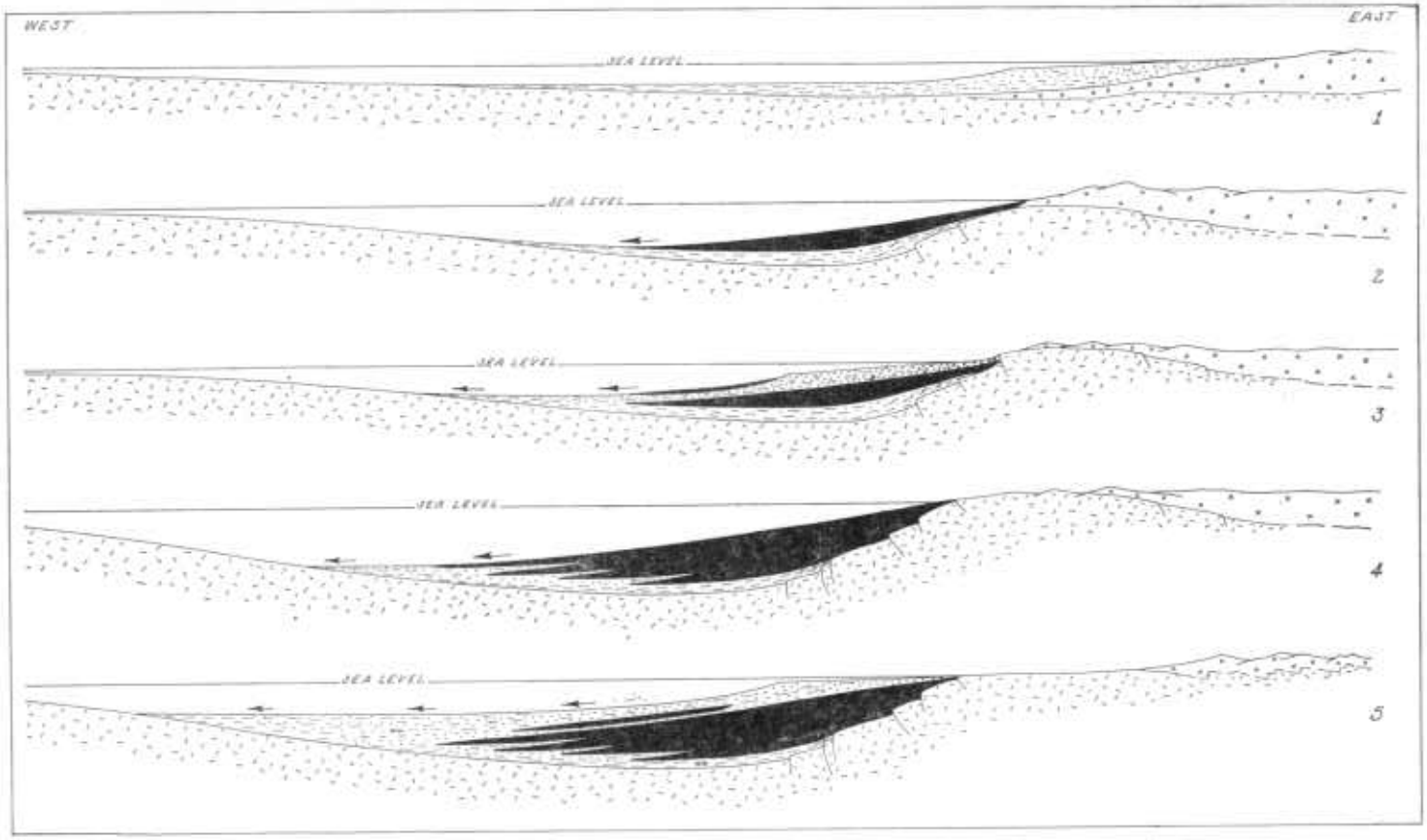


FIGURE 29. Schematic cross sections, showing possible origin of the Sykesville Formation. Five successive stages are shown diagrammatically: 1, beginning of geosynclinal (Wissahickon) sedimentation; 2, soft geosynclinal sediments slide from a rising geanticlinal ridge into an adjacent rapidly deepening trough; 3, geanticline continues to rise, shedding coarse clastic sediments (including quartz gravel and blocks of basement rock) into the trough; 4, continued rise of the geanticline and vigorous deepening of the trough, with the axes of both simultaneously migrating westward; earlier chaotic and bedded deposits, not yet firmly consolidated, sporadically slide down the steepening submarine slope and mix together; 5, basin subsidence ends, and sediments eroded from the geanticlinal highlands gradually fill the trough. Turbidity current deposition shown in stage 5 gradually changes to paralic sedimentation as the trough becomes filled and the shoreline migrates westward.

Symbols: hachures, basement rock; small crosses, unknown lithology in terrane to the east; stippling, medium to coarse clastic sediment; dashes, fine clastic sediment; black, chaotic submarine slide debris (Sykesville Fm.); arrows, turbidity currents; We, Wissahickon Fm., eastern sequence; Ww, Wissahickon Fm., western sequence.

Occurrence

The Laurel Formation emerges from beneath the sands and gravels of the Coastal Plain, between Washington. D. C. and the southern corner of Howard County. These sediments conceal a large part of the Laurel, as well as its contact with other crystalline rocks on the east and south. Along its western side the Laurel is in contact with the Wissahickon Formation (eastern sequence); this contact is gradational and concordant. In Howard County the Laurel is cut off by intrusive rocks of the Baltimore Gabbro Complex and the Guilford Quartz Monzonite.

Ernst Cloos first suggested that the Laurel and Sykesville Formations were part of the same mass, exposed on opposite limbs of a large, southward-plunging anticline. This was indicated by their striking lithologic similarity, by their similar stratigraphic position above the eastern Wissahickon sequence where it dips off the opposite sides of the Clarksville gneiss dome, and by their southward convergence toward Washington, D. C., where they almost join.

Subsequent work in the area (Fisher, 1963; Hopson, 1963) supports this belief. Fisher's mapping along the Potomac River, especially on the Virginia side, shows that the Sykesville comes very close to connecting with the Laurel; only a short interval of Coastal Plain Sediments covers the actual junction (Fig. 25). The rocks of the two formations are indistinguishable near here, both in the field and petrographically. It is interesting to note that back in 1895 G. H. Williams (p. 665) thought these rocks were the same:

> " They may best be studied along the Potomac, between Washington and the Chain Bridge, where they are extensively exposed in the great quarries [Sykesville Fm.] They are also typically shown near the north end of Analostan Island [Sykesville Fm.], at the falls of the Patuxent west of Laurel, and along Sligo Branch, Piney Run, and the lower portions of Rock Creek, near Washington [Laurel Fm.] They may also be seen to advantage where Four Mile and Holmes runs have cut deeply into the Coastal Plain deposits in Fairfax County, south of Washington [Sykesville Fm.]."

In this report, therefore, the Sykesville and Laurel are considered stratigraphic equivalents that outline the southward plunging nose of the Baltimore anticlinorium. Their inferred stratigraphic relationship is shown in Figure 26.

Lithology

Most of the Laurel rocks differ little from those of the Sykesville Formation except in degree of metamorphism. The Laurel is similar mineralogically (compare Tables 28 and 29) and has about the same bulk composition (Fig. 27) as the Sykesville, and it contains the same distinctive suite of inclusions. Rounded quartz pebbles, fragments of pelitic schist, blocks and slabs of psammitic granulite, amphibolite, and basement gneiss are all present. The Laurel rocks are, however, generally more coarsely crystalline than the Sykesville, and metamorphic differentiation has caused greater segregation of the micas. The Sykesville typically has the "salt and pepper" appearance of a fine-grained granite, whereas the Laurel may display a coarsely segregated gneissic texture or may be spangled with very large muscovite porphyroblasts. Under the microscope the Laurel rocks are more crystalloblastic than the Sykesville, which commonly reveals its former clastic texture. Some mineralogical features also reflect the difference in metamorphism: chlorite is no longer stable in the Laurel, and staurolite appears in some of its most pelitic rocks.

Even the differences in metamorphism disappear, however, as the Laurel is traced

TABLE 29

MODAL ANALYSES, LAUREL FORMATION

	1	2	3	4	5	6	7	8	9	10	11	12
Quartz	44.8	54.6	57.3	46.1	62.4	21.0	35.1	34.2	38.9	47.3	36.0	59.5
Plagioclase	25.7	21.8	20.5	17.9	18.4	3.7	20.4	26.9	16.9	25.6	18.6	23.6
K feldspar	—	—	—	—	—	—	—	.1	—	—	—	—
Muscovite	18.7	15.5	14.0	26.4	12.7	42.0	31.3	21.0	28.3	16.3	27.1	10.3
Biotite	8.6	6.7	6.8	7.3	4.5	8.8	10.0	14.8	13.2	8.2	15.4	5.5
Chlorite	—	—	tr	0.6	tr	6.8	0.2	—	tr	—	—	—
Garnet	0.4	0.2	0.1	0.4	—	11.4	0.7	—	0.8	0.3	0.2	—
Staurolite	—	—	—	0.3	—	4.4	—	—	—	—	—	—
Epidote	0.5	0.3	0.4	—	1.2	—	—	—	—	—	—	—
Allanite	tr	—	tr	—	tr	tr	—	—	—	—	—	—
Tourmaline	—	—	—	—	tr	—	0.1	—	—	—	—	—
Calcite	—	—	0.2	—	—	—	—	—	—	—	—	—
Sphene	—	—	—	—	—	—	—	—	—	0.1	—	—
Apatite	0.1	0.5	0.2	0.1	tr	0.1	0.3	0.4	0.2	0.4	0.2	0.2
Zircon	tr	tr	tr	tr	tr	tr	tr	tr	tr	0.1	tr	tr
Opaques*	1.2	0.4	0.5	0.9	0.8	1.8	1.9	2.6	1.7	1.7	2.5	0.9
Total	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0
Points	120.4	1281	130.9	1279	1333	1346	1224	1325	1271	1273	1256	1282

(1) Gneiss (M-262). Sligo Branch at Seven Oaks, Montgomery County.
(2) Gneiss (M-263). Burnt Mills Creek near Burnt Mills, Montgomery County.
(3) Gneiss (H104-1). Burnt Mills Creek near Burnt Mills, Montgomery County.
(4) Gneiss (H102-4). Patuxent River, half a mile west of Laurel, Prince Georges County.
(5) Psammitic granulite bed (H102-2). Patuxent River, half a mile west of Laurel, Prince Georges County.
(6) Pelitic schist bed (H102-3). Patuxent River, half a mile west of Laurel, Prince Georges County.
(7) Gneiss (H103-1). Maryland Route 216, near All Saints Road, Howard County.
(8) Gneiss (103-2). Maryland Route 216, near All Saints Road, Howard County.
(9) Schist (H101-1). Murry Hill Road, half a mile north of Gorman Road, Howard County.
(10) Gneiss (H-11). Middle Patuxent River above Murry Hill Road, Howard County.
(11) Schist (H100-1). Middle Patuxent River above Murry Hill Road, Howard County.
(12) Psammitic granulite bed (H100-2). Middle Patuxent River above Murry Hill Road, Howard County.
* Chiefly magnetite.

southward into Washington, D. C. Grain size decreases, metamorphic segregation lessens, and chlorite becomes a prominent constituent. Both megascopically and microscopically the Laurel in Rock Creek Park (Washington, D. C.) is indistinguishable from Sykesville rocks along the Potomac River at Georgetown, Analostan Island, and Arlington.

Strongly distorted and disrupted bedding can still be recognized at a few localities in the Laurel. This is best displayed along the Patuxent River above Walker Branch, near Laurel. Beds of fine-grained psammitic granulite that resemble quartzite (Table 29, col. 5) and a bed of staurolite schist (Table 29, col. 6) are interstratified with semipelitic gneiss (Table 29, col. 4). The striking feature here is the evidence of great plasticity and mixing of the beds. The psammitic beds swirl irregularly, swell into bulbous masses, pinch out, or even bifurcate. The contacts may be sharp at one place but fade imperceptibly into the gneiss a few inches or a few feet away. Locally all traces of bedding give way to a chaotic mixture of material. These features strongly resemble some of the soft-sediment slump structures in the Wissahickon schist at Bear Island.

Origin

The Laurel Formation was originally mistaken for gneissic granite—just as the Sykesville was—because of the abundant inclusions which resemble xenoliths (Williams, 1895, p. 665). All geologists who have worked on the Laurel in recent years (Chapman, 1942; Cloos and Cooke, 1953; Fisher, 1963; Hopson, 1963) interpret it as a metasedimentary rock. I believe that the present metamorphic rock was derived from a chaotic mixture of fragmental rocks and pebbles in an unsorted matrix of sand, silt, and mud; that it was part of the same apron of submarine slide debris as the Sykesville Formation (Fig. 26). Much of the material could have come from sliding and mixing of Wissahickon sediments, but the quartz pebbles and fragments of basement rock are exotic. Most of the evidence for these conclusions is the same as that given for the Sykesville Formation.

Chapman (1942) was first to show that the Laurel was not igneous and that it could have been derived by flowage and mixing of stratified rocks of the Wissahickon Formation. He visualized this as a metamorphic process, however, involving intense tectonic deformation and recrystallization (1942, p. 1300):

> "... the Laurel gneiss ... is believed to have originated by granulation, flowage, and recrystallization of the Wissahickon formation under conditions of stress, high temperature, and abundant water. This is suggested by: (1) the mineralogical and chemical similarity of the two rocks, (2) the granulated and recrystallized texture of the gneiss, (3) the abundance of schist and quartzite remnants which do not show granulation or recrystallization, and (4) the complete gradation from Laurel gneiss into Wissahickon formation across the contacts."

The origin of the Laurel must be the same as the Sykesville, however, for they have the same structures and composition and are part of the same body. Chapman's explanation for the Laurel fails for the Sykesville. Metamorphic flowage so extreme as to transform the well-stratified Wissahickon metasediments into the completely chaotic Sykesville would have destroyed all trace of original texture, but the clastic texture of the Sykesville is well preserved (Pl. 25). These and other objections are overcome, however, if the flowage and mixing took place while the sediments were still soft. Most of Chapman's evidence that the Laurel was derived from the Wissahickon still holds, if transformation occurred before metamorphism.

The origin of the chaotic rocks is best worked out in the Sykesville, and in the Wissa-

hickon at Bear Island, where all stages in the disruption of beds can be seen and where primary clastic textures are still well preserved. Conclusions from these rocks can then be extended to the Laurel, which shows many of the same features but in which stronger recrystallization and less favorable exposures hide much of the critical evidence.

PETERS CREEK FORMATION

Name

The name Peters Creek schist was given by Jonas and Knopf (1921, p. 447) to "an upper member of the Wissahickon formation that is less highly anamorphosed than the Wissahickon itself." These rocks, exposed near Peters Creek, a tributary to the Susquehanna River in southern Lancaster County, Pennsylvania, were subsequently described (Knopf and Jonas, 1923, p. 46) as:

> ". . . a series of chloritic and sericitic quartzites, interbedded with chlorite-muscovite schist and grading toward the top into mildly anamorphosed quartzose and conglomeratic sediments. It comprises part of what was originally mapped as Wissahickon gneiss and part of the Octoraro schist."

In the Baltimore County report the name was modified to Peters Creek Formation, and the character of the rocks and their relation to the Wissahickon were further clarified (Knopf and Jonas, 1929, p. 176):

> "The formation consists of quartzite and mica schist. Quartzite is the dominant member, but the beds vary in thickness and the intervening layers of mica schist range from a few inches to several hundred feet in thickness. The conditions of deposition must have been irregular, so that accumulation of fine uniform quartz sand was frequently interrupted by muddy deposition. The sandstone and shale have been closely folded and recrystallized so that the arenaceous part of the formation has been transformed into a white to light greenish-gray crystalline quartzite with a subordinate amount of feldspar and a variable amount of muscovite, and the shale has become a muscovite-chlorite schist."

Also (1929, p. 177):

> "The field relations of the Peters Creek formation and the Wissahickon formation show that the two grade into one another and that they represent continuous deposition in a conformable sequence of formation. On the northwest side of the syncline occupied by the Peters Creek formation, the Wissahickon albite-chlorite schist grades upward into the schists and quartzites of the Peters Creek. Likewise there is a similar transitional relation between the Peters Creek formation and the oligoclase-mica schist that adjoins the syncline on the southeast side. The lithologic character of the Wissahickon and Peters Creek formations, together with their conformable relations indicates that they belong to one depositional period."

Thus defined, the Peters Creek Formation has been extended farther north into Pennsylvania (Gray, 1960) and southward through Maryland (Mathews, 1933) and deep into Virginia (Nelson, 1928).

Relation to the Wissahickon Formation

It is clear from these descriptions, and from viewing the Peters Creek at its type locality, that this formation differs from the enclosing Wissahickon mainly in having a higher proportion of sandy beds. Additional observations show their close genetic relation. The rocks called quartzites are mainly fine-grained metagraywackes (Table 30) and metasiltstones that occur in thin continuous beds that alternate rhythmically with the pelitic schist. They display graded bedding, thin-bedding laminations, and slump structures. It seems fairly evident that the sandy and silty beds were laid down by

TABLE 30

MODAL ANALYSES OF METAGRAYWACKE

Peters Creek Formation

	1	2	3	4	5
Quartz	43.7	44.6	36.8	44.6	37.3
Plagioclase	21.4	19.1	32.9	23.9	20.8
Muscovite	19.2	22.2	15.6	15.5	24.4
Biotite	10.6	8.5	12.1	5.6	11.2
Chlorite	0.2	1.8	0.2	7.2	4.1
Garnet	—	0.1	1.4	—	—
Epidote	1.8	1.0	0.3	1.6	1.0
Allanite	t	—	0.3	tr	tr
Tourmaline	—	—	—	tr	0.1
Apatite	0.2	0.6	0.3	—	0.2
Zircon	t	t	—	t	t
Opaques	2.9	2.1	0.1	1.6	0.9
Total	100.0	100.0	100.0	100.0	100.0
Points	1207	1257	1294	1440	1254

(1) B30-1. Westminster Road near Glen Falls, Baltimore County.
(2) B30-3. Westminster Road near Glen Falls, Baltimore County.
(3) Cl-2. Susquehanna River at Broad Creek, Harford County.
(4) CE2-2. Susquehanna River below Conowingo Dam, Cecil County.
(5) CE2-6. Susquehanna River below Conowingo Dam, Cecil County.

turbidity flows and related processes (*see* Lombard, 1963, p. 21–22; Bouma, 1962, p. 98–100), a concept unknown when Knopf and Jonas made their study.

Chaotic zones and other evidence of submarine slumping and sliding are also widespread in the Peters Creek Formation along the Susquehanna River, especially near Broad Creek and below Conowingo Dam. As one goes southward, the rocks mapped as Peters Creek Formation in Baltimore County (Knopf and Jonas, 1925) and Carroll County (Jonas, 1928) become increasingly chaotic, and in Carroll County they grade along strike into massive submarine slide deposits of the Sykesville Formation (Fig. 30). In Howard County the Peters Creek can no longer be recognized. There, and in Montgomery County farther south, the Sykesville Formation appears to have taken its place (Fig. 30). It will be recalled from earlier discussion that the Sykesville Formation forms a thick wedge within the Wissahickon Formation (Fig. 26). This suggests that the Peters Creek also lies within the Wissahickon rather than above it, as Knopf and Jonas (1923; 1925; 1929) had supposed.

Along the Potomac River metagraywacke sequences are virtually identical to the Peters Creek Formation at its type locality. Exposed at Bear Island (Pl. 11, fig. 1), Rocky Island, and at localities above Great Falls, these have been regarded as sandy zones within the Wissahickon, and no one in recent years has attempted to correlate them with the Peters Creek (Cloos and Anderson, 1950; Cloos and Cooke, 1953; Fisher, 1963; Reed and Jolly, 1963; also this report). There are at least two such sandy zones, at different stratigraphic horizons (Fisher, 1963, Pl. 1), and neither can be traced northward in Montgomery County for any great distance.

The Wissahickon appears to have been a great complex of deep submarine deposits, formed by gradual accumulation of mud and fine silt interspersed with periodic influx

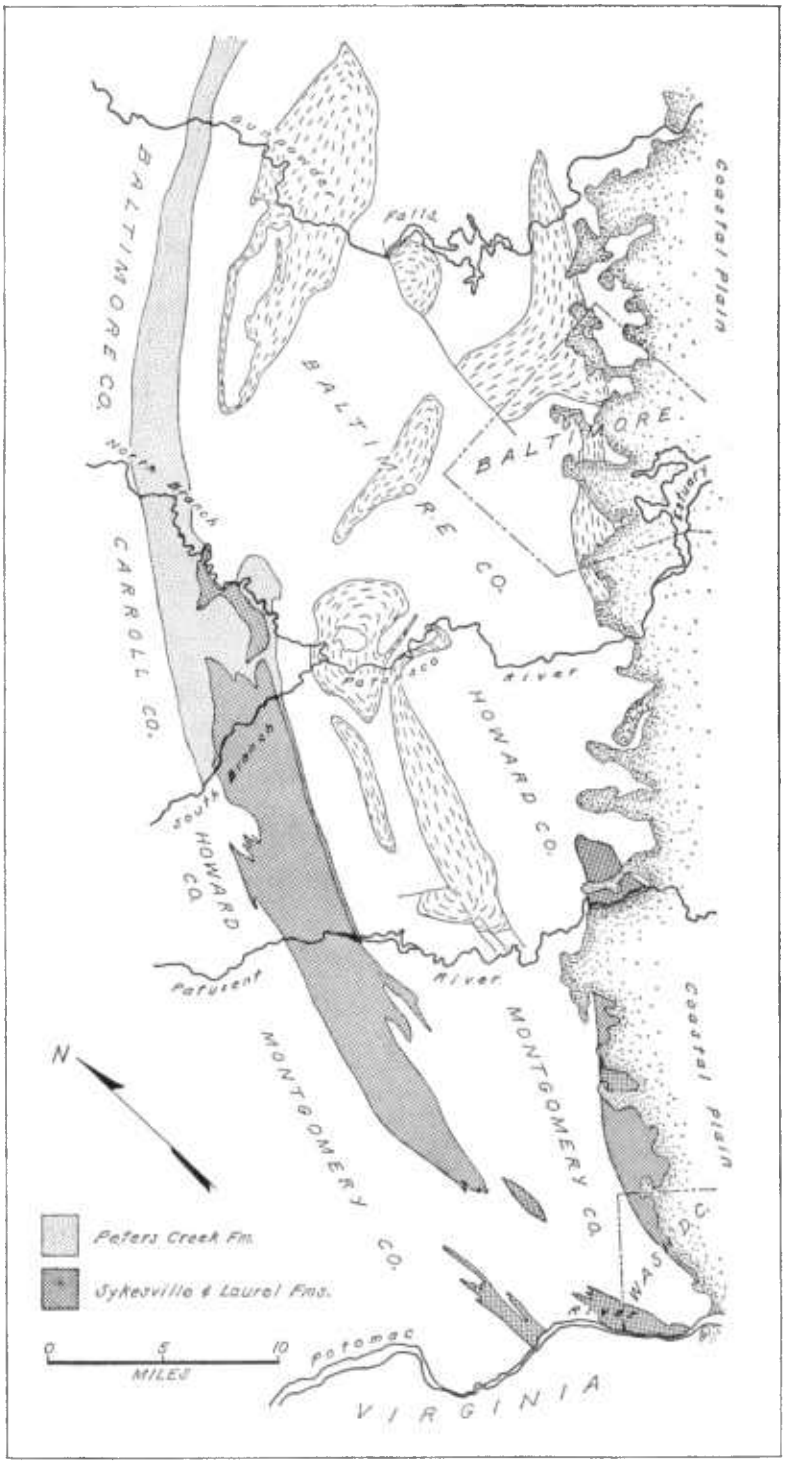


FIGURE 30. Generalized geological map of parts of Montgomery, Howard, Carroll, and Baltimore Counties, showing relationship of the Peters Creek and Sykesville Formations. Stratified turbidites and laminites of the Peters Creek Formation become chaotic toward the south and grade laterally into submarine slide deposits of the Sykesville Formation. Geology taken from the Maryland Geological Survey maps of Baltimore, Carroll, Howard, and Montgomery Counties.

of coarser silt and sand brought in by turbidity currents, and with submarine slides and mud flows. In my opinion the Peters Creek, like the rocks at Bear Island, is a sandy turbidite facies within this complex. These turbidite sequences do not appear to have been continuous sheets but broad lenticular zones, occurring at different stratigraphic horizons. They grade laterally into chaotic slide deposits and into finer, more pelitic beds. Perhaps a close modern analogue are the deposits building up in the off-shore basins along the southern California coast (Gorsline and Emery, 1959). There, sand and chaotic slide debris form aprons along the base of the steep submarine slopes. Turbidity flows, emerging from the mouths of submarine canyons, move out across these aprons and spread broader fans of graded sands and silts farther out. Between and beyond the turbidite fans still finer silt and mud are accumulating. The lenticular turbidite zones within the Wissahickon, among which I include the Peters Creek Formation, may correspond to some of these fanlike deposits. This picture suggests that long-distance correlations with the Peters Creek Formation have little value.

CRYSTALLINE ROCKS OF WESTERNMOST MONTGOMERY AND HOWARD COUNTIES

STRATIGRAPHIC PROBLEMS

Rocks of the western Piedmont, lying between the Wissahickon Formation and the Cambro-Ordovician sedimentary rocks of Frederick Valley, are chiefly phyllites with interbedded metasandstones, metavolcanic rocks, and marble. The metavolcanic rocks and marble occur mainly in Carroll County and in Frederick County north of U. S. Route 40, and they greatly diminish along strike to the south. In western Montgomery County nonvolcanic phyllites and metasandstones predominate.

Despite geological work in this region since 1890 there is still disagreement about the stratigraphic and structural relationships. The uncertainties stem from poor exposures, complex structure, lack of fossils, and lithologic similarity throughout much of the section.

Sugarloaf Mountain, a mile beyond the Montgomery County line, provides the last stratigraphic link with the known Paleozoic rocks farther west. Massive quartzites which hold up the mountain have been correlated with the Lower Cambrian Weverton Quartzite on near-by Catoctin Mountain by nearly all geologists who have worked in the area (Keyes, 1891; Williams, 1892; Jonas, 1924; Jonas and Stose, 1938; Cloos and Hietanen, 1941). Stose and Stose (1946) questioned this correlation, reversing their earlier conclusion, but it has been reaffirmed by more recent investigators (Scotford, 1951; Thomas, 1952; Cloos and Cooke, 1953). Thus the Sugarloaf Mountain Quartzite is a key unit to which the stratigraphic section in the western Piedmont may be tied.

Different structural interpretations of Sugarloaf Mountain, however, have led to conflicting interpretations of the surrounding stratigraphy. If the mountain is synclinal (Jonas and Stose, 1938; Stose and Stose, 1946; 1951) then the rocks along its eastern flank (called Urbana phyllite by Jonas and Stose, 1938a) would lie stratigraphically below the Sugarloaf Mountain Quartzite. Cloos and his students (Scotford, 1951; Thomas, 1952) have found good evidence that the mountain is anticlinal, however, which puts this phyllite above the quartzite. They accordingly correlated it with the Lower Cambrian Harpers Phyllite, which overlies the Weverton Quartzite at Catoctin Mountain. The Ijamsville Phyllite, which borders the Harpers on the east, was thought to be still higher in the section, since the thick Sugarloaf Mountain Quartzite does not reappear between them (Cloos and Cooke, 1953, section A-A′).

Stose and Stose (1951) have objected, on the grounds that the phyllites east of Sugarloaf Mountain contain volcanic material and are therefore unlike the Harpers or other formations above the Weverton. Traced north into Frederick County the phyllites show relict tuffaceous textures and interfinger with metabasalt and metarhyolite. The Frederick County metavolcanic rocks in turn are correlative with the Catoctin metavolcanics, which underlie Weverton Quartzite on Catoctin Mountain. This evidence suggests that the phyllites east of Sugarloaf Mountain are stratigraphically lower than the Lower Cambrian quartzite.

More detailed work is needed in the Sugarloaf area to resolve these problems. My own reconnaissance work in the area east of Sugarloaf Mountain was brief and concerned more with lithology than with regional structure.

IJAMSVILLE PHYLLITE

Name

Green and purple phyllitic slates associated with metabasalt, metaandesite, and metarhyolite in Frederick County was named Ijamsville Phyllite by Jonas and Stose (1938a, p. 346). The name was taken from Ijamsville, a small community 6 miles southeast of Frederick, where the rock was quarried for roofing slate. It was distinguished from other phyllites in the area because of its supposed tuffaceous origin.

During the geologic mapping of Montgomery County (Cloos and Cooke, 1953) it was not found practical to separate volcanic and nonvolcanic phyllites. Therefore all the phyllitic rocks along strike with the Ijamsville of Frederick County were given this name, as well as the other phyllites and interbedded rocks that were indistinguishable from them in the field. This includes rocks mapped as Marburg schist in Frederick County. Similar rocks in western Howard County directly on strike with the Ijamsville of Montgomery County are shown as Ijamsville Phyllite on Plate 7. They were previously called Wissahickon.

Occurrence

Thus defined, the Ijamsville Phyllite covers a wide area in western Montgomery County, as well as the western tip of Howard County. It is bounded on the east by the Wissahickon Formation and on the west by the Harpers Phyllite (Urbana phyllite of Jonas and Stose, 1938). On the south it passes beneath the unconformable cover of Triassic sedimentary rocks 6 miles north of the Potomac River.

Its contact relations with the Wissahickon Formation have been described. Graded bedding in the Wissahickon near the Ijamsville contact at Woodfield and also at Blockhouse Point shows tops of beds to the west, which suggests that the Ijamsville is stratigraphically higher than the Wissahickon.

The contact with the Harpers Phyllite in Montgomery County is not well exposed, but it appears to be gradational and conformable. Cross-bedding in the Harpers and cleavage-bedding relationships in the Ijamsville, in quartzites which border the contact both west and south of Barnesville, indicate that the beds are overturned and their tops are to the west. This suggests that the Harpers overlies the Ijamsville at this locality. However more observations are needed.

Lithology and Field Relations

The formation consists chiefly of silvery phyllite and green to purple phyllite and slate interbedded with metasiltstone and metagraywacke. A few white quartzite beds also

occur. Most of the quartzites are only a few inches to a few feet thick, but toward the western side of the formation—presumably the top—at least one quartzite bed is more than 50 feet thick. Greenschist, derived from mafic volcanic rock, occurs in the Ijamsville near Hyattstown but appears to die out about 1½ miles north of Comus.

The phyllites are poorly exposed but appear to form uninterrupted sequences tens or even hundreds of feet thick. Cleavage is commonly the only structure, but fine bedding laminations may also be visible. Purple, green, and gray layers in the slate also mark bedding. Elsewhere the phyllite is interbedded with dark-gray metasiltstone or metagraywacke. The metagraywacke is mostly very fine-grained, but locally it becomes coarse. Unlike metagraywackes in the Wissahickon those in the Ijamsville do not show graded bedding, convolute bedding, or slump structures. The bedding is generally even and regular, but in some places it is discontinuous or lensoid. Occasionally coarse graywackes are in beds more than 10 feet thick. Thus the bedding characteristics of the Ijamsville and Wissahickon metagraywackes are different, although they are lithologically similar.

Most of the quartzites are massive, although some have traces of bedding. The thick quartzites in the western part of the formation appear to have cross-bedding, but it is poorly shown.

Petrography

The slaty and phyllitic rocks are composed chiefly of muscovite, with chlorite, magnetite, and quartz as lesser constituents. Paragonite and chloritoid are also common. Minor minerals include tourmaline, iron oxides, leucoxene, or rutile. Common mineral assemblages in the most aluminous phyllites and slates are:

(1) muscovite-chlorite
(2) muscovite-paragonite-chlorite
(3) muscovite-paragonite
(4) muscovite-chloritoid-chlorite-paragonite
(5) muscovite-chloritoid-chlorite.

Magnetite and quartz may accompany all these assemblages, and hematite may also be present. It is difficult, however, to tell metamorphic hematite from that which has formed by weathering. The quartz is commonly segregated in thin layers, lenses, or knots.

The paragonite is in very tiny flakes and was identified only by X-ray diffraction. It is always greatly subordinate to muscovite and has not been found with albite.

Chloritoid is in very tiny prisms, often visible only under the microscope. It is greenish blue in thin section and commonly shows well-developed polysynthetic twinning. The prisms are randomly oriented, and many lie athwart the cleavage (Pl. 26, fig. 1). Where the phyllite is crinkled by a later discordant cleavage the chloritoid was generally earlier and was bent, broken, or twisted by the shearing.

Magnetite occurs in porphyroblasts up to 2 mm in diameter that are conspicuous in hand specimen. In thin section these may show striking sieve structure, enclosing trains of quartz grains that preserve the trend of the enclosing cleavage, as well as randomly oriented chloritoids.

Some phyllites, evidently derived from more silty muds, are rich in quartz. These may also contain albite and sparse stilpnomelane. The common mineral assemblages are:

(1) muscovite-albite-chlorite (+ quartz)
(2) muscovite-albite-chlorite-stilpnomelane (+ quartz)

(3) muscovite-chlorite-stilpnomelane (+ quartz)
(4) muscovite-albite-chlorite-chloritoid (+ quartz)

Magnetite, hematite or leucoxene may also be present.

The stilpnomelane forms very tiny brown flakes and shreds. It is easily told from the green chlorite but may resemble biotite; however, it does not show "birds eye" extinction. It is generally a distinctive "sooty" brown to almost black, but in a few rocks it is deep olive.

Very fine bedding laminations, marked by layers richer in opaque minerals or in quartz, are well displayed in thin section. Features suggestive of volcanic origin are remarkably rare. Phyllites from the area around Hyattstown, however, contain small blebs rich in muscovite, chlorite, and stilpnomelane that appear to have been relict pumice lapilli.

The metasandstones consist of quartz and lesser plagioclase, in a fine matrix of muscovite, chlorite, quartz, and albite. Small amounts of stilpnomelane may also be present, together with magnetite and various heavy minerals. The original clastic textures are well preserved, showing that the original sandstones were fine-grained, with unsorted angular grains set in argillaceous matrix. These rocks were typical graywackes. There were all gradations from fine-grained graywackes through siltstones to silty mudstones. The graywackes show no evidence of admixed volcanic material. Relict pumice lapilli, shard structures, particles of volcanic rock, and other features of volcanic sandstones are lacking. The relict plagioclases show no evidence of zoning or complex twinning, and they were quite sodic even before metamorphism; they thus appear to have come from a metamorphic rather than a volcanic source.

Greenschist near Hyattstown shows good evidence of volcanic origin. It was evidently an andesitic tuff, now converted to chlorite, albite, epidote, and calcite. The original lath shape of the plagioclase is well preserved, and remnants of squashed and stretched pumice lapilli, now replaced by chlorite and calcite, are still visible.

The quartzites were derived from quartz sands. Those that occur in thin beds in the western part of the formation were poorly sorted, with subrounded to angular grains and abundant sericitic matrix. Those forming thick beds farther west were composed of well-rounded sand grains about 1 mm in diameter, in a siliceous matrix. These were cleanly washed orthoquartzites.

Origin

The Ijamsville slate and phyllite in Montgomery and Howard Counties were derived chiefly from nonvolcanic mud and silt, rather than from tuff. This is shown by the lack of pumiceous blebs, lath-shaped feldspars, and other relict volcanic features, and also by their chemical composition (Table 27, col. 5). They are exceptionally rich in a alumina but low in alkalis, and the K_2O/Na_2O ratio is high. These chemical features, which are reflected in the high muscovite content of the phyllites, are typical of clay shales but not of volcanic rocks. Moreover, the interbedded silts and sands also lacked volcanic testures or volcanic plagioclase. Some tuffaceous slates may be present, like those near Hyattstown that display pumiceous blebs, but they do not seem to be abundant.

These conclusions do not imply that the Ijamsville is not tuffaceous in Frederick and Carroll Counties. Slates with smeared-out pumiceous blebs appear to be more common there. The Frederick County map shows volcanic material greatly diminishing toward the south, and relatively little of it near the Montgomery County line. Also, much of

the rock mapped as Ijamsville in Montgomery County is a continuation of Marburg schist, which was not proven to be volcanic in Frederick and Carroll County.

The sandy beds in the Ijamsville seem to record a change in depositional environment from that of the Wissahickon. Graded beds, convolute beds, and slump structures are lacking, which suggests that sediment transport by gravity mechanisms such as turbidity flows, slumping, and sliding was no longer important. Many of the beds are still even and regular, but some discontinuous and lensoid bedding is also in evidence. Some sandy beds are many feet thick. Toward the western part of the formation beds of clean-washed quartz sand was present—a feature not found in the Wissahickon Flysch. The cross-bedding in some of this quartzite indicates deposition in an environment of current action.

These and other features suggest that the change from Wissahickon to Ijamsville sediments represents a change from Flysch to Molasse. The change is not sharply defined, however, and perhaps much of the Ijamsville was transitional. In the Harpers, however, the change becomes clear cut.

HARPERS PHYLLITE

Name

Harpers Shale (also called Phyllite) was named in separate publications by Williams and Clark (1893) and by Keith (1894) for exposures in the Potomac River gorge near Harpers Ferry, West Virginia. Keith (1894, p. 333) gives the following description:

> "The shales when fresh are dull bluish gray and weather to a light greenish gray. In composition they are argillaceous and sandy, although the sandy feature is not pronounced except at certain horizons. Taking the southern half of the belt . . . there is scarcely any variation from one bed to another, and the whole is a uniform, argillaceous, slightly sandy shale. Northward from the Potomac, beds of sandstones and sandy shales begin to appear, especially in the upper portion of the shale. In the northern portion of the belt, toward the Pennsylvania line, these beds of intercalated sandstones attain a considerable thickness, varying from 4 or 5 up to 50 feet."

Phyllites along the Little Monocacy River at the far western edge of Montgomery County are shown as Harpers Phyllite on the Geologic Map of Montgomery County (Cloos and Cooke, 1953). Scotford (1951, Pl. 1) had earlier applied that name not only to these rocks but to a much broader area of phyllites and related rocks surrounding Sugarloaf Mountain. As already noted, the correlation of these phyllites with the known Harpers at Catoctin Mountain and Harpers Ferry rests on the structural interpretation of Sugarloaf Mountain: if it is an anticline these phyllites overlie the Sugarloaf Mountain Quartzite, considered correlative with the Weverton Quartzite which underlies the Harpers Shale at its type locality.

Jonas and Stose (1938; 1938a, p. 346) proposed the name Urbana phyllite for the same group of rocks, after the town of Urbana in Frederick County. They believed that the mountain is synclinal and that the phyllites around it lie stratigraphically beneath the Sugarloaf Mountain Quartzite (Jonas and Stose, 1938; 1938a; Stose and Stose, 1946; 1951).

This report follows the Montgomery County map and tentatively retains the name Harpers Phyllite, but the need for further work to resolve this conflict is recognized.

Occurrence

The Harpers Phyllite is quite restricted in Montgomery County, forming only a small area along the upper Little Monacacy River just inside the county line. About half a mile east of this stream the formation is in contact with the Ijamsville Phyllite, while to the south and west of Dickerson it is covered by sandstone and shales of the Triassic New Oxford Formation. On the northwest the Harpers extends across the line into Frederick County, where the formation has its greatest extent (mapped as Urbana Phyllite, Jonas and Stose, 1938).

Lithology and Field Relations

Dark-gray phyllites, metasiltstone, and various impure metasandstones predominate. Some of the phyllites and metasandstones are calcareous, and a few thin lenses of impure marble occur locally. Thick, ledge-forming beds of massive white quartzite are prominent but subordinate members of the sequence. The beds are closely folded and cleavage is developed in the micaceous rocks, but deformation has done little to obscure primary textures and structures.

The phyllites are dark sericitic-chloritic rocks with discordant cleavage. In general they are coarser-grained than the typical slaty phyllites in the Ijamsville; many of the so-called phyllites in the Harpers were derived from silty rocks and fine-grained sandstones, rather than clay shale. Thin, dark-gray bedding laminations, brought out by concentrations of finely divided magnetite and leucoxene, stand out at many outcrops. The calcareous phyllites are often reddish and crumbly, owing to oxidation and solution of calcite.

The metasandstones range widely in texture, composition, and mode of occurrence. Dark-gray to greenish-gray metagraywackes are most common. Typically these are rather fine-grained, and they grade into the silty phyllite. The metagraywackes and metasiltstones occur in beds ranging from a few inches to more than 10 feet thick. The thick beds show crude stratification, formed by alternating layers of slightly different grain size and color. The alternations are random, rather than cyclic, and the contacts between layers are gradational. Many of the individual layers lense out within a few feet. Some layers may be somewhat calcareous; these have a pitted appearance, from differential solution of calcite.

Cross-bedding is developed in some of the metagraywacke and siltstone units, but because of the dark color one must look closely to see it. Ripple marks, apparently of the oscillation type, are well preserved on the base of a fine-grained metagraywacke bed along the Little Monocacy River, southeast of Dickerson (Pl. 26, fig. 2). At the same locality cross-bedding and channel structure are found.

Sericitic quartzites and quartzose phyllites are also abundant. These were derived from fine- to medium-grained sandstones that were richer in detrital quartz. They occur in thin beds and lenses in metagraywacke, as well as in massive beds many feet thick. Thin, bluish laminations rich in magnetite are found in some of the quartzose metasandstones; others show crude cross-bedding.

Quite distinct from the sericitic and phyllite quartzites are the massive white quartzites, which occur in thick, resistant beds that form ledges and hold up low ridges. The thicker of these beds are shown separately on the Geologic Map of Montgomery County (Cloos and Cooke, 1953). They range from 10 to 30 feet or more thick. Generally they are massive, but in places bedding planes and fairly coarse cross-bedding can be recog-

nized. Similar quartzites on Sugarloaf Mountain have well-preserved current ripple marks. Relict clastic grains, medium-sized and well rounded, are plainly visible with a hand lens or even with the maked eye.

Petrography

The fine phyllites are composed chiefly of quartz, sericite, and chlorite, with subordinate magnetite. Albite may be present, and tourmaline, ilmenite, hematite, and zircon are the common accessories. Thomas (1952, p. 57) describes the calcareous phyllites as composed of granular calcite, albite, quartz, ilmenite, and sphene, in thin laminae that alternate with schistose chloritic layers.

The metagraywackes consist of poorly sorted, dominantly angular and subangular clastic grains of quartz and plagioclase in a matrix of muscovite, chlorite, and fine quartz and magnetite (Table 31, cols. 10, 11, 12). Small amounts of stilpnomelane are generally present, and the accessory minerals include tourmaline, apatite, zircon, monazite, xenotime, rutile, hematite, and leucoxene. The matrix may be weakly schistose, but the relict clastic texture is well preserved (Pl. 27, fig. 1). The relict plagioclase is albite, and the lack of associated epidote indicates that the plagioclase was not much more calcic originally. None of the relict plagioclase has lath-shaped outlines, relict zoning, complex twinning, or other indication of volcanic origin. Moreover, the rest of the rock contains no grains of volcanic rock, relict pumice lapilli, or other volcanic features. These sands appear to have had chiefly a metamorphic source, like the Wissahickon graywackes.

The metagraywackes show all gradations into other metasandstones that were better sorted, better rounded, had less argillaceous (micaceous) matrix, and had chiefly quartz as detrital grains. This includes the sericitic quartzites, which consist chiefly of moderately well sorted, subangular to subrounded quartz grains, packed closely together in a weakly schistose matrix of sericite and finely crystalline quartz (Table 31, col. 9). The original matrix was partly silica and partly argillaceous paste. Calcite was an important matrix constituent in some. These rocks were formerly protoquartzites (Pettijohn, 1957, p. 316–321). Other metasandstones, less well rounded and sorted, and with more clastic feldspar, were originally subgraywackes.

The white, thick-bedded quartzites are composed of well-sorted, well-rounded detrital quartz grains in a finely crystalline quartz matrix (Pl. 27, fig. 2). Sericite and iron oxides plus a few heavy minerals are minor matrix constituents (Table 31, cols. 7, 8). Texturally and mineralogically these rocks are identical with the Sugarloaf Mountain Quartzite (Table 31, col. 6). Some of the rounded clasts are second-generation quartz grains, consisting of worn overgrowths enclosing an earlier detrital grain. These quartzites came from cleanly washed, mature quartz sands, cemented by silica. They would be classed as orthoquartzites by Krynine (1948) and Pettijohn (1957).

Origin

The outstanding feature of the Harpers in Montgomery County is the abundance of rocks derived from immature fine-grained sands and silts. Much of the detritus was unsorted, poorly rounded graywacke sands and silts, with abundant clastic feldspar and argillaceous matrix. These materials evidently record rapid erosion and deposition, with little opportunity for prolonged weathering or reworking. Perhaps they were shed from an area of high relief. Interbedded with this material were other fine sands with better sorting and rounding, and predominance of stable quartz as detrital grains. All

TABLE 31

MODAL ANALYSES OF QUARTZITE AND METAGRAYWACKE

Ijamsville and Harpers Formations

	1	2	3	4	5	6	7	8	9	10	11	12
Quartz	76.6	75.2	84.3	87.5	98.2	97.8	93.2	96.9	70.1	55.2	39.4	6.5
Albite	8.0	—	—	4.7	—	—	—	—	.1	11.7	12.0	33.4
Muscovite	9.2	24.4	15.4	5.3	1.5	2.2	5.6	0.1	18.2	8.9	16.1	42.5
Biotite	—	—	—	0.2	—	—	—	—	—	—	—	—
Chlorite	3.1	—	—	2.2	—	—	—	—	9.0	17.2	20.4 }	6.4
Stilpnomelane	—	—	—	—	—	—	—	—	tr	—	1.6 }	
Epidote	1.0	—	—	—	—	—	—	—	—	—	—	—
Tourmaline	.2	tr	—	tr	—	—	—	—	tr	tr	tr	tr
Apatite	—	tr	tr	tr	tr	—	tr	—	—	tr	0.1	0.3
Sphene	—	—	—	—	—	—	—	—	—	—	—	0.7
Zircon	—	tr	tr	tr	tr	tr	—	tr	tr	tr	tr	tr
Heavies*	—	tr	tr	tr	tr	tr	tr	tr	tr	tr	0.1	0.2
Rutile	—	—	—	—	0.1	—	—	—	—	—	tr	—
Magnetite	1.1 }	0.4	0.1	0.1	—	tr }	1.1	3.0	2.4	6.2	9.3	8.9
Hematite	— }		—	—	0.1	— }			—	—	—	—
Limonite	0.8	—	—	tr	—	—	0.1	—	0.2	—	0.2	.2
Leucoxene	—	tr	0.2	—	0.1	tr	—	—	—	0.8	0.8	0.9
Total	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0
Points	1674	1159	1379	1518	1514	1342	1151	1141	2886	3886	4738	3337

Ijamsville Phyllite

(1) Feldspathic quartzite (M298-2). Woodfield, Montgomery County.
(2) Sericitic quartzite (M472-1). West of Barnesville, Montgomery County.
(3) Sericitic quartzite (M472-2). West of Barnesville, Montgomery County.
(4) Quartzite (H131-5). Cabin Creek, 1½ miles west of Florence, Howard County.
(5) Quartzite (M418-1). Md. 109, 1½ miles north of Comus, Montgomery County.

Sugarloaf Mountain Quartzite

(6) Quartzite (M477-1). Summit of Sugarloaf Mountain, Frederick County.

Harpers Phyllite

(7) Quartzite (M473-1). One mile northeast of Dickerson, Montgomery County.
(8) Quartzite (M476-1). One mile southeast of Dickerson, Montgomery County.
(9) Sericite quartzite (M473-3). One mile northeast of Dickerson, Montgomery County.
(10) Metagraywacke (M475-1). Half a mile southeast of Dickerson, Montgomery County.
(11) Metagraywacke (M475-2). Half a mile southeast of Dickerson, Montgomery County.
(12) Metagraywacke (M474-2). One mile northeast of Dickerson, Montgomery County.

* Chiefly monazite, xenotime, and anatase.

gradations from graywacke to subgraywackes and protoquartzites are present. Interbedded with this material was abundant fine silt and mud.

The graywackes and other sands show clear evidence of shallow-water deposition. Cross-bedding, ripple marks, channeling, lensoid bedding, and other features point toward deposition in an environment of current action. The thick beds of stratified, nongraded sand must have build up gradually, by relatively continuous deposition. This indicates sediment transport by continuously flowing currents, a phenomenon characteristic of shallow-water environments.

Muds and local zones of limestone deposition were interspersed with the sands and silts. Perhaps these limy beds accumulated in lagoons or esturaries, or other places temporarily protected from coarser clastic deposition.

This assemblage of shallow-water deposits is characteristic of the molasse facies and contrasts strongly with the Wissahickon flysch sediments farther east. Judging from descriptions by Trümpy (1960, p. 880–881) it resembles, in a general way, some of the molasse sequences marginal to the Alps.

The few thick beds of cleanly washed orthoquartzite are obviously out of place in this pile of immature clastics. These sands were the product of prolonged weathering, winnowing, and abrasion. The detrital quartz grains consisting of worn overgrowths around earlier clastic grains show derivation from an area having other quartz sandstones. One can hardly escape the conclusion that the orthoquartzites were brought in from a source different from that which supplied the molassic sediments constituting the bulk of the formation.

There is little evidence of volcanic debris in the Harpers sediments in Montgomery County. Scotford (1951), who had reached a similar conclusion, was criticized by Stose and Stose (1951), who regard the Urbana phyllite (Harpers of this report) partly as a pyroclastic facies of the metabasalt exposed in Frederick County. As evidence they cite the abundance of chlorite in the phyllites and their interfingering relation with the metabasalts farther north in Frederick County. They placed special importance on epidote and actinolite-bearing phyllites that have basaltic composition (few of these extend into Montgomery County). They add (Stose and Stose, 1951, p. 698):

> "Due to shearing, volcanic textures are largely destroyed in the epidotic phyllites near Sugarloaf Mountain, and Scotford, not finding such textures, doubts their volcanic origin."

Some beds are sheared, especially the incompetent argillites, but relict clastic textures are so well preserved (Pl. 27, fig. 1) in most of the beds that it is impossible for volcanic features to go undetected. But such features are rarely present; the plagioclase is nonvolcanic, and there are no relict pumiceous fragments or particles of aphanitic volcanic rock. Small contributions of fine tuffaceous material mixed in the graywacke matrix could pass undetected, but it is very plain that the bulk of the Harpers sediment in Montgomery County was not derived from a volcanic source. This, however, does not rule out the possibility of contemporaneous volcanic sedimentation farther north.

PIEDMONT SEDIMENTATION—SUMMARY AND INTERPRETATION

Montgomery and Howard Counties lie astride the Piedmont metamorphic belt in Maryland and span the complete section of stratified crystalline rocks. These form a sequence on the order of 25,000 feet thick. Quartzite and marble at the base rest on older Precambrian basement in Howard County. In western Montgomery County phyl-

lite and metasandstone at the top of the section are partly equivalent to the Lower Cambrian Chilhowee Series or lie just beneath it. The inferred stratigraphic relations are shown in figure 31. Metamorphism has blurred but not obliterated the sedimentary record, for original structures and textures are preserved in even the higher-grade rocks. The following is a brief summary of this record.

The crystalline basement (Baltimore Gneiss) was eroded to a surface of low relief when Glenarm sedimentation began. Subsidence and flooding were followed by deposition of quartz sand and potassic silt and mud (Setters Formation). Carbonate sediments, largely dolomite, came next (Cockysville Marble). Together they formed a blanket approximately 1500 feet thick. These sediments reflect deposition under fairly stable tectonic conditions.

The overlying Wissahickon Formation records the development of a deep submarine trough, in which shale, graywacke, and chaotic slide deposits accumulated to a thickness of more than 20,000 feet. The lowest part of the formation was mainly shale, perhaps deposited as the trough gradually deepened. Thin siliceous beds that may have been chert appear higher in the shale and might indicate periods of starved sedimentation. Above the lower 1000–2000 feet of the formation thin silty and sandy beds become common and alternate rhythmically with the shale. This characterizes the rest of the Wissahickon section, although the proportion of shaly to sandy beds varies widely, and thick intervals of pure shale recur. Slump structures, sandstone dikes, disrupted bedding, and chaotic zones formed by sliding of soft sediments are widespread and may indicate tectonic activity during deposition. These movements culminated in submarine sliding on an enormous scale: the Sykesville and Laurel Formations, perhaps up to 15,000 feet thick, are a wedge of chaotic slide material within the Wissahickon. Evidently great masses of unconsolidated sediment slid from a rising submarine ridge or scarp into a deep, rapidly subsiding trough, forming a thick apron of chaotic debris. This material grades westward into Wissahickon turbidites.

Metasediments in the western part of Montgomery County appear to lie conformably above the Wissahickon and to represent a gradual upward and westward change from deep marine to paralic sedimentation. The Ijamsville Phyllite was chiefly shale, but it contains sandy interbeds that lack the features of turbidites. Among these are quartz sandstones with cross-bedding. The formation is poorly exposed, however, and some argillaceous parts of it may represent the distal end of the Wissahickon, brought up in the crests of anticlines. The Harpers (Urbana?) Phyllite is a higher, much sandier formation, with cross-bedding, ripple marks, channeling, discontinuous and lensoid bedding, and other features of paralic deposits. The metasandstones are still partly graywacke, but subgraywacke and protoquartzite are also important. Some of the sands and silts are calcareous, and thin lenses of limestone appear locally. Thick beds of massive and cross-bedded orthoquartzite interfinger with the less mature sands and silts and were probably derived from a different source.

Volcanic activity accompanied sedimentation in the western Piedmont. Metabasalts in Frederick and southwestern Carroll Counties are evidently an eastward extension of the Catoctin metabasalt, which is much thicker farther west along the South Mountain (Blue Ridge) uplift. The Ijamsville and Urbana Phyllites partly interfinger with the metabasalt and more siliceous metavolcanic rocks in Frederick and Carroll Counties and are partly tuffaceous in that area (Stose and Stose, 1946).

Nevertheless, volcanic rocks and tuffaceous sediments form but a small proportion of the strata that fill the Piedmont trough. The Catoctin volcanics, erupted partly

from dike swarms along the Blue Ridge (Reed, 1955), encroached upon and interfingered with the western Piedmont sediments, but only after the trough was nearly filled (Fig. 31). The filling was chiefly by nonvolcanic sand, silt, and mud. Greenschist and amphibolite derived from mafic aquagene tuffs occur locally in the Wissahickon but comprise only a very small part of the section. Numerous large mafic and ultramafic ophiolitic intrusions are present, however, and the mafic tuffs may be a submarine extrusive phase of the same magma.

It is well known that geosynclinal sedimentation commonly follows a cyclic pattern (Bertrand, 1897; Krynine, 1942; Pettijohn, 1957). The usual succession of facies characterizing each cycle is (Pettijohn, 1957, p. 636–644): (1) a preorogenic facies, chiefly orthoquartzite and carbonates; (2) a euxinic facies, characterized by black shales, and commonly chert; (3) a Flysch facies, with rhythmically bedded dark shales and graywacke; and (4) a Molasse facies, in which the sandstones become coarser, cleaner, and show cross-bedding and other current structures. The succession of facies is tectonically controlled, reflecting: (1) slow, shallow marine sedimentation on a flooded craton or its margin; (2) rapid basining and starved sedimentation; then basin filling, first by (3) turbidity flows and other gravity processes taking place in deep water, and then (4) paralic sedimentation as the basin becomes filled.

Two such cycles are recognized in the Paleozoic of the Middle Appalachians: the first from Early Cambrian to about Middle Silurian, and the second from Late Silurian through the Carboniferous (Pettijohn, 1957, Fig. 165). The Maryland Piedmont reveals another, still earlier cycle. Setters and Cockeysville Formations represent the orthoquartzite-carbonate facies, Wissahickon-Sykesville the Flysch-Wildflysch facies, and the current-bedded metasandstone-phyllite formations in the western Piedmont the Molasse facies. The basal part of the Wissahickon, derived chiefly from shale and thin siliceous (chert?) beds, may correspond to the black-shale facies, although there is no proof of euxinic conditions.

An important problem is the source of the clastic detritus that filled the Piedmont trough. Much of it appears to have come from the east. The Sykesville Formation provides one line of evidence. The Sykesville is an immense wedge of submarine slide material that must have slid from a steeply rising ridge or scarp (Fig. 29). There is little possibility that this ancient ridge lay west of the Sykesville because of the stratigraphic relations, but it could have been to the east, beneath the present Coastal Plain. Along the Potomac River the mass of Sykesville slide debris thins and lenses out to the west (Figs. 25, 26) and passes into graded turbidites and then into finer silty and argillaceous beds. This suggests strongly that sliding was to the west. Moreover, the Sykesville contained a large mixture of coarse sand as well as quartz pebbles and exotic blocks of basement rock. There is no possible source for this coarse debris west of the Sykesville belt, where for many miles the rocks are fine-grained metasandstones and phyllites. Therefore, the coarse debris must have come from the east. If the submarine sliding was directed westward from an elevated area on the east, the Wissahickon turbidity flows probably came from the same direction.

These deductions are supported by observations on Wissahickon slump structures. There is much variation, but in general they indicate westerly downslope movement. Observations on slumping directions are still too few to carry much weight, however.

Still another indication of an eastern source is the westward progression from Flysch to Molasse. In the Alps, Carpathians, and elsewhere the Flysch troughs developed adjacent to the newly rising cordillera, but as uplift continued and troughs became filled

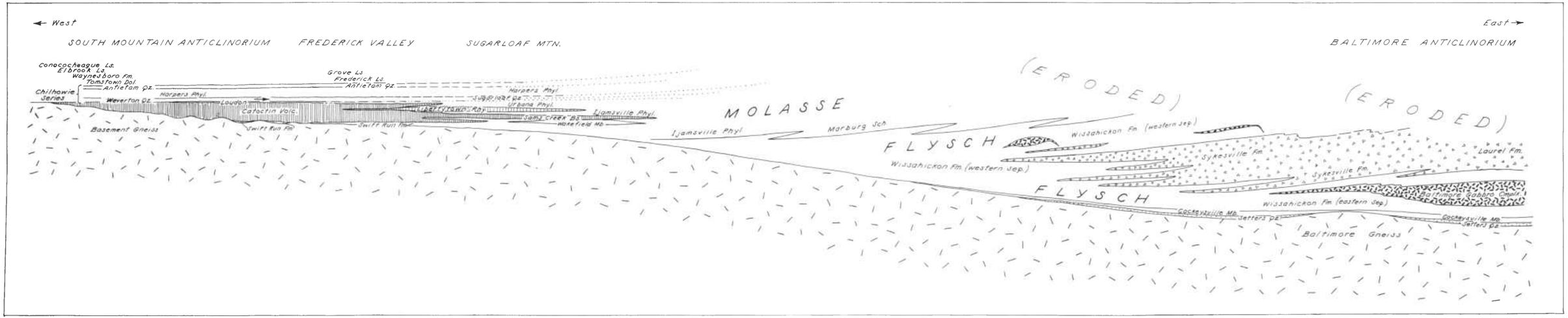


FIGURE 31. Diagram of inferred stratigraphic and facies relations in the Maryland Piedmont. The units are now folded and faulted. Basement rocks shown by hatched pattern; mafic and ultramafic intrusions by heavy hatching. The Weverton and Antietam Quartzites change eastward to rows of dots, representing transition to thinner units which finger out to the east. The arrow in the Weverton shows direction of current transport. The diagram is not drawn closely to scale.

P. 130 B

P. 130B

the site of paralic sedimentation shifted outward, toward a more marginal position. If this pattern holds for the Piedmont trough the cordillera or tectonic lands that supplied the sediment would have lain to the east.

Thick beds of clean orthoquartzite are interstratified with the immature molassic sediments in the western part of the Piedmont. Whitaker (1955) has shown that the Weverton orthoquartzite, which is approximately the same age, was washed in from the west. The western Piedmont paralic zone may therefore have received sediments from at least two directions: clean quartz sands from the craton on the west, interfingering with a flood of immature detritus from highlands rising to the east. This could explain the eastward thinning of the Weverton, between Catoctin and Sugarloaf Mountains, and its disappearance in the Piedmont beyond. A detailed study of current structures in the Sugarloaf Mountain area might provide a firm answer.

The present study does not support current opinion that the Piedmont metamorphic rocks are a eugeosynclinal facies of the Paleozoic sedimentary rocks that lie west of the Blue Ridge (Stose and Stose, 1951; King, 1959; Eardley, 1962). (1) The Piedmont sediments are older, probably Late Precambrian. (See chapter on Radiogenic Dates and Geologic Age Problems.) (2) They show the same sequence of sedimentary facies as the Paleozoic strata to the west. The Piedmont rocks therefore appear to be the product of an earlier geosynclinal cycle. This cycle differed from the later (Paleozoic) ones by having less pre-orogenic and more Flysch (including Wildflysch) sediments, and in being followed by regional metamorphism and plutonic intrusion.

Another conclusion is that the Piedmont trough in Maryland was filled with chiefly nonvolcanic detritus, much of which was derived from metamorphic rocks. This argues against the concept that the Piedmont was bordered by a volcanic island arc (Kay, 1951; King, 1959). A landmass of crystalline rock, east of the present Piedmont, is more compatible with the present data. Perhaps this was the landmass of Appalachia, a concept made popular many years ago by Schuchert (1910)

PLUTONIC ROCKS

INTRODUCTION

Widespread plutonic activity along the eastern part of the Maryland Piedmont accompanied and outlasted the Early Paleozoic deformation and regional metamorphism. Two series of plutonic rocks can be recognized: a gabbroic and a granitic series. In addition, in the Baltimore area there are several swarms of huge pegmatite dikes and sills which appear to have a closer genetic connection to the gneiss domes than to either series of igneous intrusions.

Members of the first, or mafic series include the Baltimore Gabbro Complex, the Georgetown Mafic Complex, and the assemblage of smaller serpentinite, soapstone, and metagabbro bodies that occur in the Wissahickon Schist in central Montgomery and Howard Counties. Also, the Relay Quartz Diorite and associated albite granite are grouped with this series, rather than with the other granitic rocks, because of their close genetic relationship to the Baltimore Gabbro.

The second, or granitic series, includes the Norbeck Quartz Diorite, Georgetown Quartz Diorite, Kensington Quartz Diorite, Ellicott City Granodiorite, Woodstock Quartz Monzonite, Guilford Quartz Monzonite, and the Bear Island Granite. The quartz diorites are chiefly concordant, synkinematic intrusions, and the granodiorite and quartz monzonites are late kinematic.

BALTIMORE GABBRO COMPLEX

General Statement

The Baltimore Gabbro Complex is a large mass of mafic and ultramafic rock that crops out in the western part of Baltimore city and adjacent parts of Baltimore County and extends southward into Howard County. Similar rocks in the northeastern part of the city are part of another belt that extends to the Susquehanna River and beyond into Pennsylvania (Fig. 32). They all appear to be part of one great intrusive mass.

The Baltimore Gabbro and its ultramafic rocks are well known from the early work of George H. Williams (1884; 1886; 1890). His study of the gabbro (1886) made it a classic example of uralitization and metamorphism of mafic igneous rock. This remains the best petrographic treatment of the complex. Cohen (1937) has given an accurate account of the structure of the complex and a detailed structural map. Herz (1950) mapped the Baltimore-Patapsco aqueduct tunnel, which runs completely through the complex, and has given a detailed cross section and logs; this is an exceptionally valuable record, since the mass is poorly exposed at the surface. Herz (1951) has also written on the petrology of the complex and given detailed mineralogical data. Other studies of the gabbro includes the work of Leonard (1901), Knopf (1921), Insley (1928), and Knopf and Jonas (1929).

Views on the origin of the complex have varied widely; some place most emphasis on its igneous derivation (Herz, 1951), and others on its later metamorphism (Williams, 1886; Cohen, 1937). It has also been suggested that many of the internal structures formed by semisolid flowage during intrusion of the mass (Cloos, 1937, p. 236; T.P. Thayer, personal communication). There is evidence to support each of these views, which are not necessarily incompatible. My interpretation is that the mass was intruded as magma at the earliest stage of orogeny; was caught up in compressional folding, gneiss doming, and regional metamorphism during the later stages of its crystallization; and that deformation and metamorphism continued after solidification.

Form of the Complex

The gabbro complex occurs in two major northeast-trending belts: one extends from the western part of Baltimore city southwestward to Laurel, where it passes beneath the Coastal Plain; the other emerges from beneath the Coastal Plain in east Baltimore and extends northeastward past Belair to the Susquehanna River. The State Line serpentine district lies along the continuation of this belt in Cecil County, Maryland, and Pennsylvania. A third belt, shown as "serpentine" on the geologic maps of Baltimore and Howard Counties, extends from Soldiers Delight southwestward almost to the Montgomery County line (Fig. 32). This belt is much thinner and appears to contain a higher proportion of ultramafic rocks. For convenience these three belts are informally called (from west to east) the Soldiers Delight, Laurel, and Belair belts.

Gabbro and norite of the Belair belt have ultramafic rocks along their northwest border and quartz diorite (shown as quartz gabbro on the Geologic Map of Harford County (Mathews and Johannsen, 1904) on the northeast. The belt is broadly concordant with the structure of the country rock. Knopf (1921, p. 89) believed that it was the edge of a huge sill-like mass, in which

"... the serpentinized peridotite-pyroxenite series [along the northwest border] represents the floor phase of the intrusion, derived from the gabbro magma by gravity differentiation of the ultramafic constituents in a manner analogous to the formation of the celebrated sheet at Sudbury, Ontario."

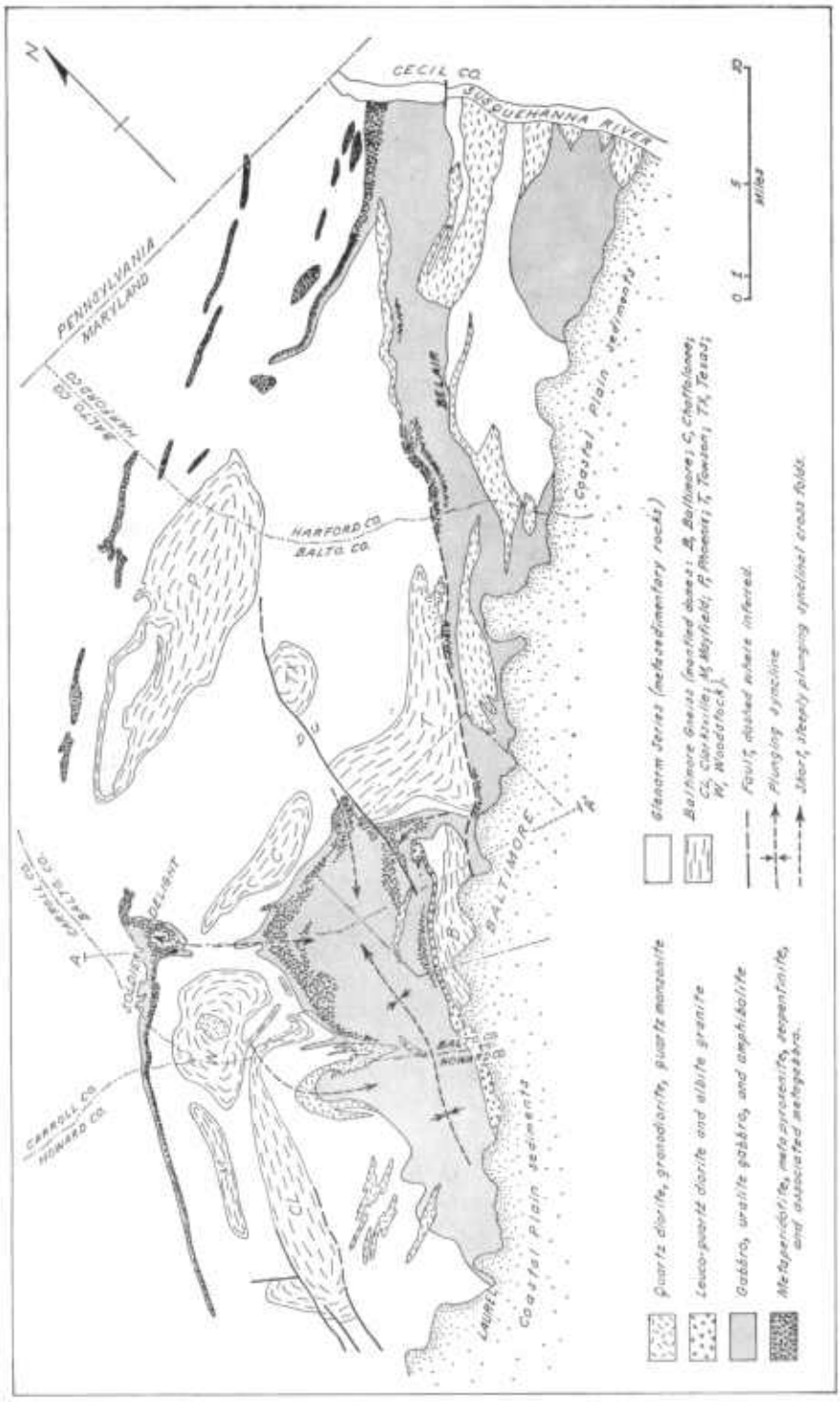


FIGURE 32. Generalized geologic map of mafic and ultramafic rocks in the Maryland Piedmont between Laurel and the Susquehanna River. Compiled from the Maryland Geological Survey maps of Howard, Baltimore, Carroll, and Harford Counties.

The Laurel belt, emerging from beneath the Coastal Plain at Laurel, extends northeastward for 20 miles and terminates abruptly in the structural depression between the gneiss domes at Baltimore (Fig. 32). This mass also is chiefly concordant with the older wall rocks. In Baltimore County Cohen (1937, p. 227) found that "... The boundaries of the gabbro against the country rock are remarkably conformable and uniform." Where it continues into Howard County the mass also appears to be broadly concordant, although the actual contacts are rarely exposed.

Ultramafic rocks are chiefly concentrated toward the margins of the Laurel mafic belt in Baltimore County (Fig. 32). This was first pointed out by Knopf and Jonas (1929, p. 116):

> "There are numerous exposures of serpentine within the large gabbro area that lies north and northwest of Baltimore. They are localized in general on the periphery of the gabbro intrusion."

Herz (1951, p. 1013) also notes that "... All exposed sides of the intrusion reveal a concentration of ultramafic rocks."

The marginal distribution of ultramafic rocks is not so uniform as might be imagined, however. The ultramafic zone is irregular and discontinuous, and many of the bodies are isolated. Moreover, smaller amounts of ultramafic rocks also occur toward the interior of the complex, leading Cohen (1937, p. 221–222) to conclude that:

> "The distribution of the types is unsystematic. In particular, the marginal distribution of ultramafic rocks, typical of undeformed gabbro lopoliths, is lacking.... When closely studied it seems that the ultrabasics occur within a mile of every point within the complex."

This is misleading. Small amounts of ultramafic rocks occur throughout the complex in the Baltimore area, but the large masses are unmistakably concentrated toward the margins. This is clearly shown on Herz's (1950) detailed cross sections and logs, taken through the complex along the Baltimore-Patapsco aqueduct tunnel. My own mapping finds that the ultramafic rocks are more uniformly concentrated along the north and northwest sides of the complex than is shown on the Baltimore County Geologic Map. Figure 32, based on all information to date, shows the generalized distribution of the large areas of ultramafic rocks; their marginal distribution is evident.

The general form of the gabbro complex in the Laurel belt is that of a trough or syncline, closed abruptly at the northeast end. This is shown by the structure of the surrounding gneiss domes and Glenarm Series, with which the gabbro is generally concordant, and by the internal structures of the gabbro itself (Cohen, 1937, Pl. XLI). According to Cohen (1937, p. 235):

> "Primary structures indicate that the gabbro has been a conformable intrusion into a synclinal area of the Glenarm Series. It may be called a synclinal pluton. Its shape may be that of a sill or basin-like body, narrowing downward from the south, east, and north sides—wherever the Baltimore gneiss domes are nearest."

Herz (1951, p. 1015) called the intrusion a funnel, which cannot be correct. Both the outcrop pattern (Fig. 32) and the internal structures show that the structure is open on the southwest and that it extends in that direction for at least 20 miles and continues for an unknown distance beyond beneath the Coastal Plain.

The Soldiers Delight belt is a narrow strip of mafic and ultramafic rocks that lies west of the line of mantled gneiss domes, isolated from the rest of the complex (Fig. 32). Serpentinite is widespread in the Soldiers Delight area proper, but throughout the

rest of the belt gabbroic rocks are also important. Most of the gabbro, however, has been metamorphosed to fine-grained greenschists, composed chiefly of hornblende, chlorite, clinozoisite (or epidote), and quartz. These rocks are well exposed at Liberty Dam and along the shores of the Patapsco Reservoir, as well as near Glenelg and elsewhere along the belt.

All three mafic belts can be interpreted as parts of a single large semiconcordant sheet with an ultramafic floor, which has been strongly folded and deformed during regional compression and gneiss doming (Fig. 33). In the middle (Laurel) belt the sheet was caught between the horseshoe-shaped line of gneiss domes which rose beneath it on three sides, pinching it into a tight syncline overturned to the southeast. The ultramafic lower part of the sheet is exposed around these three sides. The syncline is open on the southwest but terminates abruptly on the northeast where the sheet was bent upward by the rising Chattolanee dome.

To the west the sheet was arched over the Clarksville-Woodstock line of domes. The Soldiers Delight belt is the greatly thinned edge of this sheet, lying along the western limb of the arch. Rocks of the Soldiers Delight and Laurel belts nearly join across the lowest part of this arch—the structural saddle between the Woodstock and Chattolanee domes (Fig. 32). This is best seen in Herz's (1950, Baltimore-Patapsco Aqueduct sections) cross sections, which show that the mafic rocks are concordant with the underlying Wissahickon strata, which arch gently across this saddle. If the level of erosion had stopped a few hundred feet higher, the mafic rocks of the two belts would connect across this depression.

On the east the gabbro sheet was arched over the Baltimore City and Towson gneiss domes. The Belair mafic belt is the steeply dipping edge of the sheet where it comes down on the far side of these domes. Its top is to the east. This accords with Knopf's (1921, p. 89) explanation of this strip of gabbro as a sill-like mass, with the ultramafic rocks along its northwest side representing the gravity-differentiated floor. Structure of the Belair belt may be complicated by a fault along the northwest side. There is indirect evidence for such a fault where the gabbro comes in contact with the southeast edge of the Towson dome, and D. Southwick (personal communication) finds other evidence for faulting along the northwest edge of the belt in Harford County.

The interpretation of the gabbro complex as a single large folded sheet is supported by the following facts: (1) The generally concordant contacts of the mass suggest a sill-like intrusion. (2) The synclinal middle belt nearly joins the belts on either side, across structural saddles in the lines of gneiss domes. (3) The ultramafic rocks are concentrated along those sides of the intrusions that would correspond to the floor of a single great folded sheet.

For other evidence we turn to the internal structure of the complex.

Internal Structure

Cohen (1937) showed that in the complex some of the structures are igneous features and others are metamorphic. Some further details, chiefly from observations in the Baltimore area, may be added.

The mafic and ultramafic rocks locally show layering that resembles the gravitational layering in undeformed stratiform complexes. Rhythmic layering is one variety; individual layers range from an inch to several feet thick. There is also a finer, less sharply defined layering, similar to the "inch-scale layering" described by Hess (1960, p. 51, 133, Pl. 8; *also see* Rossman, 1963, Fig. 5). In addition, there is evidence of layering

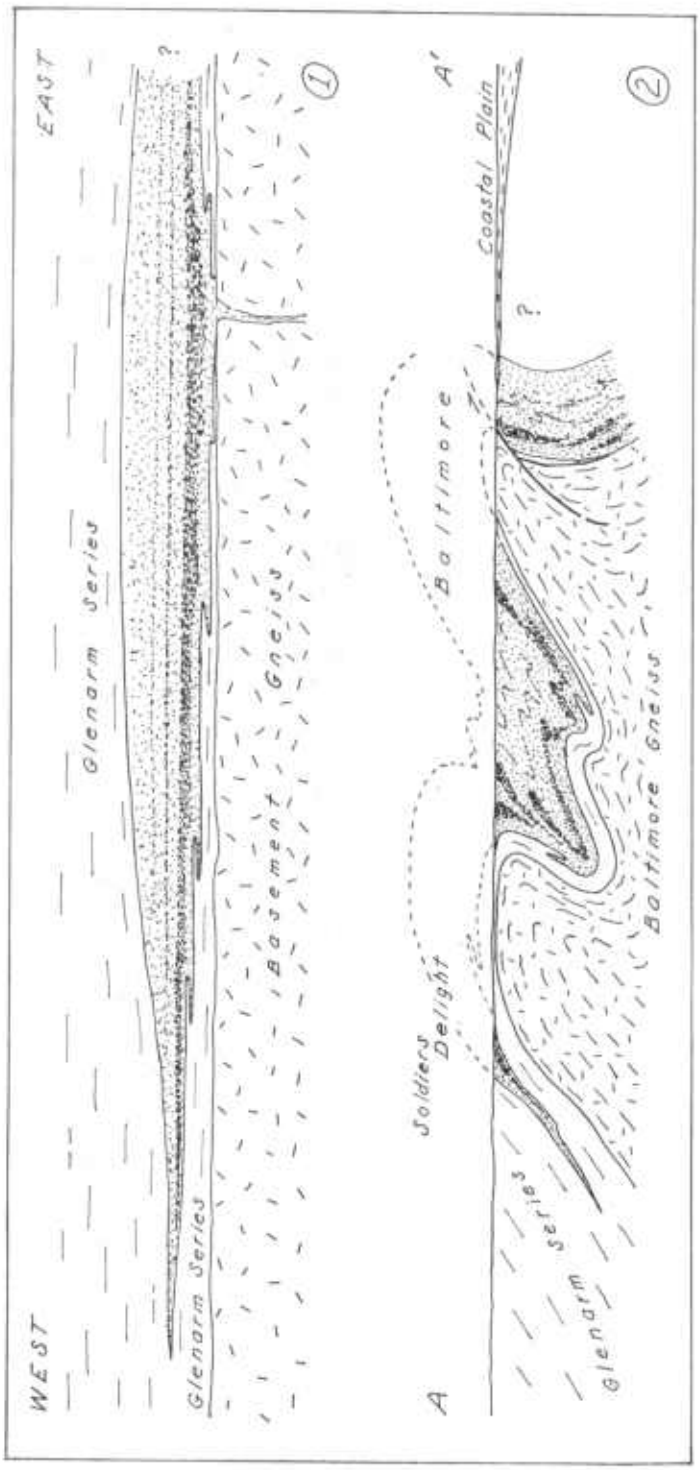


FIGURE 33. Structural interpretation of the Baltimore Gabbro Complex. 1. The gabbro complex after emplacement, but before regional deformation; the mass was not yet completely solidified. 2. The gabbro complex after folding, metamorphism, and erosion to the present level. Section taken at A-A' in Figure 32. The light stippling is gabbro (including amphibolite), and the heavy pattern is ultramafic rock. Relations greatly generalized.

on a scale of tens or hundreds of feet, in which massive rocks alternate with thinly layered rocks, or ultramafic rocks with gabbro. Coarse layering of this type could be seen in fresh road cuts during construction of the northwest quarter of the Baltimore Beltway, and they are also recorded on Herz's (1950) cross sections and logs of the Baltimore-Patapsco aqueduct tunnel.

Rhythmic layering is best displayed in the ultramafic rocks. Most commonly peridotite alternates with pyroxenite, but layers of dunite or mafic gabbro may also be present. The layering is uniform where there has not been subsequent deformation, and individual layers can be traced across the outcrop or quarry face with little change in thickness. Contacts between layers are commonly sharp, but some are gradational. The ultramafic layering is well shown in the large quarry south of Hollofield (Pl. 28, fig. 2). A measured section through the layered sequence in this quarry is given in Table 32. Other places in the Baltimore area where uniform ultramafic layering occurs is at Gwynns Falls north of Forest Park Avenue (*see* Herz, 1951, Pl. 5, fig. 2); in the area around the intersection of Belmont and Dogwood Roads; and in the new road cuts along the Baltimore Beltway between Liberty Road and Security Boulevard. Other good ultramafic layering near Gwynns Falls Parkway and the Western Maryland Railroad (Cohen, 1937, Pl. XXXIX, fig. 1) has been covered by recent construction.

Inch-scale layering is well shown in peridotite at several places. It consists of alternating more olivinic and more pyroxenic layers, mostly $\frac{1}{8}$ to $\frac{1}{2}$ inch thick. The individual layers are not sharply bounded but are rhythmically repeated dozens or even hundreds of times. The best examples come from huge blocks in new excavations just west of Belmont. Here the rock preserves good poikilitic igneous texture, so it is reasonably certain that the layering is magmatic, not tectonic.

The gabbroic rocks also show rhythmic layering (Pl. 28, fig. 1). The dark layers have more pyroxene or uralite, and the light layers more plagioclase. In the fresh pyroxene gabbro there may also be a grain-size difference between alternating layers; the light layers tend to have coarser tabular plagioclase. A prominent zone of layered gabbro runs along Gwynns Falls between Windsor Mill Road and Dickeysville and is well exposed in the Hillsdale quarry. The layers here range from less than an inch to more than 10 feet thick, but many are 2–12 inches thick. Some layers may be traced for tens of feet with little change in thickness, although others pinch out rapidly. A few of the layers are graded from pyroxene-rich at the base to plagioclase-rich at the top, and appear to have formed by differential crystal settling. Similar rhythmic gabbros may also be found in the Woodlawn area, south of Dogwood Road. Here also some layers show grading (Pl. 28, fig. 1).

Along the western side of the complex the layered ultramafic rocks in the Belmont area appear to grade upward (eastward) into layered gabbros through a transitional zone in which gabbroic and ultramafic rocks are interlayered. Rocks of this zone were exposed continuously for several hundred feet in a new road cut of the Baltimore Beltway near Windsor Mill Road. Here the gabbro is rhythmically bedded in layers 6 inches to 20 feet thick (average 10 feet); it also shows a fainter but uniform inch-scale layering. Ultramafic layers, chiefly metapyroxenite but including some metaperidotite, appear every 20 to 50 feet. These range from 1 to 5 feet thick and average 3 feet. Individual layers could be followed for 40 feet, with little change in thickness.

Primary foliation (also called igneous lamination, Hess, 1960, p. 51) is found in some of the plagioclase-rich gabbros. The tabular plagioclase crystals lie with their long dimension in the plane of the layering, producing a planar structure. Viewed perpendicu-

TABLE 32

MEASURED SECTION OF LAYERED ULTRAMAFIC ROCKS

Baltimore Gabbro Complex

Location: Quarry, Patapsco River at Old Frederick Road, half a mile south of Hollofield, Baltimore County.

Geology: The quarry exposes a small part of a thick section of metamorphosed ultramafic rocks, near the base of the Baltimore Gabbro Complex. The upper part of the quarry is in massive metapyroxenite, the lower part in thinly interlayered ultramafic and gabbroic rocks. The measured section is 69.4 feet thick and consists of metapyroxenite, 69.1%, metagabbro and mafic gabbro, 15.7%, metaperidotite and metadunite, 14.1%, chlorite-garnet rock 0.7%, and meta-anorthosite 0.4%.

Measured Section	
16 feet	*Metapyroxenite* (pale-green amphibole + colorless chlorite; minor magnetite). Massive, with no well-developed foliation or lineation. Cut by joints and small shears. Top not exposed.
2 feet	*Metapyroxenite* (anthophyllite + tremolite; minor relict clinopyroxene and olivine, Mg-chlorite, opaques). Massive, originally coarse-grained rock.
2 feet	Thinly interlayered *metaperidotite* (mesh serpentine, relict olivine, chromite; lesser tremolite, Mg-chlorite, magnetite) and *metapyroxenite* (chiefly anthophyllite, tremolite). The layers range from ½ to 6 inches thick, but most are 1 to 2 inches. Contacts between layers are sharp.
4 feet	*Mafic metagabbro* (pale-green hornblende + bytownite). Massive, fine-grained, granoblastic. Rests in sharp planar contact on underlying 4-inch peridotite layer. There is a faint layering within the gabbro, and near the middle a ⅜-inch layer of coarse metapyroxenite.
4 feet	Thinly interlayered *metaperidotite* (serpentine + olivine; lesser tremolite, Mg-chlorite, chromite, magnetite, carbonate) and *metapyroxenite* (chiefly tremolite, anthophyllite, minor Mg-chlorite). The peridotite was olivine-rich. The metapyroxenite layers are thicker (1 to 10 inches) than most of the metaperidotites (¼ to 4 inches). Contacts between layers generallly sharp, but commonly the layers are thickened, thinned, or pinched off.
1½ feet	*Metapyroxenite* (anthophyllite + tremolite; minor relict clinopyroxene and olivine, Mg-chlorite, opaques). Massive. Uniform in thickness across the quarry face (Pl. 28, fig. 2).
1 foot	Thinly interlayered *metaperidotite* (chiefly serpentine, relict olivine, tremolite) and *metapyroxenite* (chiefly pale-green amphibole). Individual layers ¼ to 2 inches thick.
10 inches	*Metapyroxenite* (Chiefly pale-green amphibole). Shearing and possibly some hydrothermal alteration. Schistosity oblique to layering.
10 inches	*Metaperidotite* (serpentine, relict olivine, tremolite; minor chromite and Mg-chlorite) Massive and uniform.
4 inches	*Metapyroxenite* (?) (pale-green amphibole; minor Mg-chlorite). Massive; fine-grained. Possibly hydrothermally altered.
8 inches	*Metaperidotite* (serpentine, relict olivine, tremolite; minor chromite and Mg-chlorite). Massive.
8 inches	*Metapyroxenite*, possibly with some shearing and hydrothermal alteration. Schistosity oblique to layering.
10 inches	*Metaperidotite* (chiefly olivine; lesser serpentine tremolite, chromite, and Mg-chlorite). Massive and uniform. Contact with underlying metapyroxenite is sharp but irregular.
3 feet	*Metapyroxenite.* Massive. Cut by small fractures, with alteration to very pale green amphibole and chlorite.
0–3 inches	*Amphibole-chlorite rock.* Very fine grained and massive, but patchy and heterogeneous-appearing. Pinches and swells; irregular contacts with enclosing layers.
5 feet	*Metapyroxenite* (tremolite + anthophyllite; minor Mg-chlorite and opaques). Massive, originally coarse-grained rock. The lower 2 inches is richer in chlorite and strongly schistose; schistosity transects layering. Small shears with chlorite developed along them, cut the massive rock.

TABLE 32—*Continued*

Measured Section	
2 inches	*Chlorite-garnet rock.* Coarse massive chloritic layer, with large clots of coarse reddish garnet. Contact irregular with underlying metagabbro.
1 foot	*Mafic metagabbro* (pale-green hornblende + bytownite). In sharp planar contact with underlying rock. Weak foliation, dipping about 20–30° more steeply than the layering.
8 inches	*Metapyroxenite* (pale-green amphibole; with lesser Mg-chlorite). Strongly schistose; schistosity parallels layering in the interior; but near the base it refracts steeply toward the underlying metagabbro. Thin discontinuous zones of coarse chlorite-garnet rock developed along the upper and lower contacts. This entire zone appears to be one of shearing and strong hydrothermal alteration.
2 feet	*Mafic metagabbro* (70% pale-green hornblende; 30% bytownite). Lowermost 8 inches is feldspathic metapyroxenite, with some thin layering; grades upward into the more uniform mafic metagabbro. Contains small metapyroxenite inclusions. Weak foliation transecting the primary layering at a low angle.
1½ inches	*Metapyroxenite* (tremolite-actinolite; minor Mg-chlorite, magnetite). Very fine grained and massive. Grades abruptly downward into the underlying coarse metapyroxenite.
2½ inches	*Metapyroxenite* (tremolite + anthophyllite; minor Mg-chlorite, magnetite, carbonate). Massive, and coarse grained.
3 inches	*Chlorite-garnet rock.* Coarse chlorite, with large clots of coarse pinkish-red garnet. No foliation. This rock is in sharp but very irregular embayed contact with the underlying anorthosite, and locally cuts down discordantly across it. The rock appears to be a product of hydrothermal alteration along a shear zone parallel to primary layering.
1–3 inches	*Anorthosite* (bytownite; minor hornblende). Very fine grained; granoblastic. Rests on the underlying gabbro in sharp planar contact.
4 inches	*Mafic metagabbro* (85% hornblende, 15% bytownite). A thin feldspathic layer at the base. Well developed foliation transects layering at about 20°.
3½ feet	*Metagabbro* (pale-green hornblende + bytownite; minor Mg-chlorite). Weak layering: alternating mafic and felsic layers, 6 inches to 1 foot thick. Foliation transects primary layering at a 20° angle.
3 feet	*Metapyroxenite* (light-colored patches chiefly pale-green amphibole; darker patches chiefly blue-green hornblende + minor Mg-chlorite. Rocks fine-grained with patchy coloration. Strongly sheared locally, with chlorite abundant along the shear zones. No well-developed foliation or lineation. Grades downward into the coarser, more massive metapyroxenite beneath.
4½ feet	*Metapyroxenite.* Massive. Sharp planar contact with thin metadunite layer of the underlying zone.
14 inches	Thinly alternating layers and lenses of metapyroxenite (tremolite + anthophyllite; minor chlorite, serpentine, relict olivine) and *metadunite* (mesh serpentine with relict olivine; minor magnetite, chromite, carbonate, chlorite). Most layers range from ½ to 2 inches thick; one very continuous dunite layer ½ inch thick. Contacts between layers sharp and planar to ragged and irregular. Chiefly massive, but locally there is a weak schistosity at low angles to layering.
3 feet	*Metaperidotite* (serpentine + relict olivine + tremolite + anthophyllite; minor Mg-chlorite, chromite, magnetite). Outlines of original coarse pyroxene. No foliation or lineation. Contact with underlying layer not well exposed.
6 feet	*Metapyroxenite* (tremolite + anthophyllite; minor chlorite, opaque, carbonate). Composition and grain size uniform; no foliation or lineation. Base of layer not exposed.
	69.4 feet total thickness

lar to the layering the crystals are randomly oriented, like dominos dropped on the floor. Foliation of this kind is customarily ascribed to accumulation of settled crystals on the floor of the magma chamber (Wager and Deer, 1939).

The features described so far—rhythmic layering, inch-scale layering, graded layering, and tabular plagioclase foliation—are believed to have formed by crystal settling and related processes in the magma, rather than by deformation. This is suggested by their close resemblance to structures in stratiform complexes such as the Stillwater. Also, relict igneous textures are best preserved in these rocks. However, there are other structures that have formed by flowage and deformation, to which we now turn.

The gabbro layering is most commonly irregular, lenticular, or streaky. Individual layers pinch out, fray out, or fade out within a few feet or tens of feet. Some layers bifurcate, swell into bulbous masses, or buckle and swirl. Contacts between light and dark layers may be sharp at one point but become diffuse a few inches or feet away. In places the layers have been pulled apart and smeared out, forming schlieren gabbros.

Structures of this sort in the Hillsdale quarry in Baltimore have formed by distortion and disruption of the earlier rhythmic layers, evidently by plastic flowage. The evenly layered gabbroic and metapyroxenite layers at the Baltimore Beltway and Windsor Mill Road are thrown into large-scale folds. Here the layers become blurred and streaked out, as if they had been mushy when the folding occurred. At Ilchester, metapyroxenite layers in streaky metagabbro are boudinaged. Some of the boudins are widely separated and rotated, indicating considerable flowage. Features of this sort are commonplace throughout the gabbro complex and far exceed well-preserved examples of igneous layering.

Feldspathic net veins are common in the massive, unlayered gabbros, especially toward the center (top?) of the complex in the Baltimore area. They are irregular, anastomosing vein-like bodies that are more feldspathic than the gabbroic host (Pl. 29, fig. 1). They tend to follow the direction of banding or foliation of the host rock, but also commonly wander across it. The field relations make it clear that the veins are a feature of the gabbro itself, not some later granitic intrusive. Mineralogically they also correspond to the gabbro.

These net veins seem best explained as feldspathic liquid residuum that was squeezed from the gabbro before it had completely solidified, injecting the surrounding plastic rock as dikelike shoots along directions of weakness. At some places the injection was followed by plastic flowage of the entire mass, smearing out the veinlets into irregular streaks and schlieren. This process in all its stages is well displayed in an old abandoned quarry between Franklintown Road and Rognel Heights, in west Baltimore.

Amphibolite dikes appear to be associated with deformation and plastic flowage. The dikes are up to a foot wide and cut discordantly across the layering or foliation of the host. A typical example, east of Franklintown (Pl. 29, fig. 2), has narrow feldspathic margins and small feldspathic schlieren in the interior. This dike occurs in an area with abundant net veining and appears to have formed in a complimentary manner—by intrusion of a gabbroic crystal mush, in which the feldspathic residuum was squeezed from the mush and collected along the dike margins. Amphibolite dikes such as this are also found in the Hillsdale quarry near Dickeysville and in other places. Perhaps some of the much larger discordant masses of serpentinized peridotite and dunite (Williams, 1886, p. 17; Knopf and Jonas, 1929, p. 118) were also intruded as crystal mushes or plastic solids during deformation of the complex.

These features—irregular layering, schlieren, boudinage, net veins, and amphibolite

dikes—seem best interpreted as phenomena formed during deformation and flowage of the gabbro complex, while much of it was still a crystal mush or a highly plastic solid. They postdate the igneous layering and were partly derived from it by distortion and flowage.

Other structures are still later, and were imposed on the gabbro after it had completely solidified. Chief among these is the discordant schistosity or foliation, recognized by Cohen (1937, p. 229–231). Herz (1951, p. 1001) later denied the existence of discordant foliation, but he was mistaken. It is well displayed throughout the complex, both in the ultramafic and gabbroic (amphibolite) rocks. The two quarries near Hollofield provide typical examples. In the larger quarry the gently dipping ultramafic layering is cut by weak foliation that dips about 20° more steeply. It is best developed in the metapyroxenite layers and is commonly refracted near their contacts. The metagabbro layers show the same cross-cutting foliation and also contain a strong down-dip lineation, due to alignment of hornblende. The smaller Hollofield quarry exposes metagabbro with steeply dipping layers. Here again the layers are cut by a still more steeply dipping planar and linear foliation. Good examples are also found at Belmont, Ilchester, Melvale, and many other places in the west Baltimore area.

It has been suggested that the foliation and lineation, both concordant and discordant, may have formed during plastic intrusion of the mass, rather than during later metamorphism. One reason is because the internal structures of the gabbro complex conform more closely to the contacts of the mass than to some of the country-rock structures outside (Editor's note in Cohen, 1937, p. 236). Also, some of the foliation is earlier than the net-vein complexes and mobilized amphibolite dikes—features inferred to have formed before the mass had completely solidified. On the other hand, some of the foliation and lineation extends continuously through the granitic and pegmatitic intrusions which sharply crosscut the gabbro complex. The granitic gneiss east of Ilchester, the granodiorite gneiss of Leakin Park, and the Relay quartz diorite and albite granite are all sizeable masses that intruded the gabbro after it had solidified, but all have been intensely deformed and metamorphosed with it. In addition, numerous smaller granitic dikes which crosscut the gabbro have been intensely granulated and recrystallized—some so thoroughly that they resemble finely crystalline quartzite. Examples are found at Thistle, Ilchester, Gwynns Falls along Franklintown Road, the Hillsdale quarry near Dickeysville, the Arundel quarry at Jones Falls near 40th Street bridge, at Greenspring Avenue west of Melvale, and in many other places in the Baltimore area. There can be no question of the fact that deformation and metamorphic recrystallization long outlasted the solidification of the gabbro.

Foliation and lineation that formed during the protoclastic stage seem to grade into that which formed after the granite was solidified and cut by the granitic intrusions. It is doubtful if the early and late foliations could be mapped separately or even distinguished from one another in thin section. These structures apparently developed continuously in the gabbro throughout a long act of deformation, which began when the mass was still partly a crystal mush, and continued long after it had solidified.

Contact Metamorphism

Little is known about contact metamorphism around the gabbro complex. Cohen (1937, p. 220) and Herz (1951, p. 1014) reported that garnet in the Wissahickon schist increased in frequency toward the gabbro contact, and they regarded this as a contact-metamorphic effect. My own work argues against this. Garnets in the Wissahickon in-

crease in size and frequency toward the gneiss dome area in general, where the gabbro occurs, but I could find no special concentration of garnets around the gabbro itself. Moreover, in thin section the garnet schists from near the gabbro had no textural or mineralogical features any different from those several miles away.

Detecting contact effects around the gabbro complex is hampered by two things: (1) The Wissahickon schist weathers more readily than the other rocks in the area, and few fresh exposures really close to the gabbro contact are available. No one has yet made an effort to study the schist systematically within a few hundred feet of this contact. (2) The Wissahickon Formation in this area has been regionally metamorphosed to the kyanite zone; this metamorphism reached greatest intensity during the rise of the mantled gneiss domes, after the gabbro was already emplaced. This is sure to have blurred any record of contact metamorphism, and very careful petrographic study of schist in the innermost contact zone might be necessary to recognize it. No such study has yet been made.

Petrography

Some rocks in the gabbro complex have well-preserved igneous minerals and textures, while others show varying degrees of alteration and recrystallization. Still others are thoroughly recrystallized and have a new metamorphic mineral assemblage. Throughout the part of the complex that lies in Howard and Baltimore Counties recrystallization has been extensive, and rocks with original textures and minerals well preserved are found only at a few localities. The original, least-altered rocks will be described first.

Websterite, lherzolite, and dunite were the most common ultramafic rocks, and hypersthene gabbro was the predominant mafic variety. Olivine gabbro, norite, and anorthositic gabbro are greatly subordinate. Figure 34 shows the range of original rock types that were important in the complex. An interesting feature is the abundance of clinopyroxene in most of these rocks. In this respect the Baltimore complex is unlike nonorogenic stratiform complexes such as the Bushveld and Stillwater, where harzburgite, bronzitite, and norite predominate.

The mineralogical details will be summarized only briefly. The most detailed work has been done by Herz (1951), and the reader is referred to his paper.

Plagioclase ranges from An_{86} to about An_{50} but most commonly is around An_{75-80} in the least-altered gabbros. The amphibolite plagioclases tend to be more slightly sodic (about An_{65-75}) but there is considerable overlap. Few plagioclases show zoning; however, some exceptions noted farther on. Olivine, which occurs chiefly in the ultramafic rocks, is Mg-rich. Herz (1951, p. 986) found that it ranges from Fo_{96} to Fo_{70} with 70 per cent of the determinations falling in the range Fo_{90-86}. It does not show zoning.

Orthopyroxene displays a much wider composition range, En_{99} to En_{50} (Herz, 1951, p. 987). In the ultramafic rocks it is chiefly about En_{90-83}, but in the gabbros it is richer in iron and is often about En_{75}. Herz found zoning to be common, with the rims invariably more iron rich than the core. I also noted zoning but found that unzoned hypersthenes were also common. Where zoning occurs the greatest composition change is generally restricted to a narrow rim. The orthopyroxenes commonly show exsolution, the exsolved clinopyroxene phase occurring as thin plates in some rocks and as clusters of tiny blebs in others. The large clinopyroxene crystals commonly show diallage parting. Their composition, according to Herz (1951, p. 991), ranges from $En_{56}Fs_{5}Wo_{39}$ to $En_{40}Fs_{21}Wo_{39}$. Herz notes zoning in some of the clinopyroxenes, especially the larger

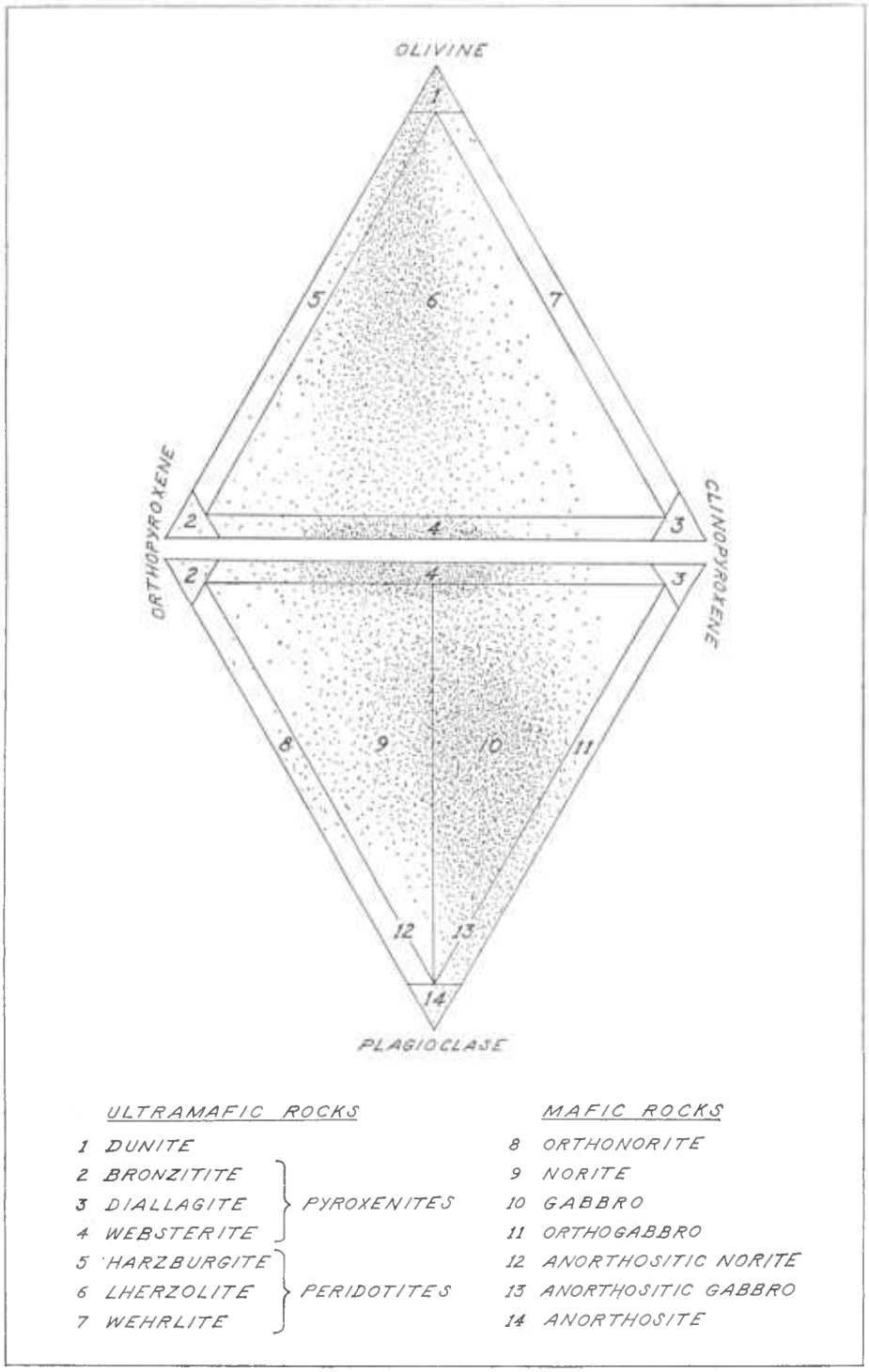


FIGURE 34. Classification of ultramafic and mafic rocks. Stippled areas show the chief rock types of the Baltimore Gabbro Complex. Dense stippling shows the most common varieties.

crystals. I found that in general strong zoning is not a characteristic feature of clinopyroxene in these rocks.

Iron-rich olivines and pyroxenes are not found in the Baltimore Gabbro Complex, suggesting that differentiation did not closely follow the Skaergaard trend. Hornblendes from the amphibolites (metagabbro) may have slightly higher Fe/Mg ratios than do the pyroxenes in corresponding gabbros. This is because the titaniferous magnetite or ilmenite in the original gabbro broke down to form sphene, releasing iron that went chiefly into the hornblende (Table 33). However even hornblende in the typical amphibolite (Table 34, col. 2) is not unusually iron rich.

The mafic and ultramafic rocks show relict cumulate textures, in which large well-formed settled crystals are enclosed by interstitial or poikilitic minerals that crystallized from the interprecipitate magma. The ultramafic rocks were more susceptible to alteration and plastic flowage, and they preserve these textures more rarely. Layered peridotite along Johnny Cake Road west of Belmont displays large bronzites poikilitically enclosing small rounded to euhedral olivines (Pl. 30, fig. 2). Small amounts of plagioclase, clinopyroxene, and their alteration products occupy the interstices between the bronzite oikocrysts. Olivine was the chief settled mineral, while bronzite, followed by clinopyroxene and plagioclase, grew from the enclosing melt.

Pyroxenites (websterite) with cumulate texture occur at Gwynns Falls near Forest Park Avenue in Baltimore and at the intersection of Windsor Mill Road with the Baltimore Beltway, west of the city. Orthopyroxene and clinopyroxene were both settled minerals. Clinopyroxene continued to crystallize from the interstitial melt, rimming the settled crystals so that their edges are partly molded around the peri-euhedral bronzites. Small amounts of plagioclase and brown hornblende also crystallized from the interprecipitate magma, filling the final interstices. The same relations hold for the mafic gabbros, which occur as thin layers in the ultramafic rocks.

The more feldspathic gabbros have plagioclase as a settled mineral. In anorthositic gabbro southeast of Ellicott City large tabular plagioclase with well-defined planar lamination was the only settled mineral; the pyroxene formed wedge-shaped interstitial fillings. More commonly, however, plagioclase, hypersthene, and clinopyroxene all form peri-euhedral settled crystals; however, the same three minerals have generally also grown from the interprecipitate magma, adding to the rims of the settled crystals so that their borders interlock (Pl. 30, fig. 1).

Wager *et al.* (1960) distinguished between cumulates in which the interprecipitate minerals crystallized from completely entrapped magma (orthocumulates) and those in which the composition of the interstitial melt was changed by diffusion with the overlying magma (adcumulates). Where there is more than one variety of settled mineral, each enlarged by interprecipitate minerals of the same species, the two cumulate types are distinguished by the presence or absence of zoning in the interprecipitate margins of the crystals. Zoning results from crystallization of completely entrapped magma but is impaired or not developed when there was free diffusion. Based on these criteria a high percentage of the Baltimore gabbros, in which plagioclase is unzoned or zoned in only the thinnest marginal rind, were evidently adcumulates. This may indicate that the settled crystals accumulated very slowly (Wager *et al.*, 1960, p. 77).

Brown hornblende is a common interprecipitate mineral in both the gabbros and pyroxenites. Williams (1886, p. 24) first recognized it as a pyrogenetic mineral, earlier and distinctly different from the green uralitic hornblendes. Herz (1951, p. 979, 998, 1015) disagreed, stating that the brown hornblende was also uralitic. The textural evi-

TABLE 33

MODAL ANALYSES OF METAGABBRO

Baltimore Gabbro Complex

	1	2	3	4	5	6	7	8	9	10	11	12	13	14
Plagioclase	25.7	26.2	29.4	29.6	31.6	33.1	40.7	42.3	46.2	48.6	50.1	56.8	57.9	67.0
Quartz	—	2.5	10.4	—	1.0	.2	—	1.0	2.6	—	—	.2	7.4	.3
Hornblende	72.8	68.6	57.8	70.4	60.5	62.6	50.4	50.9	46.2	50.6	45.1	42.9	29.6	27.3
Clinopyroxene	—	—	—	—	—	.2	2.8	—	—	—	—	—	—	.6
Epidote	—	—	—	—	tr	3.0	5.8	2.6	3.9	—	—	—	—	4.7
Clinozoisite	1.5	—	.8	—	—	—	—	—	—	—	2.8	—	1.3	—
Biotite	tr	—	—	—	—	—	—	—	.4	—	—	—	tr	—
Chlorite	—	—	.5	—	.3	—	—	.1	—	.8	.1	—	1.4	tr
Garnet	—	—	—	—	—	—	—	—	—	—	—	—	.6	—
Apatite	tr	—	—	—	.2	—	tr	.1	—	—	.3	.1	.2	—
Sphene	tr	1.2	.2	—	.7	.7	tr	2.8	.7	—	tr	—	.1	.1
Magnetite, Ilmenite	tr	1.5	—	—	5.7	.2	0.3	.2	tr	—	1.6	tr	1.4	tr
Carbonate	—	—	.9	—	—	—	—	—	—	—	—	tr	.1	—
Total	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0
Points	1602	1304	1693	1643	1291	2978	1619	1669	1276	1647	1210	1304	3180	1632

(1) Gneissic amphibolite (H54-4). Patapsco River at Thistle, Baltimore Co.
(2) Weakly gneissic amphibolite (B-61). Patapsco River, half a mile below Ilchester, Howard Co.
(3) Schistose amphibolite (Gb16-3). U. S. Route 1 at Deep Creek, half a mile north of Waterloo, Howard Co.
(4) Weakly gneissic amphibolite from layer in ultramafic sequence (H83-10). Quarry near Hollofield, Baltimore Co.
(5) Gneissic amphibolite (B-60). Near Thistle, Baltimore Co.
(6) Gneissic amphibolite (H54-1). Near Thistle, Baltimore Co.
(7) Massive uralite gabbro (H87-1). Md. Route 102, 1 mile south of Ellicott City, Howard Co.
(8) Massive amphibolite (Gb20-2). Leakin Park at Franklintown Road, Baltimore City.
(9) Gneissic amphibolite (H56-3). Patapsco River above Orange Grove, Howard Co.
(10) Weakly gneissic amphibolite from layer in ultramafic sequence (H83-6). Quarry near Hollofield, Baltimore Co.
(11) Massive uralitized norite (B-72). Hillsdale quarry near Wetheredsville Road, Baltimore City.
(12) Gneissic amphibolite (Gb15-1). Half a mile north of Westley Grove, Howard Co.
(13) Massive amphibolite (Gb16-2). U. S. Route 1 at Deep Creek, half a mile north of Waterloo, Howard Co.
(14) Massive uralite gabbro (H89-1). Md. Route 102, 1.5 miles south of Ellicott City, Howard Co.

TABLE 34

CHEMICAL ANALYSES OF MAFIC ROCKS AND MINERALS

	1	2	3	4	5	6	7
SiO_2	43.69	48.02	48.2	47.03	46.63	43.53	41.74
TiO_2	3.26	.66	.7	3.10	.40	none	nd
Al_2O_3	14.70	8.47	15.8	17.03	24.19	1.89	1.02
Fe_2O_3	5.59	3.45	2.6	4.00	1.28	.49	.87
FeO	11.63	9.19	6.4	11.23	6.34	4.21	.17
MnO	.22	.21		.23	.17	.04	.01
MgO	7.42	14.26	9.5	3.91	6.35	37.52	37.90
CaO	9.34	12.64	13.6	9.10	11.03	none	none
Na_2O	1.94	1.00	1.7	1.66	.51	none	nd
K_2O	.32	.27	.2	.54	.11	none	nd
H_2O^+	1.71	1.97	1.3	1.74	2.75	11.69	13.16
H_2O^-	.11	.10	—	.11	.05	.55	4.85
F	nd	.05	—	nd	nd	tr	none
CO_2	.19	nd	—	.20	.17	nd	nd
P_2O_5	tr	nd	—	.10	.02	nd	—
Cr_2O_3	nd	nd	—	nd	nd	.01	none
NiO	nd	nd	—	nd	nd	.20	.06
CoO	nd	nd	—	nd	nd	none	nd
ZrO_2	nd	nd	—	.02	0.00	nd	nd
Total	100.12	100.29	100.0	100.00	100.00	100.13	99.78

(1) Hypersthene-augite gabbro, partly uralitized (H55-5). Contains 6.6 per cent ilmenite-magnetite. Baltimore Gabbro Complex; Patapsco River at Ilchester, Baltimore County. Analyst, O. von Knorring.

(2) Green hornblende, from amphibolite (H54-1). Baltimore Gabbro Complex; Patapsco River near Thistle, Baltimore County. Analyst, O. von Knorring.

(3) Amphibolite (H54-1). Calculated from the modal analysis (table 33, col. 6), using the analyzed hornblende in column 2. Baltimore Gabbro Complex; Patapsco River near Thistle.

(4) Coarse grained amphibolite (M55). Large sill at Bear Island, Montgomery County. Analysts, Penniman and Browne.

(5) Gabbro (M121). Small intrusive body in Sykesville Formation, one-third mile northeast of Halpine Station, Montgomery County. Analysts, Penniman and Browne.

(6) Serpentine (antigorite + chrysotile) (F-19). Serpentinite mass in Baltimore Gabbro Complex, Bare Hills, Baltimore County. Analyst, K. J. Murata. (Faust and Fahey, 1962).

(7) Serpentine (lizardite and clinochrysotile) (F-3). Serpentinite body near Forest Hill, Harford County. Analyst, K. J. Murata. (Faust and Fahey, 1962).

dence clearly supports Williams' interpretation. The brown hornblende occurs as interstitial fillings between the peri-euhedral minerals or encloses them poikilitically. In each case, it is in sharp contact with the pyroxenes and has clearly grown *around* them. In contrast, the green uralitic hornblende eats into the pyroxenes along their margins, cleavages, and parting planes, gradually replacing them.

Magnetite also occurs as an interprecipitate phase, partly molded around the settled minerals. Occasionally it forms up to 7 per cent of the total rock (approximately 20 per cent of the interprecipitate)—far more than would have been present in the original entrapped magma. This appears to be another case in which material diffused in from the overlying magma, adding to a few magnetite crystals which had nucleated from the interstitial melt.

In summary, relict textures show that some of the layered ultramafic and gabbroic

rocks were igneous cumulates. This supports the structural evidence that the layering formed by crystal settling rather than by some tectonic process.

Mineral alterations and recrystallization have obscured original igneous textures and mineralogy throughout most of the complex, however. Dunite has gone to serpentinite, peridotite to serpentine-amphibole rocks, and pyroxenite to various amphibole or amphibole-chlorite rocks. Metasomatism has further changed some of them to rocks rich in talc and carbonate minerals. In the gabbros, pyroxene has gone to hornblende, forming uralite gabbro and ultimately amphibolite.

Serpentinization has affected all the olivine-bearing rocks, although olivine residuals persist. The initial alteration is to one or more varieties of fibrous serpentine, but antigorite appears in the more strongly deformed rocks. Serpentinite along U. S. Route 40 just east of the Patapsco River bridge consists of mesh-structure serpentine with sagenitic magnetite and was evidently derived from nearly pure dunite. More often the serpentinites contain bastite, tremolite, or other evidence of original pyroxene and were evidently derived from peridotite. The large serpentinite mass at Bare Hills is derived from peridotite.

As is usually the case, it is difficult to tell from textures whether the serpentinization was deuteric or occurred during metamorphism. An exception is the layered peridotites west of Belmont, which are partly serpentinized along fracture cleavage that cuts nearly at right angles across the layering. In this case the serpentinization was metamorphic.

The pyroxenites are changed chiefly to fibrous amphibole rocks. Tremolite is the most common amphibole, but it may be accompanied by anthophyllite and chlorite. These rocks are massive in outcrop, but under the microscope the amphiboles form a decussate mass of interpenetrating fibers and spears. Some of the pyroxenites have recrystallized to rock composed chiefly of a bright-green amphibole (very pale green under the microscope) that Williams (1886, p. 58) called smaragdite. Where there has been strong deformation the pyroxenites have gone to chlorite-amphibole schist.

Alteration of the pyroxene gabbro to rocks with green hornblende has been described in detail by Williams (1886, p. 27–32). The alteration commonly starts at the margins of the pyroxene and works inward along cracks and cleavages until gradually the entire crystal is replaced. In some cases each pyroxene is replaced by a single hornblende crystal (Pl. 31, fig. 1), but more commonly they alter to an interlocking aggregate of many smaller hornblendes (Pl. 31, fig. 2). In still other cases the pyroxenes break down to a maze of pale amphibole fibers, many tinted deeper green at the margins. Possibly this coloration reflects zoning of the original pyroxenes to more iron-rich rims.

Other mineralogical changes accompanying the uralitization are the release of quartz, which forms small grains enclosed in the hornblende, and the formation of clinozoisite (or epidote). The clinozoisite commonly forms reaction rims between hornblende and plagioclase, or very large xenoblasts that poikilitically enclose the same two minerals. Small amounts of sphene appear, released from ilmenite or titaniferous magnitite.

Some of the uralitized gabbros retain a relict igneous texture even where pyroxene is nearly all replaced (Pl. 31, fig. 2), but with further recrystallization it is gradually lost. Where deformation accompanied alteration, as was often the case, the original texture was wiped out by shearing and granulation (Pl. 32, fig. 1). The final product is gneissic or granoblastic amphibolite, composed chiefly of green hornblende and plagioclase (Table 33). At this stage the thin sections reveal no trace of the rock's former gabbroic origin. In the field, however, all gradations from incipiently uralitized gabbro

to schistose amphibolite can be traced, in some places even in the same outcrop. Good examples are found at Ilchester near the Patapsco River bridge and at the Hillsdale quarry in west Baltimore.

Herz (1951, p. 998–1001) thought that the amphibolites were primary igneous rocks, which he called bojite. He criticized Cohen's (1937) interpretation of them as crystalloblastic amphibolites, stating (p. 1001) that: "Crystalloblastic texture is never discernable; alleged metamorphic mineralogic changes are more easily attributed to normal late magmatic conditions." Nevertheless, crystalloblastic textures (Pl. 32, fig. 1) and all stages in their development are widely found. Herz (1951, p. 1010) also failed to recognize the gradual transition from gabbro to amphibolite, for he says:

> "The environmental change causing hornblende rather than pyroxene to crystallize [from the magma] was as dramatic and sudden as that causing plagioclase. Both came out in floods without any extended series of transition rocks."

The fact is, however, that amphibolite is intimately associated with relict gabbro throughout the complex, and transitions between them are seen both in the field and in thin section (Pls. 31, 32). Williams (1884, p. 267) recognized this many years ago:

> "In the great gabbro area west of Baltimore the massive diallage and hypersthene rocks occur everywhere interbedded in and passing by gradual transitions into more or less schistose amphibolites, which differ from them mineralogically only in the crystalline form of the bisilicate constituent. These amphibolites have throughout the whole area a nearly parallel strike and dip, and many other facts, which cannot here be enumerated, indicate that their schistose structure is, like slaty cleavage, the result of lateral pressure. That the amphibolites have resulted from the paramorphosis of the pyroxene in the gabbros is abundantly proven both by microscopic study and their relations in the field, and the fact is very significant that throughout the area, as a rule *the schistose structure is developed in proportion to the completeness of the paramorphosis.*"

This does not necessarily mean that the amphibolites developed long after the gabbro magma had crystallized. Structural evidence suggests that the magma crystallized under tectonic conditions, with protoclasis merging gradually into metamorphism. The conversion of gabbro to amphibolite may have begun during the protoclastic stage.

The feldspathic net veins, whose field relations were described earlier, appear to have formed where residual liquid was squeezed from the enclosing gabbro mush and concentrated along zones of weakness or low pressure. Thin sections from the net veins give supporting evidence, although metamorphic recrystallization of both the veins and host have done much to obscure it. The veins have fewer ferromagnesian minerals, as might be expected of residual melt. Moreover, some of the net-vein plagioclase shows relict progressive zoning. Relict plagioclase from a net vein in metagabbro at Leakin Park is zoned from about An_{55} to An_{40}, while the host rock plagioclase is about An_{65}. Relict plagioclase from a net vein in metagabbro along the Patapsco River below Ilchester is zoned from about An_{60} to An_{45}. Here recrystallized plagioclase in the host rock is also about An_{60}, but originally it was probably more calcic for it is accompanied by metamorphic clinozoisite. These observations suggest that the magma was differentiating toward diorite, but more detailed work is needed.

Origin

The petrography and structure of the complex reveal a complicated history, which began with magmatic crystallization and development of gravity layering, passed into

plastic flowage and protoclasis, and concluded with solid deformation and metamorphism.

The complex appears to be a huge folded and faulted sheet, formed initially by magmatic intrusion into flat-lying strata. This is suggested by: (1) its concordant contacts with the country rock; (2) its rhythmic layering parallel to the contacts; (3) its symmetrical distribution around the mantled gneiss domes, except where there was faulting; and (4) the concentration of ultramafic rocks near contacts that would correspond to the base of the sheet. Relict cumulate textures (Pl. 30) and rhythmic gravity layering (Pl. 28) support the idea that the ultramafics accumulated as a floor phase of the intrusion. Smaller amounts of rhythmically layered pyroxenite and peridotite that recur at higher levels in the complex can be attributed to recurrent periods of crystal settling, as new drafts of mafic magma came up from depth.

Regional compression and folding began before the mass had solidified and perhaps even before it had grown to full size. In the Baltimore area the mantled gneiss domes rose as diapirs directly beneath the gabbroic sheet, which was draped over the domes and partly pinched between them (Fig. 33). The evidence of mushy and plastic flowage shows that the mass was still solidifying and cooling when the tectonic movements were in progress. Magmatic flowage and protoclasis merged continuously into metamorphism as the deforming mass gradually solidified. Deformation and metamorphism evidently continued for a long time thereafter, as shown by the sequence of later granitic intrusives which have also been crushed and recrystallized.

Strong tectonic forces, after the mass had solidified but while it was still at high temperature and permeated by hydrous fluids, evidently squeezed mobile, partly serpentinized masses of peridotite and dunite from the places where they had formed by crystal settling. Some moved only far enough to distort or obliterate their settled textures and structures, but others were squeezed to a different part of the complex or into the country rock. Examples of serpentinized ultramafic rock intruding the gabbro are given by Williams (1886, p. 17) and Knopf and Jonas (1929, p. 118). The best example is at Bare Hills north of Baltimore, where a large mass of serpentinite cuts across the contact of the gabbro complex into the Wissahickon Formation.

Many of the serpentinized intrusive masses, following the direction of greatest weakness, were evidently squeezed *sideways* into the thin edges of the gabbro complex in response to pressure from beneath, as the gneiss domes began to punch upward into the base of the cooling igneous sheet. The two serpentinite tongues that cut across the northern edge of the complex at Pikesville and at Bare Hills (*see* Knopf and Jonas, 1925) may be cited as examples, and also the gently-dipping serpentinite tongue at Soldiers Delight. Other serpentinite masses were squeezed beyond the edge of the gabbro sheet and continued to move sideways along bedding planes in the Wissahickon, forming "grapefruit-seed" intrusions fringing the complex. The concordant serpentinite intrusions along the northwestern side of the Phoenix dome at Piney Run, Yeoho, and Blue Mount appear to be examples, for they lie along a continuation of the Soldiers Delight belt and occupy approximately the same stratigraphic horizon in the Wissahickon Formation. (Fig. 32). Following their emplacement, during the early stages of gneiss doming, the Phoenix dome continued to rise and tilted them steeply to the northwest.

Thayer (1960) has outlined the critical differences between alpine-type peridotite-gabbro complexes and the nonorogenic stratiform complexes. The Baltimore mass does not ideally fit either category but retains features of both. Relict structures and tex-

tures show that it began to develop under relatively quiescent conditions as a stratiform mass, but that orogenic forces intervened and strongly influenced its later development. Crystallization was completed during strong deformation; protoclasis and solid flowage distorted or wiped out many primary features and imparted characteristics of the alpine complexes. An important difference between the Baltimore mass and the alpine bodies described by Thayer is that the rocks at Baltimore solidified and were deformed in the same chamber they now occupy; only the mobile serpentinized masses were squeezed from their place of origin.

C. H. Smith (1958, p. 74–82) also lists differences between the undeformed, dominantly gabbroic Bushveld-type layered plutons and the strongly deformed ultramafic [alpine-type] plutons, but emphasizes that they are linked by a continuous series of plutons with intermediate characteristics. Both may have originated from a similar parent magma, but differences in cooling, emplacement history, and post-emplacement deformation have brought about variations in structure and in the relative abundance of gabbroic and ultramafic rocks now exposed to view. Smith's study of the Bay of Islands Complex in Newfoundland, an intermediate member of the series, shows that (1958, p. 82):

"... deformation, flowage, and faulting have developed ultrabasic plutons of batholithic dimensions from plutons that originally had a large component of gabbro. Although similar to ultrabasic plutons, these plutons have not formed from an ultrabasic magma, but by the mechanical dislocation of primary plutons composed partly of gabbro."

The Baltimore Gabbro Complex and its cortege of associated serpentinite masses shows many similarities to the Bay of Islands Complex.

The Baltimore complex offers many unsolved problems. One is the course of differentiation of the original gabbroic magma. Herz (1951, p. 1001–1012) has cited evidence for iron enrichment. Other evidence, such as net veins in the gabbro and relationships involving the Relay intrusive, (p. 160), suggest a trend toward diorite, quartz diorite, and albite granite. Perhaps the problem is best attacked in Harford and Cecil Counties, where the grade of regional metamorphism is lower and the gabbro sheet is seen in simple cross section.

Another problem is the source of the great amount of water that was necessary to convert gabbro to hornblendic rocks on such a large scale. If the gabbro was emplaced before metamorphism or during its early stages, as the present work suggests, then a possible source is the water expelled from the enclosing pelitic sediments (Wissahickon). If the gabbro was emplaced into already crystalline schist (Herz, 1951), then the only apparent source of water is its own magma. In this sense the problem is similar to the serpentinite controversy.

During this work I have benefited from field trips into the gabbro complex with A. C. Waters and T. P. Thayer. Both these men have been a rich source of ideas and stimulation.

SERPENTINITES AND SOAPSTONE

Numerous bodies of serpentinite, magnesian schist, and soapstone derived from intrusive ultramafic rocks are scattered throughout the broad belt of Wissahickon (western sequence) metasediments in central Montgomery and western Howard Counties (Pl. 7). They form thin sheets and thick lenticular bodies, mostly concordant with the structure of the country rock.

The larger masses occur in Montgomery County. Massive serpentinite forms a body more than 4 miles long and 1 mile wide near Hunting Hill. Another mass, $2\frac{1}{4}$ miles long and half a mile wide, lies along strike to the northeast between Gaithersburg and Redland. Farther west are several thinner sheetlike bodies. One crops out as a narrow belt more than 6 miles long, extending from the Potomac River near Pennyfield Lock to a point about 1 mile north of Quince Orchard. Another mass 2 miles long but only a few hundred feet wide occurs at Middlebrook, and a body of similar dimensions lies along strike with it in the area west of Etchison. A swarm of smaller lenses and slivers of magnesian schist and soapstone extend from Gaithersburg northward 6 miles to upper Great Seneca Creek. These bodies are best shown on the Geologic Map of Montgomery County (Cloos and Cooke, 1953; also Pearre and Heyl, 1960, Pl. 40).

In Howard County a line of small serpentinite lenses occur just west of the Sykesville Formation, near Carrs Mill and Hoods Mills. Along the east side of the Sykesville is the long narrow belt of serpentinite, amphibole-chlorite schist, and metagabbro that extends southward through Howard County from Soldiers Delight. It was described in connection with the Baltimore Gabbro Complex.

As a rule the large, thick bodies consist chiefly of massive serpentinite, whereas the thin lenses and sheets are largely schistose and metasomatically altered rock. Antigorite schist, chlorite schist, chlorite-talc schist, and talc-carbonate rock (soapstone) are the chief types. These rocks also form a rind around the larger serpentinite masses and occur along shear zones through their interior. It thus appears that serpentinite has survived mainly in the large massive bodies that withstood penetrative deformation and entry of metasomatizing solutions.

Serpentinization has been thorough, and evidence of the original rock has survived only in a few places. Some of the massive serpentinites in the Hunting Hill and Washington Grove bodies appear to have come from peridotite. Large bastites with relict pyroxene lamellar structure occur in a matrix of mesh serpentine and magnetite probably derived from olivine. A small schistose serpentinite body exposed along U. S. 40 west of Cooksvile has a massive core that was evidently derived from dunite. Mesh serpentine with sagenitic magnetite outline the former olivine grains; the rock consists wholly of this material. Scattered through many of the other bodies are massive to schistose rocks composed chiefly of tremolite-actinolite and chlorite; occasionally a little relict clinopyroxene remains. Their composition suggests that pyroxenite was the original rock. Brown (1958, p. 41) found that metapyroxenites of this kind were the most abundant ultramafic rock type in the Lynchburg district, Virginia.

Talc-rich rocks, commonly called steatite or soapstone, rim the serpentinites and occur as small separate bodies. These soapstone lenses are most abundant in Montgomery County between Gaithersburg and Woodfield. Another small mass of soapstone crops out along Cedar Lane, 2 miles east of Ashton. In the field these rocks are either schistose or massive, and they are distinguished by their "slippery feel". Many outcrops have a rough, pitted appearance, where carbonate porphyroblasts have weathered out. The observed mineral assemblages are:

(1) Talc + carbonate + magnetite
(2) Talc + carbonate + chlorite + magnetite
(3) Talc + chlorite + magnetite
(4) Talc + tremolite + magnetite.

Wiik (1953, p. 39–40) notes that tremolite and carbonate never occur together in soapstones. This observation holds true for Maryland.

Serpentinite seems to have been the starting point for the various alterations that led to soapstone. There is no evidence in this area that steatite developed directly from peridotite, without passing through a serpentinite stage. Several paths of alteration were noted. In some cases the serpentinite (commonly coarse sheaves of antigorite) were directly replaced by carbonate (magnesite) and talc. The carbonate appears first, randomly disseminated or working out from tiny cracks, and then is joined by talc. When the last serpentine is used up the rock consists chiefly of talc and magnesite, plus minor magnetite. A general equation for this reaction (Wiik, 1953, p. 37) is:

$$\underset{\textit{serpentine}}{2Mg_6Si_4O_{10}(OH)_8} + 6CO_2 \rightarrow \underset{\textit{talc}}{Mg_6Si_8O_{20}(OH)_4} + \underset{\text{magnesite}}{6MgCO_3} + 6H_2O$$

Thus the addition of CO_2 and removal of water is the chief chemical change. However, in many of the Montgomery County soapstones talc greatly predominates over magnesite, so that silica must have been added as well:

$$\underset{\textit{serpentine}}{Mg_6Si_4O_{10}(OH)_8} + 4SiO_2 \rightarrow \underset{\textit{talc}}{Mg_6Si_8O_{20}(OH)_4} + 2H_2O$$

Another common alteration in this area is the replacement of serpentine by a fine-grained, very pale-green chlorite. The end product is a rock composed solely of chlorite plus minor magnetite. The chlorite from a rock of this type, collected half a mile northeast of Quince Orchard, was identified as sheridanite (Fisher, 1963, p. 58). Its composition, based on optical and X-ray data, is approximately $Mg_{4.3}Fe_{0.4}Al_{2.6}Si_{2.7}O_{10}(OH)_8$. The alteration evidently necessitated the addition of aluminum, since no aluminous mineral appears to have initially been present.

In some cases this chlorite rock undergoes further partial replacement by talc, or by talc + carbonate, leading to talc-chlorite (-carbonate) soapstone. Here addition of silica (+ carbon dioxide) and release of water were the main chemical changes. In addition there are other alterations involving the formation of tremolite, which were not studied closely.

Opinion differs about the source of the hydrothermal solutions that alter ultramafic rocks to soapstone. The main possibilities are: (1) solutions given off from near-by acidic intrusions (Hess, 1933, p. 406–407); (2) residual solutions from the parent ultramafic magma (Benson, 1918, p. 727); or (3) solutions derived from the enclosing sediments or schists. The first possibility can be ruled out for the Montgomery County soapstones. The nearest granitic intrusions lie many miles to the east, too far away to have caused the alterations. The second possibility also seems unlikely. Figure 35 shows that alteration from peridotite to serpentinite to soapstone involves first a strong hydration (to about 11 per cent H_2O), and then a strong dehydration (to about 5 per cent H_2O) and addition of silica. Such a sequence of chemical changes is not likely to occur during simple cooling of a crystal mush with its own residual fluids. The field relations also argue against an internal source for the fluids: the alteration begins at the margins of the bodies and along shear zones indicating that the solutions were externally derived. The third possibility seems the only simple explanation. Metamorphism of the enclosing Wissahickon sediments would have liberated the water and other substances needed. The abundance of quartz pods and segregation veinlets in the mica schists attests to the chemical mobility of silica in these rocks during metamorphism.

A fact not yet mentioned is that metagabbro commonly occurs in the serpentinite and magnesian schist bodies in Montgomery County. During brief reconnaissance such

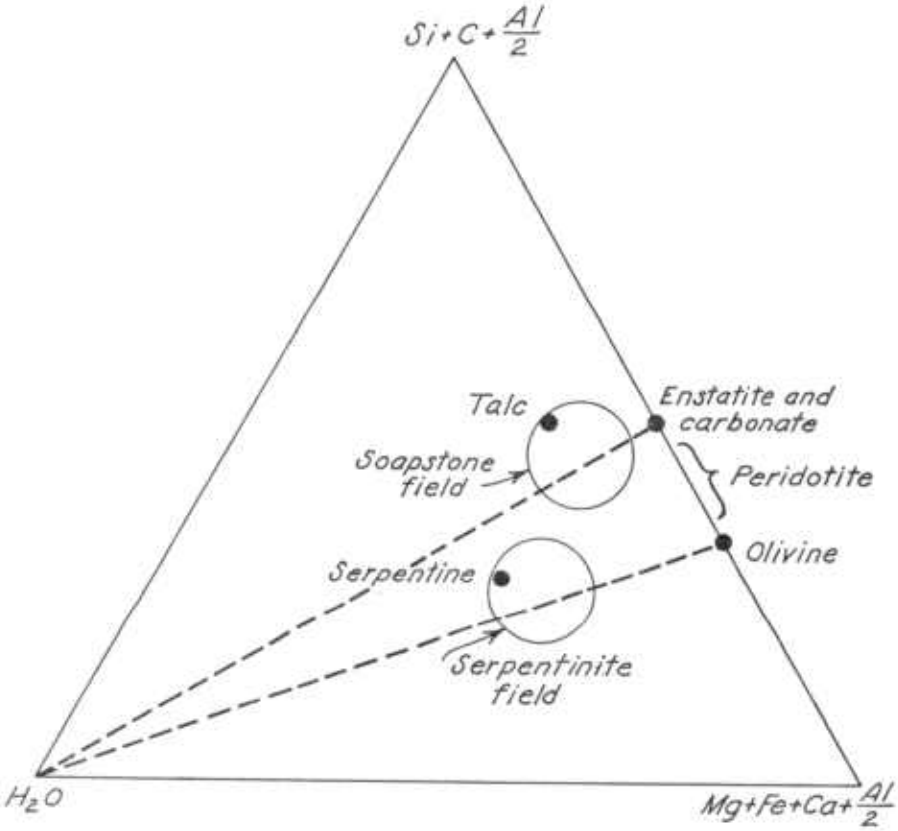


FIGURE 35. Chemical composition of serpentinite and soapstone. Modified from Wiik (1953, p. 29).

rock was found in the Hunting Hill (2 localities), Quince Orchard, Middlebrook, and Stewart Town serpentinite bodies, and detailed search would probably reveal many other occurrences. Where seen, it occurs as massive, coarse-grained metagabbro with small amounts of relict clinopyroxene. Pyroxene is chiefly replaced by tremolite-actinolite and chlorite, and plagioclase by granular clinozoisite and fine masses of nearly opaque material that may be hydrogrossular. Ilmenite is partially replaced by sphene. The metagabbro is easily recognized where it is coarse-grained and massive, but more often it is smeared out into actinolite-clinozoisite-chlorite schist, which is fine-grained and dark green, and resembles serpentinite. The field relations, best seen in the large quarry of the Rockville Crushed Stone Company south of Hunting Hill, show that the metagabbro is in large streaky masses enclosed in the serpentinite. These look like gabbroic schlieren, not later intrusive bodies.

The widespread occurrence of metagabbro and metapyroxenite with the serpentinite bodies suggests that the ultramafic material was derived from some kind of differentiated gabbroic pluton—perhaps similar to the Baltimore Gabbro Complex—rather than directly from the earth's mantle. The schlieren could be remnants of gabbroic layers in differentiated ultramafic rock, that were torn apart and smeared out during solid intrusion.

Few details are known about the emplacement of the bodies or the time at which serpentinization occurred. The intense shearing, deformation, and alteration to chloritic

schists and soapstone evidently occurred during regional metamorphism. Possibly some of the deformation and steatitization occurred as masses of serpentinite were squeezed as "grapefruit-seed" intrusions from some parent mass and migrated along zones of weakness in the Wissahickon strata.

GEORGETOWN COMPLEX

A poorly exposed complex of ultramafic, gabbroic, dioritic, and dark quartz dioritic rocks underlies a large part of western Washington, D. C., and adjacent parts of Montgomery County. They are concentrated along a zone 1–2 miles wide that extends for about 10 miles from Georgetown, in the District of Columbia, northwest to Beane in Montgomery County. They also crop out along the Potomac River, just west of Glen Echo. During the geologic mapping of Montgomery County Cloos broadly separated them into (1) tonalite; (2) meladiorite, gabbro, and amphibolite; and (3) undifferentiated basic rocks (Cloos and Cooke, 1953). For convenience the entire group is called the Georgetown Complex. This does not imply that all these rocks are differentiates from the same magma, or even that they are closely related in time.

These rocks were not studied during the present investigation. A general description is given by Keith (Darton and Keith, 1901), and more detailed work has been done by Fellows (1950) in Rock Creek Park and by Fisher (1963, p. 61–75) along the Potomac River. A new chemical analysis of gabbro from the complex is given in Table 34, col. 5.

AMPHIBOLITES AT BEAR ISLAND

Concordant amphibolite bodies appear locally in the Wissahickon Formation at various points in Montgomery and Howard Counties, but nowhere are they so well exposed as along the Potomac River at Bear Island. Here they form sheetlike masses up to 100 feet or more thick in the thinly stratified metagraywacke-mica schist sequence. They have been tightly folded into anticlines and synclines and metamorphosed with the enclosing sediments. The larger bodies are shown in detail on the plane-table map of Bear Island by Cloos (Cloos and Anderson, 1950, Fig. 1), and on smaller-scale maps by Fisher (1963, Pl. 2) and by Reed and Jolly (1963, Pl. 2).

Contacts between the amphibolite and enclosing metasediments are mostly concordant, but locally they crosscut at low angles. (Fisher, 1963, p. 51, 53) has also found small dikes of amphibolite. Thus, many of the amphibolites appear to be intrusive bodies rather than lava flows.

The amphibolites are composed chiefly of green hornblende and plagioclase, generally sodic andesine. The hornblende, which commonly occurs in large sieved porphyroblasts, predominates. Variable lesser amounts of quartz, opaque minerals, biotite, and epidote are generally also present. The minor accessories are apatite, sphene, allanite, and zircon. Chlorite, hematite, and calcite occur locally. Reed and Jolly (1963, p. H7) report finding relict pigeonite. A chemical analysis of coarse-grained amphibolite from a large body near the middle of Bear Island is given in Table 34, col. 4. The composition of this rock corresponds to diabase or gabbro, although the low Mg and high Ti are unusual features.

Some of the large amphibolite bodies have coarse-grained relict texture which suggests that the original rock was gabbro or coarse-grained diabase. In thin section clots of hornblende appear to have replaced large pyroxene crystals; rarely they contain small relics of pyroxene. Finely granular recrystallized plagioclase is interstitial to the hornblende, but in some places, according to Fisher (1963, p. 52–53):

"The plagioclase crystals in the matrix occur in elongated patches, each made up of small crystals in sub-parallel orientation; the patches are more or less randomly oriented, and closely resemble the disoriented plagioclase laths characteristic of metamorphosed diabase."

The massive amphibolite becomes finer-grained and progressively more schistose toward the margins. The foliation and lineation here are clearly tectonic (*see* section on Structural Geology), but the rock apparently was finer-grained toward the margins even before the schistosity was imposed. The marginal decrease in grain size is so consistent, even where schistosity is weak, that it can probably be interpreted as relict chilled borders inherited from the original intrusive.

Thus the structure and texture of the amphibolites, and the presence of relict pigeonite, indicate that they were originally diabase sills, emplaced in the sediments before folding and metamorphism.

A few of the amphibolites occur in thin sheets, 1–4 feet thick, that are strongly boudinaged. The boudins are separated up to 10 feet or more from one another and rotated to various orientations. Some of the longer boudins are strongly folded or intensely contorted. The boudinaged amphibolites invariably occur along chaotic zones in the metagraywacke sequence, which are inferred to have formed by mass flowage or sliding before the sediments were firmly consolidated (p. 90–93). The amphibolites appear to have been torn apart during this movement, rather than during the later metamorphism. This suggests that ophiolitic magma, from which these mafic sheets formed, was invading the sediments at shallow depth while deposition was still going on. Possibly some of this magma erupted onto the sea floor, forming mafic aquagene tuffs or palagonite. Some of the thin, dismembered amphibolites might well have been sheets of fragmental extrusive material, rather than sills.

RELAY QUARTZ DIORITE

The Baltimore Gabbro Complex is cut by leucocratic granitic rocks all along its southeastern margin, from northern Baltimore to Dorsey, in Howard County (Fig. 32). For most of the distance there appears to be one main body several hundred feet thick, emplaced along the contact between the gabbro and its wall rock. North of Druid Hill Lake in Baltimore it branches into dozens of smaller sills and dikes, forming an injection zone which can be traced northward for at least another 3 miles. The main body is best exposed along the Patapsco River and in railroad cuts near Relay, the locality for which it was named (Knopf and Jonas, 1929, p. 130). Elsewhere it is poorly exposed, and in Howard County from Elkridge southward it is easily confused with weathered outcrops of leucocratic Baltimore gneiss, which borders it on the southeast.

The mass looks like a sill from the map pattern, but it may be crosscutting at depth. Knopf and Jonas (1929, p. 130–131) report that the quartz diorite contact cuts across the gabbro near Relay. On the south side of the Patapsco River the contact is sharp and concordant with the foliation in the gabbro, but quartz diorite dikes cut the gabbro not far away. Intrusive relations are best shown in the Jones Falls area in Baltimore. In the Woodberry quarry numerous leuco-quartz diorite dikes up to 50 feet thick ramify through the gabbro. In the Baltimore-Patapsco aqueduct, which passes underground near this area, Herz (1951, p. 1014) reports that the quartz diorite forms an injection complex in the gabbro nearly 1000 feet wide, and that the two rocks alternate in bands that average 50 feet thick.

The Relay Quartz Diorite is a light-colored, fine-grained, sugary-textured rock. A

strong lineation or foliation is common but is not conspicuous, because of the scarcity of dark minerals. Along the Baltimore and Ohio Railroad tracks west of Relay it is sheared out into phyllonite, with lustrous coatings of sericite and chlorite along the closely spaced shear surfaces. Elsewhere these same minerals heal more irregular, widely spaced shears, giving the rock a "seamed" appearance.

Silicification has accompanied shearing in most places. This is shown west of Relay, and also along Rockburn Branch on the south side of the Patapsco River. Quartz pervades the rock, occurring as minute lenticles parallel to the foliation and in concordant veinlets and pods up to 5 feet thick. The vein quartz is generally foliated and in some places sheared to ribbons. Silica-bearing solutions evidently permeated active shear zones, and quartz was deformed soon after it was deposited.

The microscopic textures of the quartz diorite also reveal evidence of deformation and recrystallization. The textures are mostly crystalloblastic, and the quartz and feldspar are segregated and aligned into a lineation or foliation. Some of the rocks are granoblastic. The original igneous texture is rarely well preserved, but some deformed relics survive, especially large crystals of zoned and complexly twinned plagioclase.

The Relay rocks consist chiefly of plagioclase and quartz (Tables 35, 36). Biotite, chlorite, epidote, and muscovite (chiefly sericite) are the main varietal minerals, and

TABLE 35

MODAL ANALYSES OF RELAY QUARTZ DIORITE

	1	2	3	4	5	6	7	8
Plagioclase	48.4	50.5	46.7	49.0	47.9	48.0	60.3	82.5
K feldspar	—	—	—	0.4	7.5	—	1.2	—
Quartz	48.3	44.6	44.4	43.5	39.8	43.1	30.1	3.4
Hornblende	—	—	—	—	—	—	3.7	8.6
Biotite	1.2	tr	tr	5.7	4.0	4.2	1.6	tr
Chlorite	0.4	1.1	0.3	tr	—	0.1	—	1.2
Muscovite	0.7	1.8	6.5	0.6	0.1	3.8	tr	tr
Garnet	—	—	0.6	—	0.2	0.1	0.6	—
Epidote	0.3	1.8	0.5	0.4	tr	0.5	0.1	3.7
Allanite	—	—	tr	0.3	—	0.1	0.1	—
Apatite	tr	tr	—	—	—	—	0.1	0.2
Sphene	—	—	—	0.1	—	—	—	—
Zircon	tr	—	tr	tr	tr	tr	—	—
Monazite	—	tr	tr	—	—	—	—	—
Magnetite	0.7	0.1	0.8	—	0.5	0.1	2.2	0.1
Hematite	—	—	0.2	—	—	tr	—	—
Calcite	—	0.1	—	—	—	—	—	0.3
Total	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0
Points	1,508	1,476	1,577	1,573	1,494	1,574	1,581	1,388

(1) Leuco-quartz diorite (BG-8). Woodberry Quarry, Jones Falls, Baltimore City.
(2) Leuco-quartz diorite (BG-11). Woodberry Quarry, Jones Falls, Baltimore City.
(3) Leuco-quartz diorite (H37-1). Patapsco River at Rockburn Branch, Howard County.
(4) Granoblastic quartz diorite (H53-1). Gwynns Falls near Edmondson Ave., Baltimore City.
(5) Granoblastic quartz diorite (H53-2). Gwynns Falls near Edmondson Ave., Baltimore City.
(6) Gneissic quartz diorite (H38-1). Patapsco River at Rockburn Branch, Howard County.
(7) Gneissic hornblende quartz diorite (H41-2), from a small dike-like mass in metagabbro. Patapsco River near Gun Road, Howard County.
(8) Leuco-diorite (BG-12). Woodberry Quarry, Jones Falls, Baltimore City.

TABLE 36

MODAL ANALYSES OF RELAY ALBITE GRANITE

	1	2	3	4	5
Albite	58.0	41.7	47.1	56.1	41.8
K feldspar	—	—	—	—	1.9
Quartz	37.8	45.7	42.2	41.6	39.1
Biotite	—	0.1	2.8	0.7	3.6
Chlorite	—	1.8	2.7	0.2	tr
Muscovite	3.4	10.7	4.9	0.3	8.9
Garnet	—	—	—	—	0.1
Epidote	0.1	—	—	tr	4.0
Allanite	0.1	—	—	—	—
Sphene	0.2	tr	tr	—	0.2
Zircon	—	tr	tr	tr	—
Magnetite	0.1	—	—	1.1	0.4
Hematite	—	—	0.3	—	—
Calcite	0.3	—	—	—	—
Clay	—	—	—	—	—
Total	100.0	100.0	100.0	100.0	100.0
Points	1,278	1,703	1,670	1,516	3,160

(1) Gneissic albite granite (H-13). Baltimore & Ohio Railroad tracks, near Relay, Baltimore County.

(2) Phyllonitic albite granite (H59-3). Baltimore & Ohio Railroad tracks, near Relay, Baltimore County.

(3) Gneissic albite granite (H59-1). Baltimore & Ohio Railroad tracks, near Relay, Baltimore County.

(4) Crystalloblastic albite granite (BG-10). Woodberry Quarry, Jones Falls near 40th Street, Baltimore City.

(5) Gneissic albite granite (H39-1). River Road near Rockburn Branch, Patapsco State Park, Howard County.

small amounts of garnet are commonly present. Potassium feldspar (microcline) is minor or absent. The plagioclase compositions are mostly in the range An_{0-30}, but more calcic plagioclase occurs in the less quartzose rocks. Based on plagioclase the rocks fall into two groups: quartz diorite, with plagioclase mostly An_{20-30}, and albite granite, with plagioclase An_{0-10} (Fig. 36). Quartz diorite is more abundant, but the albite granite is also widespread. At the type locality near Relay albite granite appears to predominate. In general, albite granite seems to be more common where the rock has been more intensely sheared and silicified.

The field relations between the quartz diorite and albite granite are shown in the Woodberry quarry in Baltimore. The quartz diorite occurs in dikes up to 50 feet thick cutting the gabbro, and both rocks have a common foliation and lineation which often transects their contacts. Within the dikes the pale-gray quartz diorite fades imperceptibly into small areas of white sugary-textured rock, identified in thin section as albite granite. The granite forms diffuse, irregular zones in the quartz diorite, and the lineation or foliation passes without interruption through both.

The albite granite appears to have replaced the quartz diorite rather than intruded it. This is partly confirmed by their microscopic relations, which also indicate that the quartz diorite was severely granulated in the area where replacement occurred. The quartz diorite shows relict hypidiomorphic texture in other parts of the quarry, but

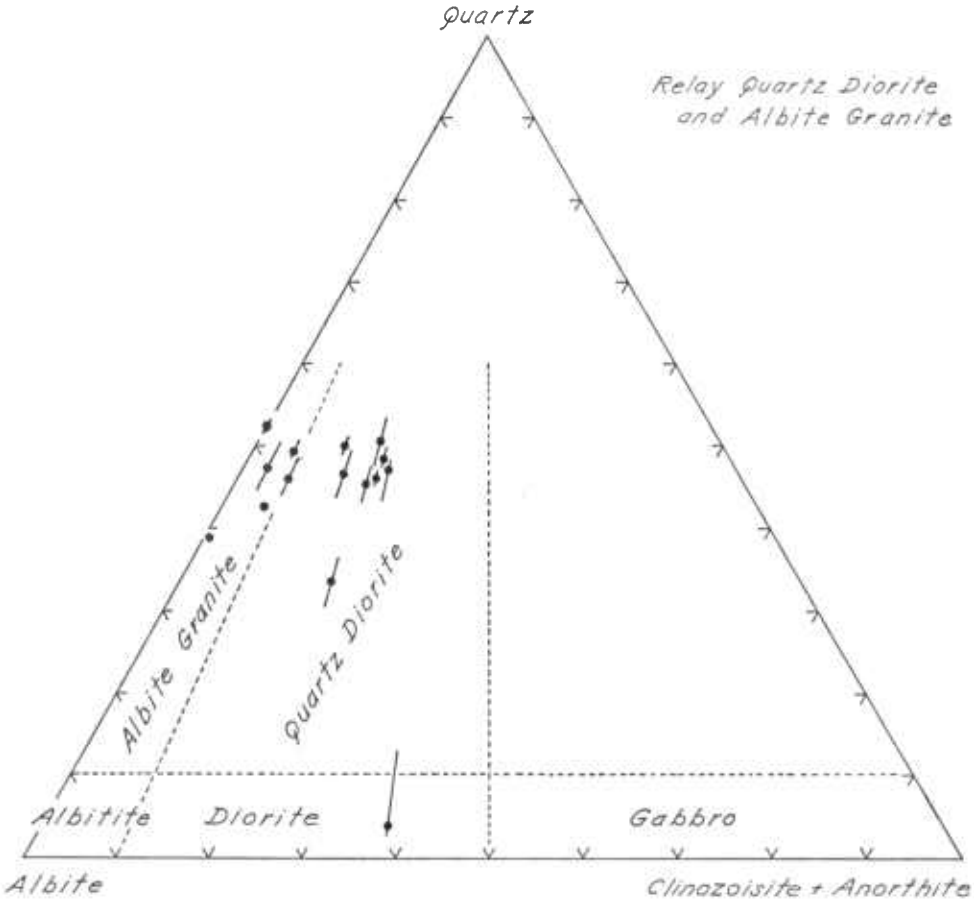


FIGURE 36. Modal composition of Relay quartz diorite and albite granite. The vertical distance between ends of the line through each point gives the percentage of mafic minerals.

adjacent to the albite granite it is thoroughly recrystallized to a gneissic or granoblastic texture. This texture is also found in the granite, where albite appears in place of oligoclase. The albite shows chessboard twinning, sieve structure, pseudo-myrmekitic and pseudo-micrographic intergrowths with quartz (Pl. 32, fig. 2), and other features suggestive of replacement. The albite granite appears to have formed by albitization of the quartz diorite along zones of granulation.

Clear-cut evidence for albite metasomatism is found in a 3-foot quartz diorite dike in the west face of the Woodberry quarry. Relict hypidiomorphic texture, with quartz molded around euhedral plagioclase, has partly survived even though the rock has been severely sheared. Most of the plagioclase has been bent and broken, and parts of the matrix crushed into mortar. Large lobes of myrmekitic albite have eaten into the plagioclases and replaced large portions of the matrix. Replacement myrmekite now constitutes at least 30 per cent of the rock and must have entailed addition of Na. The same rock contains thin, crosscutting albitite veinlets, in which tiny prismatic albites appear to be filling very narrow fissures. This also shows that Na was migrating through the rock.

The Relay albite granite may be compared with the Sparta granite in Oregon (Gilluly, 1933), formed by transformation of quartz diorite by albitization and silicification along

zones of crushing. Gilluly showed that the replacement occurred at a late magmatic or early postmagmatic stage, when the quartz diorite was almost solidified, by solutions derived from the lower part of the same mass. Transformation of the Relay quartz diorite to albite granite probably also occurred during and immediately following solidification, but subsequent metamorphism has destroyed much of the evidence needed for a firm conclusion.

Quartz diorite is one of the most common types of intrusive rock in the Maryland Piedmont, but the Relay Quartz Diorite has little in common with the others. It is distinctive in the following ways: (1) It is very leucocratic; the color index is less than 5, whereas color indexes of the other quartz diorites range from about 10 to 50. (2) It is very siliceous; it contains about 78 per cent SiO_2, more than even the granodiorites and quartz monzonites have (Table 37). (3) It has exceptionally low K_2O and high Na_2O (Table 37, col. 1). (4) It grades into albite granite; the other quartz diorites show a close relation to granodiorite and quartz monzonite. (5) It is closely associated with the

TABLE 37

CHEMICAL ANALYSES OF GRANITIC ROCKS

	1	2	3	4	5	6	7	8	9
SiO_2	77.94	56.15	59.62	68.53	63.73	65.39	71.01	73.08	77.01
TiO_2	0.13	0.75	0.70	0.22	0.52	0.47	0.30	0.27	0.10
Al_2O_3	11.75	18.11	13.52	14.08	17.28	17.13	14.46	13.86	13.47
Fe_2O_3	1.65	1.76	2.81	0.64	1.05	0.57	1.17	0.65	0.48
FeO	0.93	7.20	5.50	3.45	3.02	2.73	1.44	1.40	1.15
MnO	0.05	0.17	0.15	0.14	0.07	0.04	0.04	0.05	0.13
MgO	0.25	4.81	6.11	2.59	2.24	1.61	0.70	0.48	0.15
CaO	1.28	7.36	5.48	2.87	3.81	3.38	2.32	1.71	0.53
Na_2O	4.66	1.42	1.89	2.73	3.63	3.47	3.59	3.70	3.69
K_2O	0.76	0.78	1.53	2.39	3.53	3.90	4.38	4.09	2.18
H_2O^+	0.39	1.28	2.51	1.98	1.02	0.76	0.62	0.46	0.89
H_2O^-	0.07	0.02	0.10	0.02	0.04	0.03	0.03	0.10	0.01
CO_2	tr	0.10	tr	0.26	0.00	0.1	0.10	0.05	0.13
P_2O_5	0.03	0.08	0.14	0.06	0.28	0.23	0.12	0.06	0.06
ZrO_2	—	0.01	—	0.04	—	—	—	—	0.02
Total	99.89	100.00	100.06	100.00	100.22	99.81	100.28	99.96	100.00

(1) Relay quartz diorite (H37-1). Rockburn Branch at the Patapsco River. Patapsco State Park, Howard County. Analyst, O. von Knorring.

(2) Norbeck biotite-hornblende quartz diorite (M-71). Norwood Road, 1.5 miles south of Norwood, Montgomery County. Analysts, Penniman and Browne.

(3) Norbeck biotite-hornblende quartz diorite (M273-3). Greenwood Knolls near Georgia Avenue, Wheaton, Montgomery County. Analyst, O. von Knorring.

(4) Kensington biotite quartz diorite (M-202). Quarry at Broad Branch Road, Washington, D. C. Analyst, Penniman and Browne.

(5) Ellicott City biotite granodiorite (H-7A). River Road, 0.3 miles east of Patapsco River bridge at Ellicott City, Baltimore Co., Analyst, H. B. Wiik.

(6) Ellicott City biotite quartz monzonite (H18-1A). U. S. Route 29, 200 yards south of U. S. Route 40, Howard County. Analyst, H. B. Wiik.

(7) Woodstock muscovite-biotite quartz monzonite (H49-1). Sylvan Dell quarry, Granite, Baltimore County. Analyst, O. von Knorring.

(8) Guilford muscovite-biotite quartz monzonite (H-10). Middle Patuxent River, 1.5 miles west-southwest of Guilford, Howard County. Analyst, O. von Knorring.

(9) Bear Island biotite-muscovite granite (M-151). West end of Bear Island, Potomac River, Montgomery County. Analysts, Penniman and Browne.

Baltimore Gabbro. Knopf and Jonas (1929, p. 130) observed that "This rock [Relay quartz diorite] is associated in all its occurrences with gabbro. . . ."

For these reasons it is concluded that the Relay Quartz Diorite is not comagmatic with the other quartz diorites in the Baltimore-Washington area. It does, however, appear to be genetically related to the gabbro, as suggested by their close spatial relation and by the following evidence.

Leuco-quartz diorite resembling the Relay is the end member of a series of small dioritic and quartz dioritic bodies that appear to have been filter-pressed from the gabbro. The feldspathic net veins in the gabbro (described under Baltimore Gabbro Complex) are one example. Most of the net veins are leuco-gabbro, but some contain relict plagioclase zoned to about An_{40} and small amounts of quartz. Other small sinuous dikelike masses in the gabbro consist of diorite and quartz diorite. They commonly have diffuse, irregular contacts and appear to grade into the gabbro. These show all gradations from diorite with very little quartz (table 35, col. 8), to normal quartz diorite (Table 35, col. 7), to leuco-quartz diorite exceptionally rich in quartz. Paralleling the increase in quartz is a change from calcic andesine to oligoclase, and from hornblende to biotite as the chief dark mineral. Some of the plagioclases show relict zoning. In the Woodberry quarry in Baltimore some of these bodies, several feet wide, grade outward from a core of leuco-quartz diorite to a zone of dark hornblende quartz diorite, and then into uralite gabbro. The structural relations are compatible with the theory that these bodies were liquids squeezed from the gabbro during protoclastic deformation. A common lineation passes through all of them.

It is thus inferred that small quantities of quartz diorite liquid were derived from the gabbro, probably by filter pressing. The large mass of Relay quartz diorite is similar to the more siliceous of the small filter-pressed bodies, and is restricted in occurrence to the gabbro. Therefore it is tentatively concluded that the Relay quartz diorite also stemmed from the gabbro magma, but the manner of differentiation is not established. The quartz diorite was emplaced in its present position during regional deformation, and as it solidified its own residual fluids from a deeper level migrated through zones of protoclastic or cataclastic crushing, silicifying and albitizing the quartz diorite, and locally transforming it to albite granite. Deformation outlasted these stages, and the entire complex was recrystallized.

The volume of Relay quartz diorite and albite granite is probably less than 2 per cent of the gabbro complex in the Baltimore area—no greater than the relative amounts of granophyre that differentiate from diabase sills. On these grounds it is not unreasonable to suppose that the Relay quartz diorite was derived from the gabbro magma.

NORBECK QUARTZ DIORITE

An elongate pluton of dark quartz diorite (Fig. 37) in east-central Montgomery County is called the Norbeck Quartz Diorite, after the small community of that name near its center (Pl. 7). The pluton is $8\frac{1}{2}$ miles long and $1\frac{1}{2}$ miles wide and trends north parallel to the structure of the country rock. It is bounded on the north and west by the Sykesville Formation and on the south and east by the Wissahickon Formation. The Kensington Quartz Diorite also impinges on part of its eastern border. Another mass of the dark quartz diorite, 2 miles long and a quarter of a mile wide, crops out south of Norwood, a short distance east of the main body.

The contacts of the Norbeck pluton are not exposed, but the structural relations and its igneous texture indicate that the mass is intrusive. On the north it sends prongs

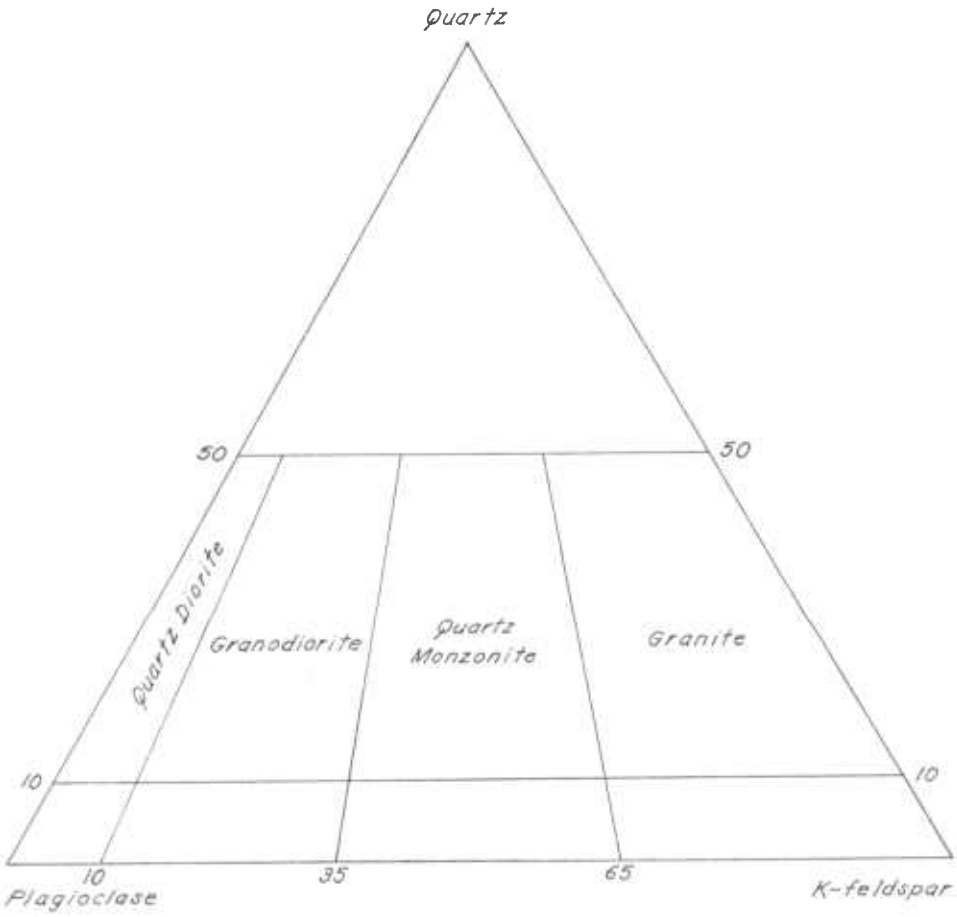
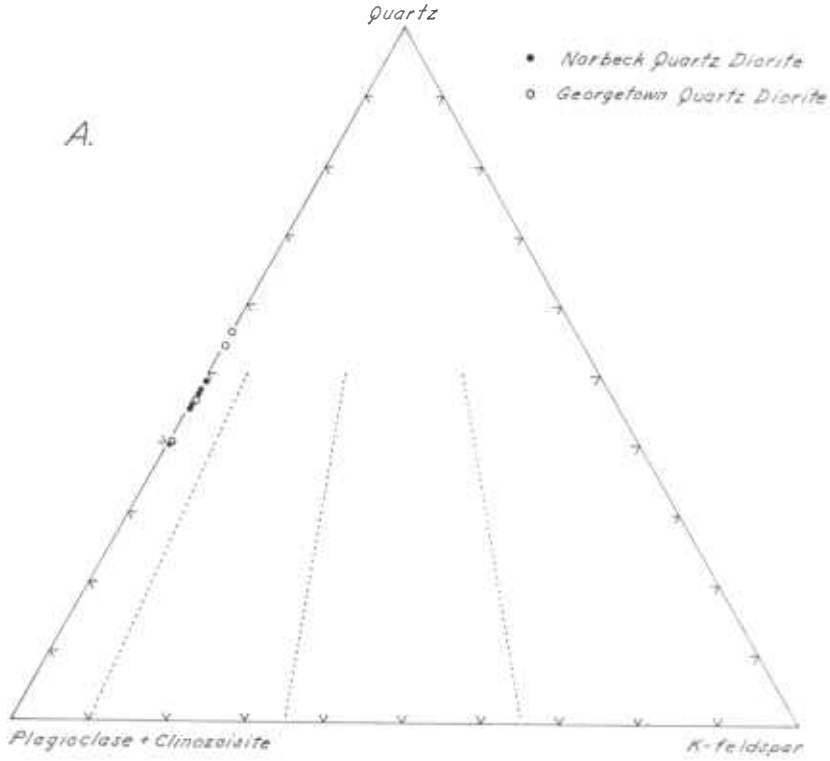


FIGURE 37. Classification of granitic rocks used in this report. Diagram modified after Johannsen (1931).

into the Sykesville Formation. At the blunt south end of the mass the schistosity in the Wissahickon Formation appears to be deflected around it, as if split apart during emplacement of the pluton. The foliation in the quartz diorite dips steeply and curves around parallel to this blunt southern contact, as might be expected if the mass were intrusive. Throughout the rest of the pluton its internal structure is also parallel with the contacts and with the structure of the country rock. The map shows the pluton cutting across the Sykesville-Wissahickon contact, but this contact is a gradual facies change which trends across Wissahickon bedding. The pluton is essentially concordant.

The rock consists chiefly of plagioclase, quartz, hornblende, and biotite, with lesser amounts of chlorite, clinozoisite, or epidote. Apatite, sphene, magnetite, and zircon are the common accessories. Modal analyses of typical specimens are given in Table 38 and plotted in Figure 38. The plagioclase is andesine and shows strong progressive zoning in the rocks that have not been strongly deformed and altered. The zoning ranges from An_{50} near the core to An_{25} at the rim. The hornblende is deep green to bluish green and shows strong pleochroism. Some of the large hornblende prisms contain rectangular patches of a much paler green hornblende. These may be remnants of uralite. The



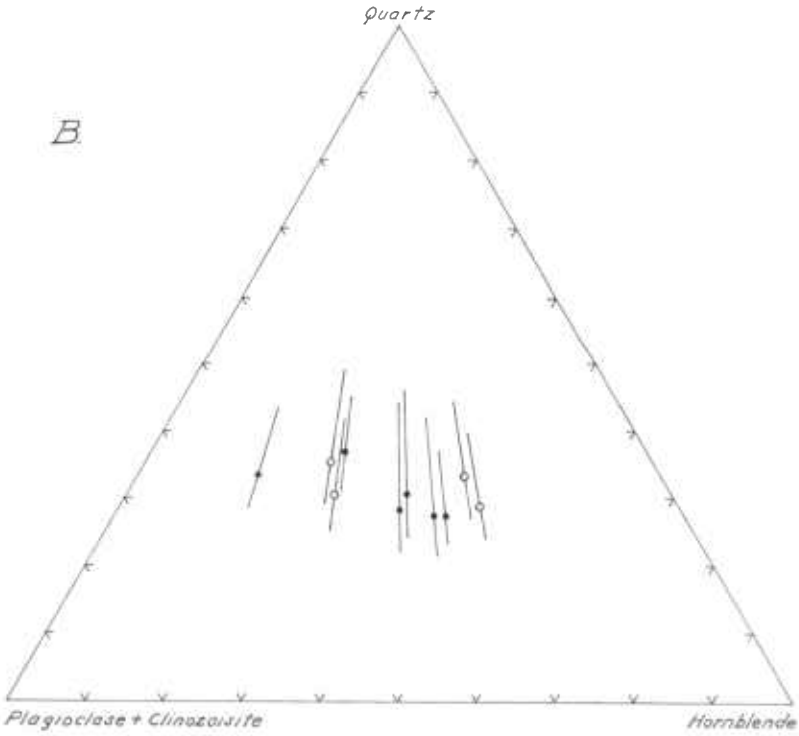


FIGURE 38. Modal composition of Norbeck and Georgetown Quartz Diorites. The dashed lines denote classification (Fig. 37).

A. Quartz — plagioclase (+ clinozoisite) — K feldspar diagram. The dashed lines denote classification (Fig. 37).

B. Quartz — plagioclase (+ clinozoisite) — hornblende diagram. The vertical distance between ends of the line through each point gives the per cent of biotite.

biotite is strongly pleochroic from deep brown to straw yellow, and much of it is partly altered to chlorite.

The massive rocks have a well-preserved igneous texture. The plagioclase is euhedral, and the zoning shows euhedral outlines. The hornblende occurs in stout euhedral to subhedral prisms, and some are molded around euhedral plagioclase. Biotite occurs in large flakes which partly enclose hornblende. Quartz fills the voids between all the other minerals. This hypidiomorphic texture shows a crystallization sequence that began with calcic plagioclase and perhaps clinopyroxene and progressed through hornblende, biotite, and more sodic plagioclase to quartz, in accord with Bowen's reaction series.

Deformation and recrystallization have blurred and in places obliterated the original hypidiomorphic texture in most of the Norbeck rocks. Plagioclase and quartz are drawn out into pods and thin lenses, and the ferromagnesian minerals are strung out in the same direction. Plagioclase has recrystallized; the more calcic cores yield dense aggregates of granular clinozoisite or epidote (Pl. 33, fig. 1). Quartz is broken and recrystallized into mosaics of smaller grains. The hornblende has also recrystallized and forms irregularly shaped crystals with pronounced sieve structure. All gradations are found from weakly foliated quartz diorites, with hypidiomorphic texture still well preserved, to strongly gneissic or even schistose rocks with recrystallized textures. However, even the most severely deformed rocks contain plagioclase relics with evidence of zoning, showing that they passed through a magmatic stage.

The Norbeck pluton was evidently emplaced as magma but was deformed and recrystallized as it solidified. The country rocks were probably undergoing deformation and metamorphism at the same time, and protoclasis in the pluton may have passed continuously into metamorphism. The lineation and foliation in the pluton is like that in the country rock (*see* Cloos and Cooke, 1953), indicating that they were deformed together.

QUARTZ DIORITE OF THE GEORGETOWN COMPLEX

Dark biotite-hornblende quartz diorite, closely resembling Norbeck quartz diorite, makes up part of the Georgetown Complex. The quartz diorite in the southeastern part of the complex is shown separately on the Geologic Map of Montgomery County (Cloos and Cooke, 1953), but to the northwest it was not differentiated from the more mafic rocks. Fisher (1963, Pl. 1) has mapped it separately (as hornblende quartz diorite) along the Potomac River near Glen Echo. On Plate 7 of this report the quartz diorite is not differentiated from the other rocks of the Georgetown Complex.

The contact relations are not well known. Most of the area in which the quartz diorite occurs is now residential district, and there are few outcrops. Fisher (1963, p. 66) describes the occurrence of the hornblende quartz diorite along the Potomac River near Glen Echo as follows:

> "It occurs mainly in elongate pod-shaped intrusions, set off by thin septa of metasedimentary rock, and contains numerous inclusions, among them fragments of the ultramafic rocks. The ultramafic inclusions suggest that the hornblende quartz diorite is younger than the ultramafic rocks, although no outcrops proving intrusions of the former into the latter have been found."

The dark quartz diorite is well exposed along Rock Creek below the Massachusetts Avenue bridge, in Washington, D. C. The rock there has a noticeable foliation, and along narrow zones it becomes intensely gneissic. The quartz diorite contains abundant dark lenticular inclusions of amphibolite, aligned with the foliation. Many of the inclu-

sions are in sharp contact with the host, but others are gradational, and some appear as nebulous dark "ghosts" in the quartz diorite.

Under the microscope the rock is indistinguishable from Norbeck quartz diorite. Plagioclase, quartz, hornblende, and biotite are the chief minerals, and the main accessories are epidote, apatite, magnetite, and zircon (Table 38). A distinctive feature is the presence of abundant coarse grains of rutile. Some of the plagioclase shows progressive zoning from An_{48} to An_{25}, but much of it is partly altered to coarse aggregates of clinozoisite or epidote. Chlorite partly replaces the biotite.

The textural description of the Norbeck quartz diorite holds equally well for the biotite-hornblende quartz diorite at Georgetown. Hypidiomorphic texture showing the normal magmatic crystallization sequence is still preserved, but in some rocks it is

TABLE 38

MODAL ANALYSES OF NORBECK AND GEORGETOWN QUARTZ DIORITES

	1	2	3	4	5	6	7	8	9	10
Plagioclase	26.5	26.4	23.3	23.1	22.6	16.9	26.8	28.3	19.1	18.1
K feldspar	—	—	—	—	—	—	—	—	—	—
Quartz	28.3	31.3	21.4	23.4	22.1	23.5	25.0	28.6	24.0	27.3
Hornblende	12.8	21.0	32.1	36.3	27.5	27.3	21.7	19.8	38.4	34.4
Biotite	11.3	13.7	15.1	11.2	15.3	16.4	11.9	12.8	11.3	13.3
Chlorite	3.7	.3	5.5	2.5	6.8	5.6	4.7	5.1	4.5	4.0
Sericite	—	.1	.4	—	.2	—	.1	tr	—	—
Epidote	16.4	5.8	1.6	2.7	3.9	9.2	9.1	4.4	1.8	2.4
Allanite	—	—	—	—	—	—	—	—	tr	—
Apatite	.1	.2	.4	.2	.5	.3	tr	.1	.4	.2
Tourmaline	—	—	—	—	tr	—	tr	—	—	—
Sphene	tr	tr	—	tr	—	—	tr	.1	.3	.2
Rutile	—	—	—	—	—	—	.5	.2	.2	.1
Monazite	tr	—	tr	tr	—	—	—	tr	—	—
Zircon	tr	tr	tr	tr	tr	tr	.1	tr	tr	tr
Magnetite	.9	1.2	.2	.6	1.1	.8	.1	tr	tr	—
Calcite	—	—	—	—	—	—	—	.6	tr	—
Total	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0
Points	1623	1110	2221	2442	1792	1588	1176	1224	1810	1232

Norbeck Quartz Diorite

(1) Gneissic quartz diorite (M272-1). Md. Route 609 at Norwood Road.
(2) Massive quartz diorite (M-69). Glenmont Road, 1 mile south of Norwood.
(3) Massive quartz diorite (M-8). One mile north of Olney.
(4) Gneissic quartz diorite (M-71). Glenmont Road, 2 miles south of Norwood.
(5) Gneissic quartz diorite (M273-1, B-47). Green Wood Knolls, Wheaton.
(6) Gneissic quartz diorite (M-75). Two miles south of Norbeck.

Georgetown Quartz Diorite

(7) Massive quartz diorite (M-99). Glen Cove, 2 miles south of Bethesda.
(8) Gneissic quartz diorite (M274-1b). Rock Creek below Massachusetts Avenue bridge, Washington, D. C.
(9) Gneissic quartz diorite (M274-1a). Rock Creek below Massachusetts Avenue bridge, Washington, D. C.
(10) Gneissic quartz diorite (M274-1). Rock Creek below Massachusetts Avenue bridge, Washington, D. C.

blurred by later deformation and recrystallization. Strongly sieved hornblendes, like those in the Norbeck, are very prominent in the more strongly recrystallized parts of the Georgetown quartz diorite.

The Norbeck and Georgetown biotite-hornblende quartz diorites are not directly connected at the surface, but their petrography is so similar that they may represent pulses of the same magma.

A lighter-colored biotite-quartz diorite is also widespread in the Georgetown Complex. It crops out along a belt extending from near American University in Washington, D. C., to Bethesda, and also in smaller concordant bodies along the Potomac River near Cabin John. Still smaller intrusions and dikes of the same rock are found along lower Rock Creek in Washington, D. C. A dike of this rock cuts the darker hornblende-quartz diorite along lower Rock Creek. Fisher (1963, p. 70) cites additional evidence that the biotite-quartz diorite is younger.

The biotite-quartz diorite consists chiefly of plagioclase and quartz, with biotite as the main dark mineral. Some of the plagioclase is progressively zoned, An_{45} to An_{25}. There is no potassium feldspar. Epidote, magnetite, apatite, and zircon are the main accessory minerals. The rock typically shows a well-defined foliation, and the microscopic textures show evidence of considerable crushing, granulation, and recrystallization. Fisher (1963, p. 68–70) gives a more detailed description.

KENSINGTON QUARTZ DIORITE

A thin wedge of quartz diorite gneiss runs north from lower Rock Creek Park in the District of Columbia to Norwood, a distance of 15 miles. Its name is taken from Kensington, located near the middle of the wedge. North of Wheaton the gneiss pinches out, then reappears and splits into two prongs. A parallel lens of gneiss, 5 miles long and 300–400 yards wide, lies a short distance east. Other masses of Kensington gneiss occur at Glen Echo, and 1 mile northwest of Georgetown, and very small lenses cut Wissahickon schist in the Bethesda-Chevy Chase area. All but the smallest of these bodies are shown on Plate 7.

Kensington quartz diorite has invaded the Wissahickon schist as thin concordant wedges and lenses. The intrusions are more irregular where they cut the plutonic rocks of the Georgetown Complex. The intrusive relations of the Kensington to these older plutonic rocks are described by Fisher (1963, p. 72):

> "Both along Rock Creek, north of Boundary Oaks Park, and along the Potomac River, near Sycamore Island, the Kensington Gneiss cuts across the trend of the foliation in the hornblende quartz diorite, and the map pattern suggests that the Kensington intrudes the hornblende quartz diorite. In addition, thin apophyses of Kensington Gneiss intrude the hornblende quartz diorite and associated biotite quartz diorite along the Virginia bank of the Potomac, west of Snake Island. These facts indicate that the Kensington Gneiss is younger than both the hornblende quartz diorite and the biotite quartz diorite."

The composition of the Kensington gneiss overlaps from quartz diorite to granodiorite (Fig. 39). Plagioclase (An_{22-32}) and quartz are the chief light minerals, and variable amounts of microcline are commonly present (Table 39). Biotite, the main dark mineral, is accompanied by muscovite and clinozoisite (or epidote). The accessory minerals include garnet, allanite, apatite, sphene, tourmaline, magnetite, zircon, and monazite. A garnet from the Kensington gneiss was identified as spessartite ($Sp_{77}Gr_{23}$) by G. W. Fisher (1963, p. 71).

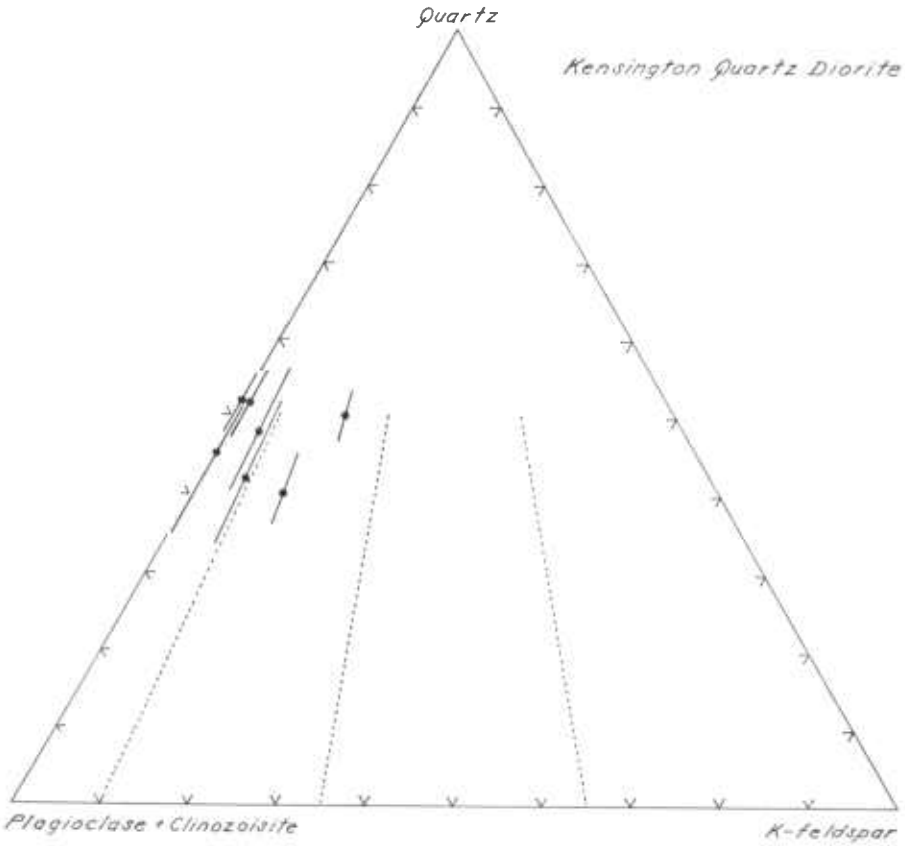


FIGURE 39. Modal composition of Kensington Quartz Diorite. The vertical distance between ends of the line through each point gives percentage of biotite.

The rock has a metamorphic texture. Large plagioclases are broken and drawn out into augen. Finer plagioclase and quartz are recrystallized into small anhedral grains and segregated into thin lenticles and ribbons. Small micas and granular clinozoisite cluster in the feldspathic lenses; the clinozoisite is chiefly derived from the plagioclase. The larger biotites and muscovites are strongly segregated into thin lenticular masses, which wrap around the feldspathic lenses and augen (Pl. 33, fig. 2). It is evident that the rocks have been granulated and recrystallized.

Few traces of original magmatic texture survive. Large plagioclase relics show complex twinning and progressive zoning, probably indicative of magmatic origin. Euhedral allanites, with sharp oscillatory zones that reflect the outline of the crystal, are probably also magmatic relics. They are mantled with epidote and closely resemble the magmatic allanites in the Ellicott City and Woodstock plutons (Pl. 35, fig. 2).

Along the main belt of the Kensington Quartz Diorite the intensity of deformation increases to the north. At Broad Branch, west of Rock Creek Park in Washington, the quartz diorite is only moderately gneissic, and the foliation could be taken for magmatic flow structure. At Rock Creek, 2 miles farther north, the rock has an intense foliation, and beyond that it changes gradually to a feldspathic schist. Along the Baltimore and Ohio Railroad in Kensington the rock is strongly schistose, with abundant muscovite

TABLE 39

MODAL ANALYSES OF KENSINGTON QUARTZ DIORITE

	1	2	3	4	5	6	7
Plagioclase	36.1	32.1	33.1	34.6	32.1	39.1	27.2
Microcline	—	—	tr	3.6	3.9	9.3	9.9
Quartz	34.2	44.6	39.0	39.2	32.2	35.4	39.4
Myrmekite	—	—	—	0.2	—	.2	.3
Biotite	17.6	7.4	6.8	15.8	18.4	8.7	5.0
Chlorite	.1	—	.5	—	—	—	—
Muscovite	6.7	7.5	17.3	2.8	5.0	2.7	16.3
Clinozoisite	4.7	8.3	2.8	3.7	8.1	4.3	1.2
Allanite	.1	—	—	—	.1	.2	tr
Garnet	—	.1	—	—	—	—	.3
Tourmaline	—	—	.1	—	tr	tr	—
Apatite	.2	—	.1	tr	.2	.1	.2
Sphene	—	tr	.3	—	—	—	.2
Monazite	tr	—	tr	tr	—	—	tr
Zircon	.1	tr	tr	.1	tr	tr	tr
Magnetite	.2	—	tr	tr	—	—	tr
Total	100.0	100.0	100.0	100.0	100.0	100.0	100.0
Points	1696	1617	1680	1644	1513	1753	1804

(1) Gneissic quartz diorite (M268-1). Beach Drive at Wyndale Avenue, Rock Creek Park, Washington, D. C.

(2) Protoclastic quartz diorite gneiss at contact (M265-4). Broad Branch Road south of Davenport Avenue, Washington, D. C.

(3) Schistose quartz diorite (M270-1). Baltimore and Ohio Railroad near Stoneybrook Drive, South Kensington.

(4) Gneissic quartz diorite (M266-1). Broad Branch, 400 feet north of Davenport Avenue, Washington, D. C.

(5) Gneissic quartz diorite (M265-1). Broad Branch Road south of Davenport Avenue, Washington, D. C.

(6) Gneissic granodiorite (M267-1). Broad Branch Road, about ¼ mile northwest of Davenport Avenue, Washington, D. C.

(7) Gneissic granodiorite (M269-2). Stoneybrook Drive at Forsyth Avenue, Rock Creek Park, Washington, D. C.

growing along closely-spaced shear planes. The quartz diorite was crushed and recrystallized after its solidified.

Some of the deformation is probably protoclastic, however. Along Rock Creek the intensely foliated gneiss contains scattered box-shaped porphyroblasts of microcline. The porphyroblasts are undeformed, although the gneiss in which they have grown has been intensely granulated. The microclines grew by replacing the granulated matrix after most of the movement had ceased. The microcline is restricted to the gneiss and does not occur in the wall rock, which suggests that it grew from residual potassic fluids derived from the Kensington magma. These fluids were still permeating the rock after much of the crushing and granulation had been accomplished, indicating that movement accompanied the solidification of the mass.

The Kensington Quartz Diorite shows a striking relation to regional structure. The main body and its satellites form thin, steeply dipping concordant sheets or wedges

localized along the plunging crest of the Baltimore anticlinorium. A tendency for the Wissahickon schist to split apart, here where tension developed across the crest of the growing anticline, may have permitted easier penetration by the Kensington magma. During slow solidification the sheets of soft quartz diorite mush probably formed zones of weakness, along which tectonic movement was concentrated. Kneading and flowage of nearly solidified quartz diorite resulted in thorough protoclasis, and recrystallization of the magmatic rock to gneiss.

ELLICOTT CITY GRANODIORITE

The Ellicott City pluton is a mass of biotite granodiorite and quartz monzonite shaped like a double phacolith, situated along the Howard-Baltimore County line (Fig. 12; Pl. 7). Its field relations and petrography have been studied by Hobbs (1888), Keyes (1893; 1895, p. 696–717), and Knopf and Jonas (1929; p. 134–135). The most detailed work, however, has been on its structure and petrofabrics (Cloos, 1933; 1947).

The pluton lies along the folded contact between the Wissahickon Formation and the Baltimore Gabbro Complex. The contact between granodiorite and metagabbro is generally sharp and concordant. Except for granitic dikes which cut the metagabbro at Thistle and Ilchester there is little evidence of diking and stoping. Along the Wissahickon contacts, however, the granodiorite has split the schist apart like the pages of a book, penetrating it in large wedges and dikes, and in stringers down to a few millimeters thick. The large schist re-entrant between Ellicott City and Oella is strongly injected in this manner. The schist is locally feldspathized on a small scale along the innermost contact. Oligoclase porphyroblasts cluster in the schist, forming small replacement dikes and more irregular feldspathized zones. This is the only evidence of contact metamorphism.

The granodiorite has well-developed flow structure toward the margins of the pluton, but it is weak or absent in the interior. Aligned biotites, feldspar phenocrysts, and flat elongated mafic inclusions are the chief flow elements. Planar structure is parallel to the contacts, and lineation plunges steeply. Cloos (1947) describes these relations in detail.

There is little evidence of stoping at the present level. Few xenoliths of Wissahickon schist are found in the granodiorite, despite their intricate intrusive contact. The granodiorite magma appears to have been forced into the schist, wedging it apart but stoping and assimilating very little of it. Dark flattened inclusions and schlieren are very abundant in the granodiorite, but few have come from Wissahickon schist. This is shown by their mineralogy: the inclusions are chiefly biotite and plagioclase, whereas the wallrock schist has abundant muscovite and quartz but only small amounts of biotite. The inclusions contain epidote, sphene, and hornblende, but no garnet, whereas the schist commonly has abundant garnet but lacks the other minerals. Moreover, there is no evidence that mica schist xenoliths have reacted with the magma to form the dark biotitic inclusions.

These inclusions were evidently derived from the underlying basement. They have about the same mineralogy and bulk composition as the dark migmatitic layers in the Baltimore Gneiss (Pls. 1, 2, 3). Moreover, schlieren-like masses of migmatitic gneiss up to 30 feet long are locally found in the granodiorite, some of them partly disrupted into smaller biotitic inclusions (Pl. 34, fig. 2). The migmatitic masses were evidently in a softened condition and flowed with the enclosing granodiorite. This is shown by their sinuous form, their contorted internal structure and contacts, and by their disruption into smaller fragments that are streaky or wavy rather than angular (Pl. 34, fig. 2).

The biotitic inclusions and masses of migmatite could represent softened remnants of blocks stoped from the basement. The dark inclusions are so abundant, however, that the question arises: why would stoping be so extensive in the basement and so negligible in the Wissahickon schist and the metagabbro? A more reasonable explanation is that the granodiorite originated by anatexis of migmatitic basement, and the dark inclusions are unmelted remnants of the more refractory biotitic layers in the migmatite which have been disrupted and smeared out by flowage. Similar dark inclusions have originated this way in the diapiric Gunpowder Granite (p. 46–47), where anatexis can be demonstrated.

The composition of the pluton ranges from granodiorite to quartz monzonite (Fig. 40A). In general, granodiorite comprises the margins of the mass and grades to quartz monzonite toward the interior. The pluton also tends to be darker (more biotitic) and nonporphyritic near the margins. Large microcline phenocrysts become more abundant toward the interior of the mass. Thus the Ellicott City resembles some of the zoned plutons of the Sierra Nevada (Moore, 1963, p. 115–116; Bateman *et al.*, 1963, p. D30–D32) and other western batholiths (Vance, 1961).

The granodiorite and quartz monzonite consist chiefly of plagioclase quartz, potassium feldspar, and biotite. The minor minerals are epidote, allanite, apatite, sphene, zircon, monazite, and magnetite. Some of the rocks contain small amounts of green hornblende, and others have muscovite. Sericitic muscovite is also present as a minor alteration product, together with chlorite and calcite. Modal analyses of the granodiorite and quartz monzonite are given in Tables 40 and 41, and those of some dike rocks and granitized wall rocks (Wissahickon schist) in Table 42. New chemical analyses of the dark granodiorite from the border of the pluton and porphyritic quartz monzonite from its interior are given in Table 37, cols. 5–6.

Plagioclase forms large euhedral and subhedral crystals, with the composition about An_{25-32}. Some crystals show progressive and oscillatory zoning, but the zones do not span a wide composition range. The potassium feldspar is chiefly microcline; most of it is nonperthitic or cryptoperthitic. Microscopically visible perthite is patchy and often occurs only where grains have been strained (Chayes, 1952).

Magmatic texture is generally well preserved except near the borders of the pluton, where protoclastic flowage has occurred. The minerals began to crystallize in approximately the following order: (1) plagioclase, allanite, apatite; (2) epidote, hornblende; (3) biotite; (4) potassium feldspar, quartz, muscovite. Primary crystallization was followed by growth of myrmekite and by other deuteric replacements.

Epidote is a primary magmatic mineral in the Ellicott City Granodiorite, as Hobbs (1889) and Keyes (1893) first pointed out. It forms large euhedral crystals that sharply enclose euhedral zoned allanite. Biotite and other pyrogenic minerals are molded around the epidote (Pl. 34, fig. 1). Deuteric or hydrothermal epidote also occurs in some of the rocks, chiefly as small granules replacing plagioclase. According to Dietrich (1961, p. 42–45), high pressure and/or high hydroxyl content in the magma favor primary epidote.

The microcline had a complex paragenesis, growing as orthomagmatic crystals, as deuteric replacements, and as metasomatic porphyroblasts. The orthomagmatic microcline occurs: (1) as anhedral grains that fill the interstices between earlier pyrogenic minerals, and (2) as large phenocrysts. The phenocrysts poikilitically enclose small grains of plagioclase, biotite, quartz, apatite, and other minerals, showing that the phenocrysts were forming at a late stage of magmatic crystallization. They are distin-

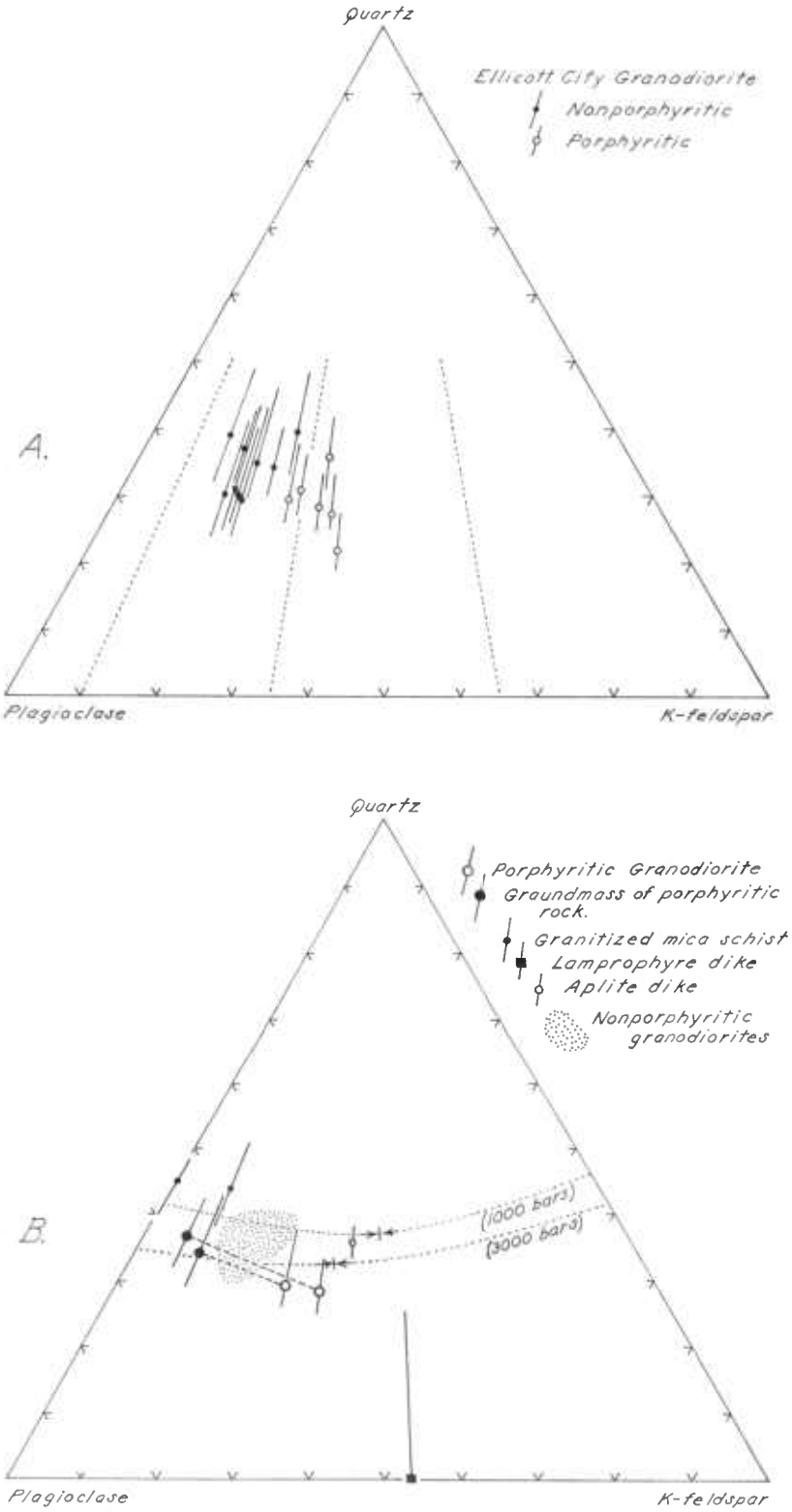

Quartz
Ellicott City Granodiorite
Nonporphyritic
Porphyritic
A.
Plagioclase
K-feldspar
Quartz
Porphyritic Granodiorite
Groundmass of porphyritic rock.
Granitized mica schist
Lamprophyre dike
Aplite dike
Nonporphyritic granodiorites
B.
(1000 bars)
(3000 bars)
Plagioclase
K-feldspar

FIG. 40

TABLE 40

MODAL ANALYSES OF NONPORPHYRITIC GRANODIORITES

Ellicott City Granodiorite

	1	2	3	4	5	6	7	8	9
Plagioclase	40.3	43.8	42.0	42.2	45.4	41.9	40.7	40.7	33.3
K feldspar	8.1	11.4	11.4	12.2	12.9	12.5	12.6	16.2	15.3
Quartz	31.1	23.7	31.2	24.7	25.5	23.1	28.7	29.5	31.4
Myrmekite	.6	1.1	1.7	.5	—	.2	.4	2.1	.9
Hornblende	—	.4	—	1.9	.2	1.1	.6	—	—
Biotite	16.3	16.2	10.2	16.2	14.2	17.4	15.5	9.6	13.3
Chlorite	.3	—	.2	—	—	—	—	tr	—
Muscovite	.9	tr	2.9	tr	tr	tr	tr	.7	4.2
Epidote	1.3	1.9	.2	1.5	.8	2.5	.5	1.0	1.1
Allanite	.3	.4	tr	.2	.1	.5	.1	tr	tr
Apatite	.5	.7	.2	.2	.5	.3	.3	.2	.3
Sphene	.2	.4	—	.3	.3	.5	.6	tr	.2
Monazite	.1	tr	tr	—	tr	tr	tr	tr	—
Zircon	tr	tr	tr	tr	tr	tr	tr	tr	tr
Magnetite	tr	tr	tr	.1	.1	tr	—	tr	—
Calcite	tr	tr	tr	tr	tr	—	—	tr	—
Total	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0
Points	1,849	1,852	1,707	2,070	1,495	2,854	1,445	1,876	1,403

(1) H20-1. One-half mile southeast of Oella, Baltimore County.
(2) H22-1. Cooper Branch, 0.4 mile above Patapsco River, Baltimore County.
(3) H19-1A. National Pike, near Old Frederick Road, Baltimore County.
(4) H108-2. Ellicott City, 200 yds. southeast of Patapsco River Bridge.
(5) B-21. Weber Quarry, Ellicott City.
(6) H-7. Weber Quarry, Ellicott City.
(7) H116-1. Highway U. S. 40, 0.4 mile west of St. Johns Lane, Howard County.
(8) H21-1. Cooper Branch, 0.7 mile above Patapsco River, Baltimore County.
(9) H-2 Patapsco River, 0.3 mile above bridge at Ellicott City.

guished from the later porphyroblasts by the alignment of included biotite flakes and small plagioclase laths parallel to the faces of the phenocryst; these tabular inclusions were pushed aside and oriented along the faces of the growing phenocryst before being enclosed. In the porphyroblasts the inclusions are random. The deuteric microcline forms small lobes and rinds that eat in along the grain boundaries between plagioclase and quartz, corroding their margins and replacing them. Where the rock has been protoclastically deformed, the microcline extensively replaces the crushed and granulated material.

Eskola (1956), Marmo (1960), and others have shown that some newly solidified granitic masses were "granitized" by their own potassic fluids, which rose from below.

FIGURE 40. Modal composition of Ellicott City Granodiorite.

A. Comparative composition of the porphyritic and nonporphyritic rocks. The vertical distance between ends of the line through each point gives percentage of mafic minerals (chiefly biotite). Dotted lines denote classification (Fig. 37).

B. Diagram showing bulk composition and groundmass composition of two porphyritic rocks (large circles); and the composition of various minor rocks. The vertical distance between ends of the line through each point gives percentage of mafic minerals + muscovite. The curved dotted line is the trace of the granite minimum in the synthetic granite system (Tuttle and Bowen, 1958, p. 75).

TABLE 41

MODAL ANALYSES OF PORPHYRITIC GRANODIORITE AND QUARTZ MONZONITE

Ellicott City Granodiorite

	1	2	3	4	5	6
Plagioclase	41.6	40.8	33.7	39.6	38.6	39.8
K feldspar	19.3	21.0	21.4	25.4	26.3	29.7
Quartz	25.1	27.2	30.4	24.9	24.0	19.0
Myrmekite	.3	.1	.7	.4	.9	.7
Hornblende	—	tr	—	—	tr	.3
Biotite	11.5	9.0	10.5	8.0	8.4	8.1
Chlorite	.1	tr	tr	—	.1	.2
Muscovite	.1	.6	2.7	.1	.2	.2
Epidote	1.1	.8	.3	1.0	.8	1.5
Allanite	.2	.1	.1	.1	.1	.1
Apatite	.2	.2	.2	.2	.2	.2
Sphene	.4	.1	—	.3	.4	.2
Monazite	tr	tr	tr	—	tr	—
Zircon	tr	tr	tr	tr	tr	tr
Magnetite	.1	.1	—	—	—	—
Calcite	—	tr	—	tr	tr	tr
Total	100.0	100.0	100.0	100.0	100.0	100.0
Points	4,896	3,307	3,002	2,489	2,032	4,425

(1) H-3. Ellicott City, Main and Church Streets.
(2) H-6. Thistle, Baltimore County.
(3) H-5. One mile northwest of Thistle, Baltimore County.
(4) H18-1. Highway U. S. 29 near U. S. 40, Howard County.
(5) H108-1. Ellicott City, 200 yards southeast of Patapsco River bridge.
(6) H142-1. Highway U. S. 40, 0.2 mile west of St. Johns Lane, Howard County.

This was the case at Ellicott City, as shown by the replacement textures in the granodiorite and by the following field evidence:

(1) Granodiorite along the border of the pluton near Thistle is strongly foliated, with well-aligned biotites and long streaked-out schlieren. Rectangular microclines up to 1 inch long are *randomly oriented*, however, indicating that they grew after movement had ceased. Seen under the microscope these crystals replace protoclastically deformed matrix but are unstrained themselves. Thus, they grew as porphyroblasts following protoclasis.

(2) Along the margin of the pluton, east of the bridge at Ellicott City, intensely gneissic zones a few inches wide cut discordantly through weakly foliated granodiorite. These were zones along which movement was localized during protoclastic deformation. Box-shaped microcline porphyroblasts up to an inch long are densely clustered along these movement zones but are sparse in the surrounding rock. The porphyroblasts are undeformed and randomly oriented, again showing that they developed after movement had ceased. They evidently grew from potassic fluids that were concentrated along the protoclastic crush zones.

It is evident from these and other relationships that some microcline grew metasomatically in the newly solidified granodiorite before protoclasis had ended. Movement zones were especially favored sites for replacement, perhaps because they facilitated

TABLE 42

MODAL ANALYSES OF DIKE ROCKS AND GRANITIZED WALL ROCKS

Ellicott City Granodiorite

	1	2	3	4
Plagioclase	32.9	23.8	41.5	48.1
K feldspar	25.5	27.5	6.3	tr
Quartz	32.3	—	37.4	39.4
Myrmekite	.4	—	—	—
Hornblende	—	1.8	—	—
Biotite	4.7	14.7	12.8	7.3
Chlorite	—	1.1	tr	4.7
Muscovite	3.9	—	.2	.1
Epidote	.3	0.1	.7	tr
Allanite	tr	0.4	.1	—
Apatite	tr	1.0	.1	tr
Sphene	—	1.4	.1	.1
Zircon	tr	tr	tr	tr
Magnetite	—	1.4	.7	.2
Calcite	—	7.9	.1	.1
Nontronite	—	18.9	tr	—
Total	100.0	100.0	100.0	100.0
Points	1,388	1,278	1,792	1,565

(1) Aplitic dike (H-2a). Patapsco River, 0.3 mile above bridge at Ellicott City.

(2) Lamprophyre dike (H108-5). National Pike, 400 feet west of Cooper Run, Ellicott City.

(3) Granitic replacement vein in mica schist (H19-3). National Pike, near Old Frederick Road, Baltimore County.

(4) Granitic replacement vein in mica schist (H19-4). National Pike, near Old Frederick Road.

entry of potassic fluids. Orville's (1963) thermodiffusion mechanism provides a ready explanation for the migration of potassium into the cooling top and sides of the pluton.

As the pluton changes from the margins inward to a more potassic granodiorite and quartz monzonite it becomes more weakly foliated or massive. Here the orthomagmatic textures are less complicated by protoclasis and porphyroblastic development of late K-feldspar. The rocks also tend to be more porphyritic, with flesh-colored phenocrysts of microcline-microperthite or cryptoperthite up to an inch or more long. These poikilitically enclose biotite and small plagioclase laths, which are often arranged in zones and aligned parallel to the faces of the phenocryst, proving that it grew in a melt. However, some of the phenocrysts have irregular extensions from their margin that penetrate along the grain boundaries between small plagioclase and quartz grains in the matrix and partly replace them, which shows that the phenocryst completed its growth after the matrix had crystallized.

The strongly porphyritic rocks have less K-feldspar in their matrix than the nonporphyritic rocks, even though the total K-feldspar content is greater (Tables 40 and 41). This is shown quantitatively in Figure 40B for two porphyritic rocks whose matrix composition and total modal composition were determined accurately by point-counting and by counting the phenocrysts with a grid on large rock slabs; in each case the matrix is relatively deficient in K-feldspar. This indicates that K-feldspar which ordinarily would have crystallized in the matrix migrated to the phenocrysts and added to their

size. It provides further evidence that the phenocrysts were still growing during the last stage of solidification.

The zonation of the pluton inward from granodiorite (Table 37, col. 5) to porphyritic quartz monzonite (Table 37, col. 6) involves an increase in K_2O and SiO_2 and decrease in the other constituents. Mineralogically the biggest change is the marked increase in K-feldspar (Fig. 40A; Tables 40–41). These effects are not well explained by marginal assimilation of wallrock, because: (1) field relations show that stoping and assimilation are negligible; and (2) the wallrocks have the wrong compositions to produce the observed effects. Assimilation of Wissahickon schist, composed mainly of muscovite and quartz, would enrich the margins of the pluton in K and Si, not the interior. Assimilation of the metagabbro would produce a hornblendic border, but hornblende is very scarce (Table 40).

The metasomatic introduction of K-feldspar along protoclastic movement zones at the margins of the pluton, described earlier, suggests a possible explanation for the zoning. The same fluids or ions that entered and locally granitized the margins of the pluton might have migrated from the deeper, hotter parts of the magma into the granodioritic mush toward the cooling top of the mass, enriching its residual liquid in the constituents necessary to crystallize excess K-feldspar. This explanation accords with the textural relations, which indicate that much of the K-feldspar formed at a late magmatic stage and even continued to grow by replacing the matrix. The means by which K and Si may have migrated upward from the deeper parts of the magma is not known, but experimental work has shown at least two possibilities: (1) alkalis will diffuse through fluid or melt along a thermal gradient, and tend to enrich the rock at the cooler end in K-feldspar (Orville, 1963); (2) hydrous fluid, which would rise toward the top of the magma chamber, could transport Si and alkalis, especially K, in excess of the other constituents (Morey and Hesselgesser, 1951; Tuttle and Bowen, 1958, p. 89–91). Thus diffusion or vapor transport might enrich the upper interior parts of the mass in K and Si as it very slowly cooled from the top and sides.

The pluton has a varied suite of dike rocks, including granodioritic apophyses in the wall rock and late dikes of aplite, pegmatite, and lamprophyre that cut through its interior. Some of the dikes are strongly deformed, like those near the contact at Thistle. There thin granitic sills in amphibolite are so thoroughly granulated that they resemble beds of quartzite. The microscope shows that these rocks are intensely cataclastic granodiorite, with most of the large feldspars milled down to mortar and partly recrystallized. Two generations of Ellicott City dikes are well displayed below the bridge at Ilchester, cutting metagabbro. The earliest of these are granodiorite, which again show much cataclastic granulation. They are cut and dilated by several undeformed pegmatite dikes that emanate from the pluton. Most of the pegmatite and aplite dikes, however, are confined to the interior of the pluton, which suggests derivation from its residual liquid. Some show relations which link them closely with the pluton in time, like the aplitic and pegmatitic dikes in the old Weber quarry at Ellicott City. The granodiorite in the quarry has 2 prominant sets of steep joints that lie parallel and transverse to the long axis of the pluton and probably formed during its solidification. Very thin aplite and pegmatite dikes in the quarry faithfully follow the same two directions, obviously recording the entry of residual magma into early-formed joints. The composition of these and other aplitic dikes are near the synthetic granite minimum (Fig. 40B), as might be expected. One small lamprophyre dike (Table 42, col. 2), exposed in a roadcut near the Weber quarry, cuts the aplite and all other igneous structures.

In summary, the Ellicott City pluton is a crudely zoned body of granodiorite and porphyritic quartz monzonite, shaped like a steeply-plunging double phacolith. It was probably emplaced during the waning stages of orogeny and may be classed as late-kinematic. Petrofabrics show that structural elements in the granodiorite and surrounding metamorphic rocks match closely, indicating that they were deformed together (Cloos, 1947). The protoclastic textures toward the margins of the pluton, and the severely granulated dikes that cut the wall rock also seem to reflect solidification in an actively deforming terrane. Textures from the weakly foliated and massive interior of the pluton, however, show fairly undisturbed crystallization. Evidently the magma welled in gradually: the first batch of granodiorite—which now forms the outer zone—participated in the strong regional deformation, but, as the pluton gradually inflated, these tectonic movements decreased in intensity and the core crystallized with little disturbance. Enrichment in K and Si accompanied the later stages of crystallization.

WOODSTOCK QUARTZ MONZONITE

A small oval intrusion of massive biotitequartz monzonite pierces the center of the Woodstock gneiss dome in southwestern Baltimore County, 1½ miles northeast of Woodstock. The small community of Granite, the site of several large granite quarries, lies near its center. Quarrying operations, facilitated by the well-developed horizontal sheet structure and vertical jointing, first began in 1835, and for many years the area was an important producer of building stone. Keyes (1895) and Knopf and Jonas (1929, p. 131–133) have reported on the geology of the area and the petrography of the granitic rocks.

The stock is 1½ miles long and a little more than a mile wide, enclosed on all sides by Baltimore Gneiss. A long re-entrant of gneiss extends into the stock from the west, nearly dividing it in two. Except for the quarries, little fresh rock is exposed. The contacts are mapped from saprolite and float. Thus little is known of the contact relations. Numerous large angular xenoliths of Baltimore Gneiss are displayed in the quarry walls, however, showing that the quartz monzonite is younger. The xenoliths are described and illustrated by Keyes (1895, p. 723–724). The quartz monzonite is an exceptionally uniform, medium-grained rock, devoid of foliation or lineation.

Modal analyses are given in Table 43 and plotted in Figure 41. A new chemical analysis is listed (Table 37, col. 7), and an older analysis is reported by Knopf and Jonas (1929, p. 133). Plagioclase, quartz, and potassium feldspar are the principal minerals, followed by biotite, muscovite, and epidote. The important accessories are allanite, apatite, sphene, monazite, xenotime (?), and zircon. Plagioclase is progressively zoned, generally in the range An_{20-30}, and the average composition is about An_{27}. Clear albite rims many of the plagioclases.

Potassium feldspar is chiefly microcline with grid twinning. Some of the grains are perthitic, especially where they have been strained. Narrow borders of albite at the contacts between microcline and quartz may represent exsolved material.

Large prismatic epidotes with euhedral allanite cores are common and are obviously primary. Biotite, quartz, and microcline are molded around them. The allanites are euhedral and show striking oscillatory zones (Pl. 35, fig. 1); many of them are also twinned.

The granite has a well-developed hypidiomorphic granular texture (Pl. 35, fig. 2), showing the normal magmatic crystallization sequence. Deuteric alteration is evident in most specimens. Interstitial microcline locally embays and partially replaces plagio-

TABLE 43

MODAL ANALYSES OF WOODSTOCK QUARTZ MONZONITE

	1	2	3	4	5	6
Plagioclase	35.0	33.7	34.7	35.8	30.9	30.9
K feldspar	16.3	19.0	21.3	24.1	20.4	25.7
Quartz	31.8	28.8	27.6	26.3	29.1	28.1
Myrmekite	1.7	0.5	1.2	2.2	—	1.2
Biotite	7.2	7.6	6.7	5.2	9.1	6.4
Chlorite	0.1	—	0.1	0.1	tr	0.1
Muscovite	4.9	6.6	5.0	3.3	6.7	4.9
Epidote	2.4	3.1	2.7	2.5	3.4	2.2
Allanite	0.2	tr	0.5	0.3	0.1	0.2
Apatite	0.2	0.3	tr	0.1	0.1	0.1
Sphene	0.1	0.2	0.1	0.1	0.1	—
Zircon	tr	tr	tr	tr	tr	tr
Monazite, Xenotime	—	—	—	—	0.1	0.1
Magnetite	0.1	tr	0.1	tr	—	—
Calcite	tr	0.2	tr	—	tr	tr
Clays	—	—	—	—	—	0.1
Total	100.0	100.0	100.0	100.0	100.0	100.0
Points	1,619	4,819	1,546	1,649	1,544	2,964

(1) H51-2. Upper quarry, southwest side.
(2) B-6. Upper quarry, north side.
(3) H46-1. Lower quarry, northwest side.
(4) H50-1. Lower quarry, east side.
(5) B61. Upper quarry, north side.
(6) H47-1. Lower quarry, north side.
} Sylvan Dell quarries at Woodstock.

clase, and lobes of myrmekite eat into the microcline. Ragged muscovites cluster on some of the plagioclase, evidently as an alteration product. Other muscovite forms large flakes sharply bounded against biotite and appears to be primary.

There is no cataclasis or other evidence of severe deformation. The rock crystallized under static conditions. The primary epidote and muscovite suggest that the magma crystallized at relatively high water-vapor pressure (Dietrich, 1961, p. 42–45), probably at a depth of several miles.

GUILFORD QUARTZ MONZONITE

The Guilford Quartz Monzonite crops out east of the Clarksville gneiss dome near Guilford, Atholton, Simpsonville, and Oakland Mills. It occurs in numerous east- to northeast-trending lenticular bodies, several hundred yards to 3 miles long. Wissahickon Formation is the principal wall rock, but the southernmost mass penetrates Baltimore Gabbro and the Laurel Formation.

The plutons are discordant, cutting across bedding and schistosity of the Wissahickon Formation. Near Guilford the quartz monzonite truncates the Laurel Formation and penetrates the Baltimore Gabbro as a narrow wedge nearly 2 miles long. Five vertical quartz monzonite dikes cut sharply across gently dipping Wissahickon schist along the Guilford road, half a mile southeast of Atholton. Granitic rocks and associated pegmatite dikes intrude Laurel schist along the Middle Patuxent River, 1½ miles southwest of Guilford. Contacts of the plutons are not well exposed, but the uniform, inclusion-free

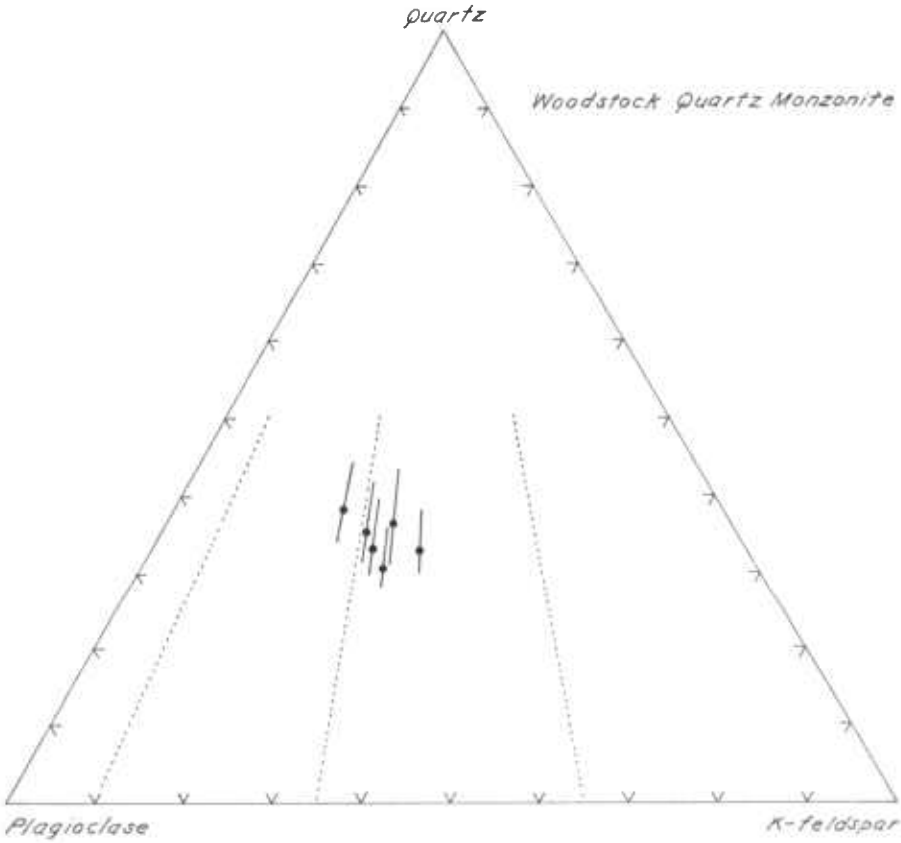


FIGURE 41. Modal composition of Woodstock Quartz Monzonite. The vertical distance between ends of the line through each point gives percentage of biotite. Dotted lines denote classification (Fig. 37).

character of the granite in exposures close to the contacts suggests that they are not complicated by stoping.

The lenticular form of the intrusions and their en echelon arrangment suggest that the magma may have been permissively emplaced along tension fractures, opening in the schist in response to a northeast-trending right-lateral couple. There is no evidence for stoping or forceful intrusion. Tiny biotitic inclusions are found but they are scarce even near the contacts. Foliation and lineation are lacking.

The quartz monzonite is light-colored, medium- to fine-grained, and exceptionally homogeneous. Plagioclase, microcline, and quartz in nearly equal proportions comprise about 90 per cent of the rock, and biotite plus muscovite make up most of the remainder (Table 44; Fig. 42). Apatite, spessartitic garnet, clinozoisite, zircon, and magnetite are the chief accessories. Plagioclase falls mostly in the composition range An_{18-24}; some zoned crystals contain slightly more calcic cores. Narrow rims of albite partly mantle some of the plagioclase, especially at the contacts with microcline.

The potassium feldspar is grid-twinned microcline. Locally it is perthitic, but the ubiquitous albite rims against plagioclase suggest that much exsolved material has migrated out of the microcline. Some of the microcline forms large "puddles" that poikilitically enclose quartz and plagioclase. Myrmekite in large lobes embays the microcline.

TABLE 44

MODAL ANALYSES OF GUILFORD QUARTZ MONZONITE

	1	2	3	4	5	6
Plagioclase	39.2	33.9	33.0	36.1	32.4	30.1
K feldspar	20.7	25.1	25.0	27.3	30.6	30.0
Quartz	30.7	28.8	34.2	28.5	30.4	32.4
Myrmekite	1.9	2.4	0.8	1.2	1.6	0.9
Biotite	3.9	5.2	3.2	3.9	2.5	1.6
Chlorite	0.1	0.1	—	—	—	—
Muscovite	3.2	4.4	3.5	2.7	2.4	4.9
Clinozoisite	0.2	tr	tr	0.2	tr	—
Apatite	—	0.1	0.2	0.1	0.1	0.1
Garnet	—	—	—	—	—	tr
Zircon	tr	tr	tr	tr	tr	tr
Magnetite	0.1	—	0.1	tr	tr	—
Total	100.0	100.0	100.0	100.0	100.0	100.0
Points	1,345	1,535	1,500	1,604	1,525	1,627

(1) RW-4. Quarry at Guilford.
(2) H42-1. Abandoned quarry, Little Patuxent River, ½ mile west of Guilford.
(3) H-9. Quarry at Atholton.
(4) H-10. Middle Patuxent River, 2 miles west-southwest of Guilford.
(5) H-8. Highway U.S. 29 at Atholton Manor.
(6) H-12. One mile west-southwest of Guilford.

Muscovite and biotite are present in about equal amounts. The muscovite is in large, well-formed flakes sharply bounded against the other minerals; quartz and microcline are sometimes molded around them. There is little doubt that muscovite as well as the biotite is a magmatic phase. Long ago Keyes (1895, p. 715–716) reached the same conclusion:

"The evidence at hand strongly indicates that in the Guilford granite a primary white mica or muscovite is present, contemporaneous in crystallization with the associated dark micas, or perhaps even somewhat older than the latter." "... to all appearances the two micas in the Guilford granite seem to have crystallized at nearly the same time."

Synthetic studies of muscovite confirm its stability at the liquidus temperature of granite above 1500 bars P_{H_2O} (Yoder and Eugster, 1955, p. 266–267). The Guilford also contains small muscovites and clinozoisite as an alteration product of plagioclase.

The texture is hypidiomorphic to allotriomorphic. There is little evidence of deformation, aside from the usual undulatory quartz. The magma crystallized with little disturbance, after the climax of regional deformation and metamorphism.

Pegmatite is associated with some of the intrusive bodies, forming intrusive dikes and thin stringers which appear to have healed early joints. The pegmatites approach ideal granite composition and were evidently formed from the residual Guilford magma.

A new chemical analysis of Guilford quartz monzonite is given (Table 37, col. 8). An older analysis is reported by Keyes (1895, p. 716).

BEAR ISLAND GRANITE

Granitic and pegmatitic dikes and small irregular-shaped intrusive bodies cut the Wissahickon Formation and the amphibolites at Bear Island, Sherwin Island, and

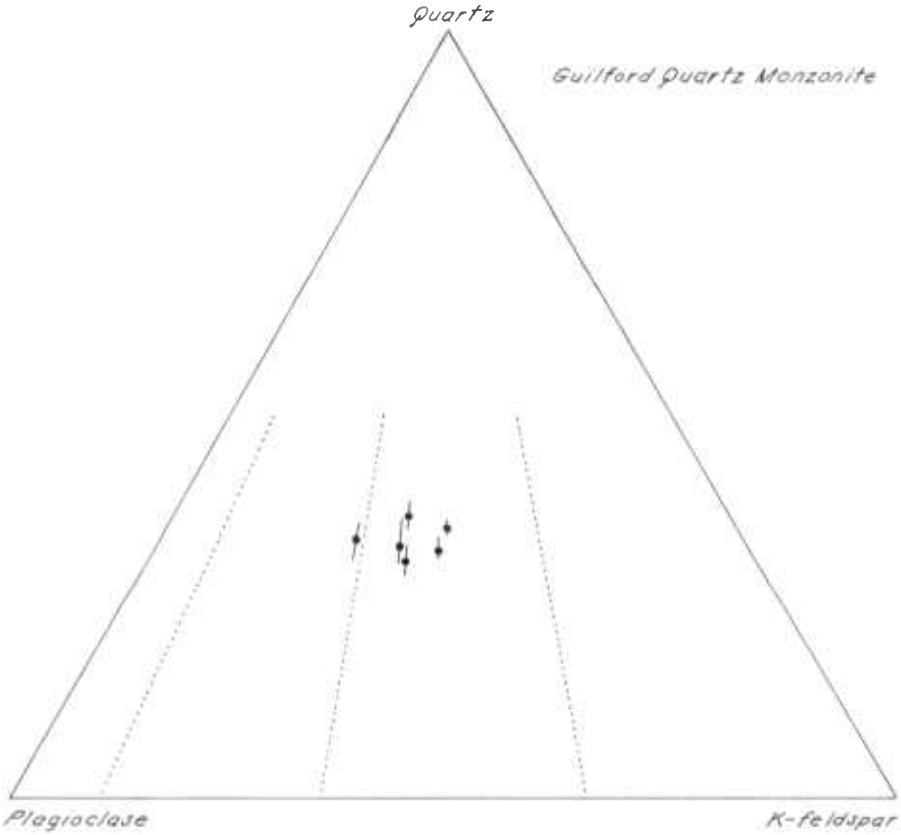


FIGURE 42. Modal composition of Guilford Quartz Monzonite. The vertical distance between ends of the line through each point gives percentage of biotite. Dotted lines denote classification (Fig. 37).

near-by areas along the Potomac River. The field and petrographic relations have been described by Cloos and Anderson (1950), Fisher (1963, p. 197–207), and by Reed and Jolly (1963, p. H9–H10).

Most of the granitic bodies are intrusive; they cut sharply across bedding and schistosity of the country rock and contain rotated xenoliths. A favored site for intrusion is along the margins and in shattered zones in the large bodies of relatively brittle amphibolite. There are also granite dikes in the schist which have evidently formed by replacement. The best example is a discordent dike of pegmatitic granite up to 12 feet thick at the south end of Bear Island. It contains numerous schist-metagraywacke skialiths whose internal structure is parallel to that of the wall rock.

Most of the granitic rocks are very leucocratic and might be called alaskite. Many of them locally coarsen into pegmatite. Their composition is variable, and different petrographic names have been given: Anderson (Cloos and Anderson, 1950, p. II-2) called them granodiorite, leucotonalite, and adamellite; Fisher (1963, p. 197) and Reed and Jolly (1963, p. H9) use the terms quartz diorite, granodiorite, and quartz monzonite. Here they are called granite because of their silicic composition (Table 37, col. 9) and because the plagioclase is chiefly albite.

Plagioclase (An_{2-15}), quartz, microcline, and muscovite are the main minerals and are generally accompanied by small amounts of epidote, biotite or chlorite, plus accessories.

TABLE 45

MODAL ANALYSES OF BEAR ISLAND GRANITE

	1	2	3	4	5	6	7	8	9
Albite	24.6	28.9	28.6	30.5	28.3	31.4	51.7	44.2	41.3
Microcline	15.4	17.3	14.7	16.6	7.4	13.2	—	—	—
Quartz	39.9	37.0	36.8	37.4	41.7	39.0	37.4	40.2	39.0
Myrmekite	—	—	—	.2	—	—	—	—	—
Muscovite I			13.5		10.7	5.0		8.2	5.1
	16.5	13.2		11.6			8.9		
Muscovite II			2.1		4.0	3.0		7.0	9.3
Biotite	.4	tr	1.1	1.4	6.0	4.5	—	tr	2.3
Chlorite	tr	.2	—	.3	—	.1	—	—	.1
Epidote	2.9	3.4	3.1	1.7	1.5	3.8	.7	.3	2.9
Allanite	—	tr	.1	—	—	tr	—	—	—
Garnet	.1	tr	—	tr	.4	—	—	—	—
Apatite	.2	—	—	tr	—	—	.1	.1	tr
Sphene	tr	—	tr	.3	tr	tr	—	tr	tr
Zircon	tr	tr	tr	tr	tr	tr	—	tr	—
Opaques	tr	—	—	tr	—	tr	—	tr	tr
Calcite	—	—	—	—	tr	—	1.2	—	—
Total	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0
Points	1248	1265	1310	1560	1258	1291	1452	1619	1296

Muscovite I: Coarse flakes of muscovite.
Muscovite II: Tiny muscovites replacing plagioclase.
(1) Granite (M407-2). Large dike cutting pelitic schist. Sherwin Island.
(2) Granite (M407-1). Large dike cutting pelitic schist. Sherwin Island.
(3) Granite (M222c). Replacement dike cutting pelitic schist. West end of Bear Island.
(4) Granite (M325-1). Small dike cutting pelitic schist. West end of Bear Island.
(5) Granite (M151). Small stock cutting amphibolite. West end of Bear Island.
(6) Granodiorite (M405-1). Small stock cutting amphibolite. West end of Bear Island.
(7) Albite granite (M430-1). Small dike cutting amphibolite. Middle of Bear Island.
(8) Albite granite (M432-1). Dike cutting pelitic schist. Widewater, Bear Island.
(9) Albite granite (M402-1). Dike cutting amphibolite. East end of Bear Island.

Modal analyses of some typical rocks are given in Table 45; all but the rock in column 3 are intrusive. The modes are shown graphically in Figure 43. The diagram shows that the quartz: feldspar ratio (including alteration products of feldspar) is rather constant near the granite minimum, but the albite: microcline ratio varies widely. Commonly the feldspar ratio varies greatly even in different parts of the same dike.

The intrusive rocks show hypidiomorphic to allotriomorphic textures. The replacement dikes are xenomorphic, and relict clastic quartz grains can occasionally still be recognized.

Fisher (1963, p. 207), and Reed and Jolly (1963, p. H10) call attention to the close spatial association of the granitic rocks with mica schists that show exudation phenomena and contain sillimanite. They postulate that some of the magma formed locally by anatexis during the height of regional metamorphism. According to Reed and Jolly (1963, p. H10):

"The transition of some of the granitic bodies into migmatites suggests that the granitic material is largely or entirely of local derivation and that the crosscutting intrusive bodies are composed of material that has been locally segregated and mobilized. The concentration of the granitic bodies in

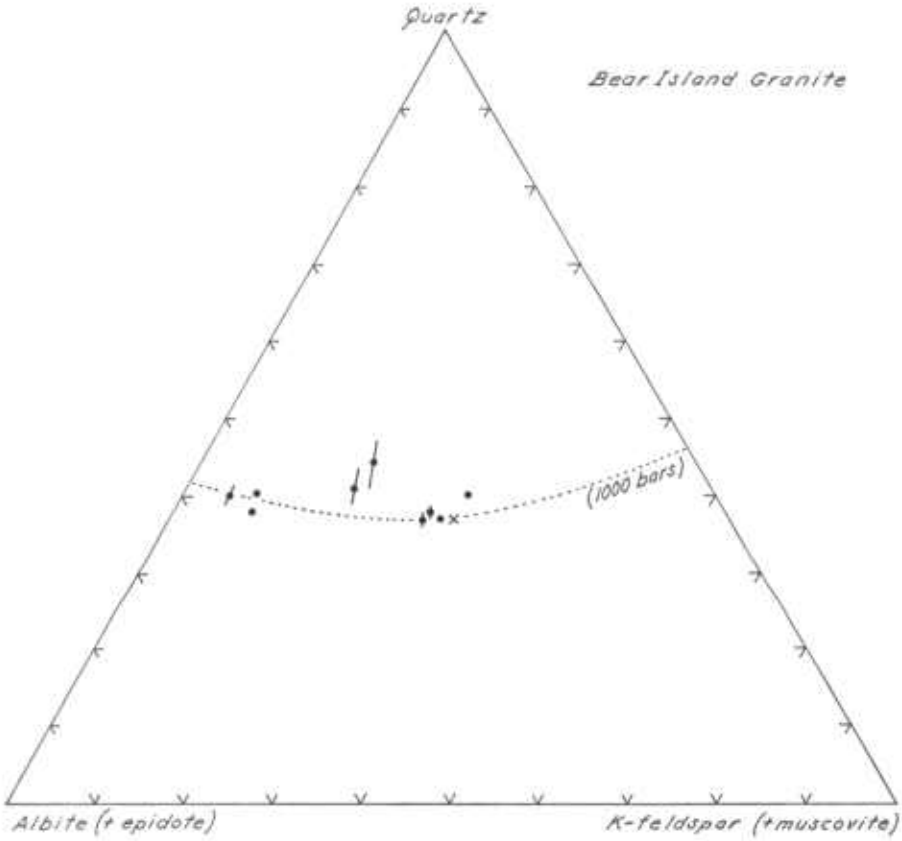


FIGURE 43. Modal composition of Bear Island Granite. The plotted feldspars include alteration products. The vertical distance between ends of the line through each point gives percentage of biotite and chlorite. The dashed line is the trace of the synthetic granite minimum at 1000 bars P_{H_2O} (Tuttle and Bowen, 1958, p. 75).

and near the zone of migmatization and sillimanite-grade regional metamorphism indicates that the granitic material was formed during this episode; but the sharp angular contacts of the discordant bodies and lack of cleavage in the granitic rocks suggest that mobilization and intrusion followed somewhat after the metamorphic climax."

The argument for local derivation of the intrusive granite hinges mainly on the migmatitic aspect of the Wissahickon country rocks. Three features give the appearance of migmatite:

(1) Widespread disruption of bedding and evidence of flowage in the schist and metasandstone.

(2) Quartzofeldspathic folia and veins in the schist.

(3) Replacement dikes and other feldspathized zones in the schist.

These features should not be taken to indicate ultrametamorphism and anatexis, however, for the following reasons:

(1) The disruption of bedding and flow structures in the Wissahickon formed by slumping and sliding of soft sediments (*see* p. 90–93).

(2) The quartzofeldspathic folia in the veined schists are composed of albite and quartz, and are not indicative of the highest grade of metamorphism or of partial fusion.

These features are also well-developed in the schists of moderate metamorphic grade, as for example at Rocky Islands, a mile upstream from Bear Island, where the schists were not metamorphosed above the staurolite zone. Similar veined schists have also been recorded in the greenschist facies (Turner, 1941).

(3) Granitization at Bear Island, characterized by the growth of metasomatic potassium feldspar in addition to albite, is developed only on a small scale and is restricted to the vicinity of the pegmatitic-granite intrusions. This suggests that the intrusions were the source of the feldspathizing fluids. In the Baltimore area large intrusive pegmatites have similar feldspathized aureoles where they invade mica schist (p. 185), but the magma came from much deeper down (p. 186–187).

The intrusive relations and igneous textures that most of the Bear Island granites display, and the lack of clearcut evidence for migmatization and anatexis, argue against local derivation of the magma. As an alternative hypothesis it is suggested that the granitic magma and pegmatitic fluids intruded and injected the Bear Island terrane, from a source at depth, well after the climax of regional metamorphism. The magma was rich in fluids, and their activity during crystallization brought about the marked local changes in texture and composition of the granites and their gradation into patches of pegmatite. The pegmatitic fluids also permeated the surrounding Wissahickon schist, forming the replacement dikes and diffuse feldspathized zones. The formation of fibrolitic sillimanite in the schist may also be related to the activity of these fluids, because: (1) the fibrolite is restricted to the vicinity of the granites; and (2) it appears to have grown after the schist had been partly retrograded, and its large aluminum-silicate porphyroblasts altered to shimmer aggregates (see p. 97–99). The discordant form of the granite intrusions and the lack of crosscutting cleavage, gneissic textures, and other evidence of strong deformation further suggest that most of the granite was emplaced after the metamorphic climax.

Nevertheless, the variability of the granitic rocks and their complex and sometimes conflicting relations with the wall rocks make it difficult to draw generalizations. In a few places there are small concordant bodies of gneissic granitic rock that appear to have folded with the schist; thus there may be more than one generation of granite. The firmest conclusion reached is that the problem deserves more study!

PEGMATITES OF THE BALTIMORE AREA

Granite pegmatite consisting chiefly of quartz, albite, microcline-perthite, and muscovite occur in sills and dikes up to 100 feet thick in an area close to the mantled gneiss domes in Howard and Baltimore Counties. They mostly occur on the upright flanks and across the roofs of the domes, in swarms that trend parallel to the strike of the host rocks. The main swarms are located: (1) along the northwest flank of the Towson dome near Glenarm; (2) along the northwest flank of the Baltimore dome near Jones Falls; (3) along the west flank of the Woodstock dome near Henryton and Marriottsville; (4) along the east flank and roof of the gneiss uplifts at Alberton and Bens Run; (5) along the west flank of the Clarksville dome; and (6) in the Baltimore Gabbro Complex between the Patapsco River and Guilford. Another short pegmatite swarm clusters over a small domal area of Wissahickon Formation, east of the Clarksville gneiss dome. Figure 44 shows their generalized distribution. For place names see Plate 7 and the County maps.

The pegmatites occur both as dikes and sills. Within short distances the sills break discordantly across the bedding or foliation to a new horizon. The pegmatites occur also

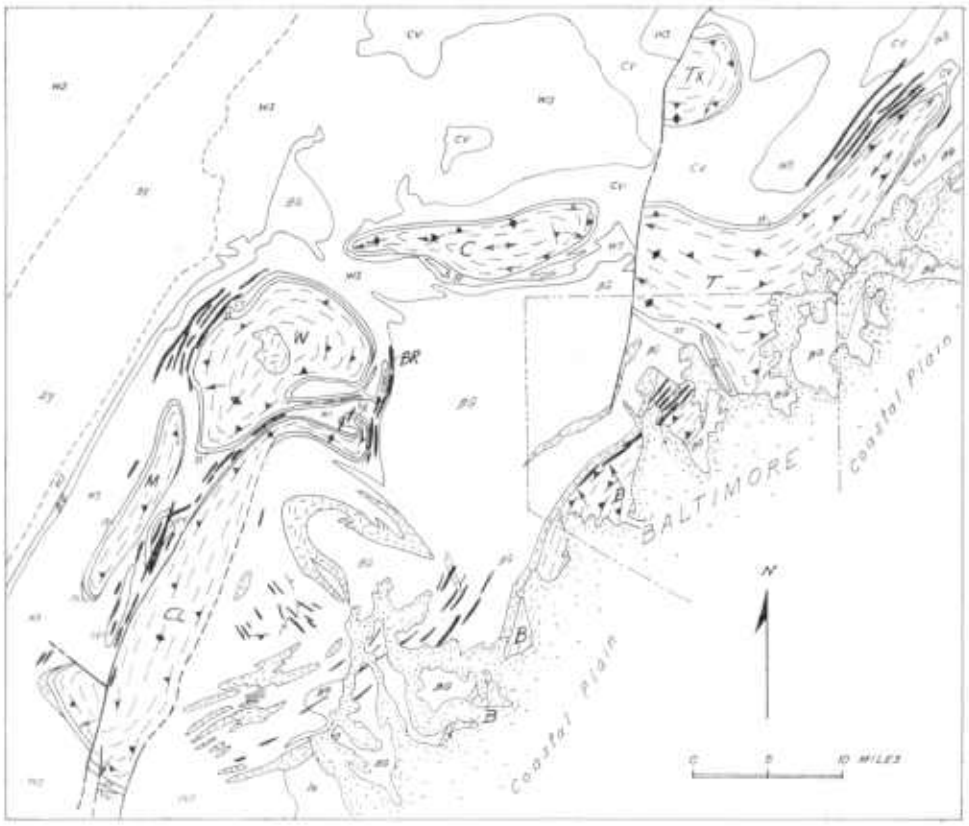


FIGURE 44. Generalized geological map showing pegmatite dike swarms and their structural setting. Pegmatites are in black; their size is exaggerated. Baltimore Gneiss domes shown are: B, Baltimore; BR, Bens Run; C, Chattolanee; CL, Clarksville; M, Marriottsville; T, Towson; Tx, Texas; and W, Woodstock. Granitic plutons shown by hatched pattern, and Coastal Plain sediments by stippling; all other units are in white: ST, Setters Fm.; Cv, Cockeysville Marble; Ws, Wissahickon Fm.; La, Laurel Fm.; Sy, Sykesville Fm.; BG, Baltimore Gabbro Complex. Refer to Plate 7 and figure 12 for geographic details. Geology modified after the geologic maps of Howard and Baltimore Counties; structure in the gneiss domes in Baltimore County is after Broedel (1937).

as isolated pods, lenses, and irregular masses. Their form varies with the host rock. In the massive marble, metagabbro, and quartzo-feldspathic gneiss the pegmatite occurs chiefly as dikes; in the well-bedded schists and quartzites sills and concordant pinch-and-swell structures are common. In the foliated mica schists diffuse, irregular masses of pegmatite and *lit-par-lit* stringers are found. One of the most favored sites for sills is along the contact between the Cockeysville Marble and adjacent formations. The contacts between the pegmatite and its host rock are generally sharp except in the mica schist. There the contacts are often diffuse, and large feldspar porphyroblasts cluster in the wall rocks.

The pegmatites do not appear to be concentrically zoned. Some are relatively fine-grained toward the margins, but others are coarse-textured right up to the walls. Coarsely porphyritic and giant-textured pegmatite is generally more abundant toward the center, but there are no massive quartz cores. Well-defined border and wall zones are lacking.

Many of the dikes are massive. They consist chiefly of medium- to very coarse-grained granitoid rock, which locally coarsens into giant-textured pegmatite. Part of the medium-grained rock is coarsely porphyritic, with megacrysts of microcline-perthite or blocks of graphic granite up to 10 inches long in a matrix of plagioclase, quartz, and muscovite. Books of muscovite up to 8 inches long occur locally. Despite the many textural variations these dikes are a single compositional and structural unit.

Other large pegmatites display a crude layered structure parallel to their walls. Alternating layers are of different texture, grain size, and composition. A large, gently dipping pegmatite dike along the Patapsco River below the dam, 1 mile downstream from Ilchester is a good example. Seen from the road, the gently northeast-dipping layering across the river looks like bedding in a metasediment. The layers range from about 6 inches to 5 feet thick and consist of aplite, medium- to coarse-grained granite, and giant-textured pegmatite. Some of the finer-grained layers contain scattered megacrysts or clusters of euhedral perthite up to 6 inches long. The composition of the various layers does not vary widely in this particular pegmatite, but it does in some of the others. A large dike 1 mile south of Henryton has layers that range from albite-quartz aplite to giant-textured perthite-quartz-muscovite pegmatite. The contacts between layers in most of the layered pegmatites is either an abrupt coarsening of grain size or a gradual coarsening over several inches; they are never sharply intrusive. The tendency for the layering to change upward from fine-grained albite-quartz rock near the footwall to coarse perthite and muscovite-rich pegmatite at the headwall (Orville, 1960, p. 1475) was not observed.

Although the large pegmatite bodies are not zoned some of them contain small zoned pods or dikelike bodies toward their center. A good example is in the massive 40-foot pegmatite sill at Gwynns Falls in Baltimore City, 200 yards below the Edmondson Avenue bridge. An 8-inch dike near the center of the sill has an outer zone of coarsely intergrown albite + quartz and an inner zone of very coarse perthite + quartz + muscovite. Crystals in the albite-quartz intergrowth are elongate perpendicular to the walls and evidently grew from the walls inward. The inner zone, with massive quartz filling the interstices between euhedral perthite megacrysts, crystallized last. Quartz from the inner zone fills small fractures which cut across the outer zone and the wall. The bulk composition of the dikelet appears to be close to that of the "granite-minimum" liquid in the synthetic granite system (Tuttle and Bowen, 1958, p. 75) but displaced toward the albite side. Thus the crystallization sequence, beginning inward from the walls with albite and quartz, is the one to be expected from a melt of this composition. Such features are common in other districts, and according to Cameron *et al.* (1949, p. 21) they "appear to have been formed as segregations, probably from isolated bodies of pegmatitic solutions that were trapped within solid pegmatite."

Late fracture fillings of massive quartz transect many of the large pegmatite bodies. Most of them are small, but some quartz lenses are a foot or more wide and at least 20 feet long. They generally stand at a high angle to the pegmatite contacts.

Most of the large pegmatites were probably emplaced by intrusion of highly fluid pegmatitic magma. The following criteria for intrusion were noted: (1) sharp discordant contacts of pegmatite against wall rock; (2) dilation offset of planar structures in the wall rock; (3) angular inclusions of wall rock that have been rotated out of their original orientation; (4) layered internal structure in crosscutting dikes, oriented parallel to the walls of the dike but at an angle to the structure of the transected host rock. Most of the pegmatites have poorly exposed contacts, so it is rarely possible to find more than two of these criteria at any one place.

Ramberg (1956) has suggested that in terrain where the composition of the pegmatites varies sympathetically with the composition of the host rock the pegmatite probably formed by replacement. Conversely, where the composition of the host rock appears to have little or no influence on the enclosed pegmatites an igneous origin seems to be indicated (Jahns, 1955, p. 1061–1062). The pegmatites which cluster around the gneiss domes in Howard and Baltimore Counties cut such diverse host rocks as gneiss, mica schist, marble, amphibolite, and quartzite with little change in their own composition; this relationship favors an igneous origin.

Some of the small pegmatite bodies probably formed by replacement. They have diffuse, irregular borders and grade out into wall rock that is peppered with feldspar porphyroblasts. Delicate, undisturbed septa of wall rock are preserved within some of the small pegmatites. Planar structures in the wall rock are not dilated where pegmatite cuts across them at a low angle. Small replacement pegmatites of this type are most common where the country rock is mica schist or gneiss.

The pegmatites have a simple mineralogy. Quartz, plagioclase, microcline, and muscovite are the chief minerals, and garnet, apatite, and tiny zircons are common minor accessories. Biotite occurs locally. Tourmaline, allanite, monazite, magnetite, and pyrite are also sporadic.

Most of the plagioclase is albite. Its average composition is about An_5, but it ranges from An_1 to about An_{16}. No antiperthitic plagioclase was observed, but some of the albites are partly replaced by microcline. Grid-twinned microcline occurs in small interstitial grains and in euhedral crystals a foot long. Crystals larger than about 3 mm are generally microperthitic, and those larger than about 1 cm are quite coarsely perthitic. Patch perthite and vein perthite are the common varieties. Some microcline is graphically intergrown with quartz.

Few of the pegmatites have unusual minerals. Near Scaggsville in Howard County, however, the pegmatites are more diversely mineralized than elsewhere. Perhaps they are younger pegmatites from the Guilford Quartz Monzonite. According to Ostrander (1940, p. 34) pegmatite of the Old Maryland mica mine northwest of Scaggsville con-

TABLE 46

MODAL ANALYSES OF FINE-GRAINED PHASES FROM PEGMATITE

	1	2	3
Albite	39.6	42.4	45.5
K feldspar	25.9	24.7	20.1
Quartz	32.6	30.6	32.3
Muscovite	.7	1.9	.9
Biotite	1.0	.1	1.1
Garnet	.1	.1	—
Epidote	—	—	.1
Allanite	—	—	tr
Magnetite	.1	.2	—
Zircon	tr	tr	tr
Total	100.0	100.0	100.0
Points	1570	1539	1615

(1) Aplite.
(2) Aplite.
(3) Medium-grained granite.
} From a large layered pegmatite at the Patapsco River, 1 mile below Ilchester, Howard County.

tains black tourmaline, micaceous hematite, ilmenite, pyrite, pyrolusite, pyrophyllite, and amethyst crystals, while pegmatite of the Ben Murphy mica mine 1 mile southwest of Scaggsville contains beryl, in translucent and opaque yellow green masses. William Melson (personal communication) also reports the uranium mineral autunite and some exceptionally fine specimens of gahnite (zinc-spinel) at the Ben Murphy mine.

At the Frost feldspar mine, half a mile south of Davis (near Woodstock), sphene, pyrite, pyrrhotite, amphiboles, pyroxene, vesuvianite, garnet, allanite, smoky quartz, muscovite, biotite, phlogopite, and dolomite occur where pegmatite cuts Cockeysville marble (Ostrander, 1940, p. 36). This is a contact mineralization caused by heating and metasomatism of the dolomitic marble.

The pegmatites commonly show evidence of moderate mechanical deformation. Plagioclase and microcline are locally bent and broken, the micas are twisted, and the quartz is undulatory. There has not been much recrystallization, however, and the primary textures are well preserved.

In summary, most of the large pegmatites appear to have been injected into the Glenarm Series and Baltimore Gabbro as fluid pegmatitic magma of alkali-granite composition. Some small masses of pegmatite also formed by replacement. The pegmatites were emplaced after the climax of regional metamorphism, because: (1) their igneous textures are not blurred by strong deformation and recrystallization; and (2) undeformed pegmatites cut sharply across metamorphosed plutonic rocks, such as the orthogneiss and the amphibolite (metagabbro) along the Patapsco River below Ilchester.

The pegmatite that forms the large dike and sill swarms was not derived from the intrusive granitic plutons, because: (1) the swarms are not satellitic to any pluton; and (2) the pegmatite is older (*see* section on radiogenic dating) than the plutons nearby that approach granite-minimum composition (Ellicott City Granodiorite; Woodstock and Guilford Quartz Monzonites)—the only ones that are likely to have differentiated much pegmatite. However, pegmatites do cluster closely around the gneiss domes (Fig. 44). This relationship is most evident when viewed on a regional scale: the only place in the Maryland Piedmont where swarms of large pegmatites occur is around the gneiss domes, whereas the granitic plutons are widespread. A close genetic relationship to the gneiss domes is thus suggested.

There is also a consistant structural relationship between the pegmatites and the gneiss domes, most of which are asymmetrical or overturned (Broedel, 1937). The pegmatite swarms strike parallel to the long axis of the domes and are localized along their *upright limb* and across their top. This is shown at the Towson, Baltimore, Woodstock, and Clarksville domes, which are overturned to the east and southeast and have pegmatite swarms along their west or northwest side (Fig. 44). The small Bens Run dome, overturned to the west, has large pegmatite dikes along the eastern flank. At the Phoenix dome near Belfast a large pegmatite dike cuts the lower Glenarm formations on the roof of the dome (Knopf and Jonas, 1925). Pegmatites also cut lower Glenarm rocks that cover buried protrusions of Baltimore Gneiss at: (1) the small domal areas that expose Cockeysville Marble between the Mayfield and Clarksville domes; and (2) the small dome in the Wissahickon Formation adjacent to the faulted eastern side of the Clarksville dome (Fig. 44). At the last locality Baltimore Gneiss is inferred to underlie the Wissahickon at shallow depth; it is a bulge of gneiss belonging to the downfaulted eastern side of the Clarksville dome.

Thus, a genetic relationship between the pegmatite swarms and the gneiss domes is indicated by: (1) their close spatial relation; (2) their parallel strike; and (3) localization of the pegmatites along the upright limbs and tops of the domes.

These relations are consistant with the theory that the pegmatites formed from palingenetic fluid that was expelled from the gneiss during the final stages of growth of the domes. During Paleozoic orogeny the basement was reactivated and rose as domal masses of mobilized gneiss and migmatite, arching its younger mantle of Glenarm strata (*see* Baltimore Gneiss, p. 52–53). The mantle, consisting of quartzite overlain by marble and schist, contained these domal intrusions at all but one place, where anatectic migmatite broke through discordantly (*i.e.*, Gunpowder Granite, p. 46–47). Small amounts of anatectic fluid, however, could have penetrated the mantle along zones of stress at many other places. The pegmatite swarms are situated along the upright limbs and tops of the overturned domes—sites where stretching was greatest as the mobilized gneiss punched obliquely upward. Here pegmatitic fluids that were strained from the migma might have risen along tension fractures developing along the backs and tops of the domes. Figure 45 shows this diagrammatically. As the highly fluid pegmatitic magma rose along systems of tension fractures it could have spread outward as sills, working along contacts and weak bedding planes in the Glenarm strata.

PLUTONIC HISTORY—SUMMARY AND INTERPRETATION

The Baltimore-Washington area had an active magmatic history that spanned the period of Early Paleozoic orogenesis. Gabbroic magma first invaded the Glenarm Series before the onset of folding and regional metamorphism, but emplacement and solidification of the gabbro and its differentiation products lasted into the period of strong orogeny. Granitic rocks ranging from quartz diorite to quartz monzonite were later intruded under synkinematic to late-kinematic conditions. Granitization of gneissic basement and rise of the mantled gneiss domes occurred at the height of regional deformation and metamorphism, and partly overlapped the two magmatic episodes.

The gabbro magma solidified mostly as deep plutonic masses, but locally it formed shallow diabasic sills. Some also broke through to the surface, erupting on the sea floor where it fragmented and was reworked into thin sheets of mafic aquagene tuff or palagonite. Slumping and sliding of the enclosing soft sediments subsequently deformed and dismembered some of the shallow sills and tuffaceous strata. They were deformed still more during regional metamorphism, and converted to greenschist and amphibolite.

The Baltimore Gabbro Complex, which extends into southeastern Pennsylvania and Delaware, is the largest mass of intrusive mafic rock. In the Baltimore area the complex was a semiconcordant sheet over 5000 feet thick, with ultramafic rocks near the base and layered to massive gabbros higher up. The magma evidently invaded flat-lying strata and began to differentiate gravitationally, but was caught up in regional deformation and metamorphism. Migmatization deeper in the crust culminated in mobilization of the gneissic basement, which rose as diapiric domes and anticlines beneath the mafic complex before the gabbro was fully solidified. The sheet-like complex was arched over two lines of rising gneiss domes and partly pinched in the deep structural depression between them. This movement outlasted solidification of the gabbro complex, and flowage and protoclastic deformation of gabbroic crystal mush passed gradually into shearing, granulation, and metamorphic recrystallization as the mass became increasingly rigid. Primary igneous textures and structures were distorted and obliterated, as much of the gabbro was converted to uralitic schlieren-gabbro and gneissic amphibolite.

Masses of ultramafic rock that had differentiated from the gabbro were squeezed from their original positions in the complex, invading the gabbro and neighboring Wissahickon Formation as quasi-solid intrusions. Some of the serpentinization and steatitization of the ultramafic rocks probably took place during this movement.

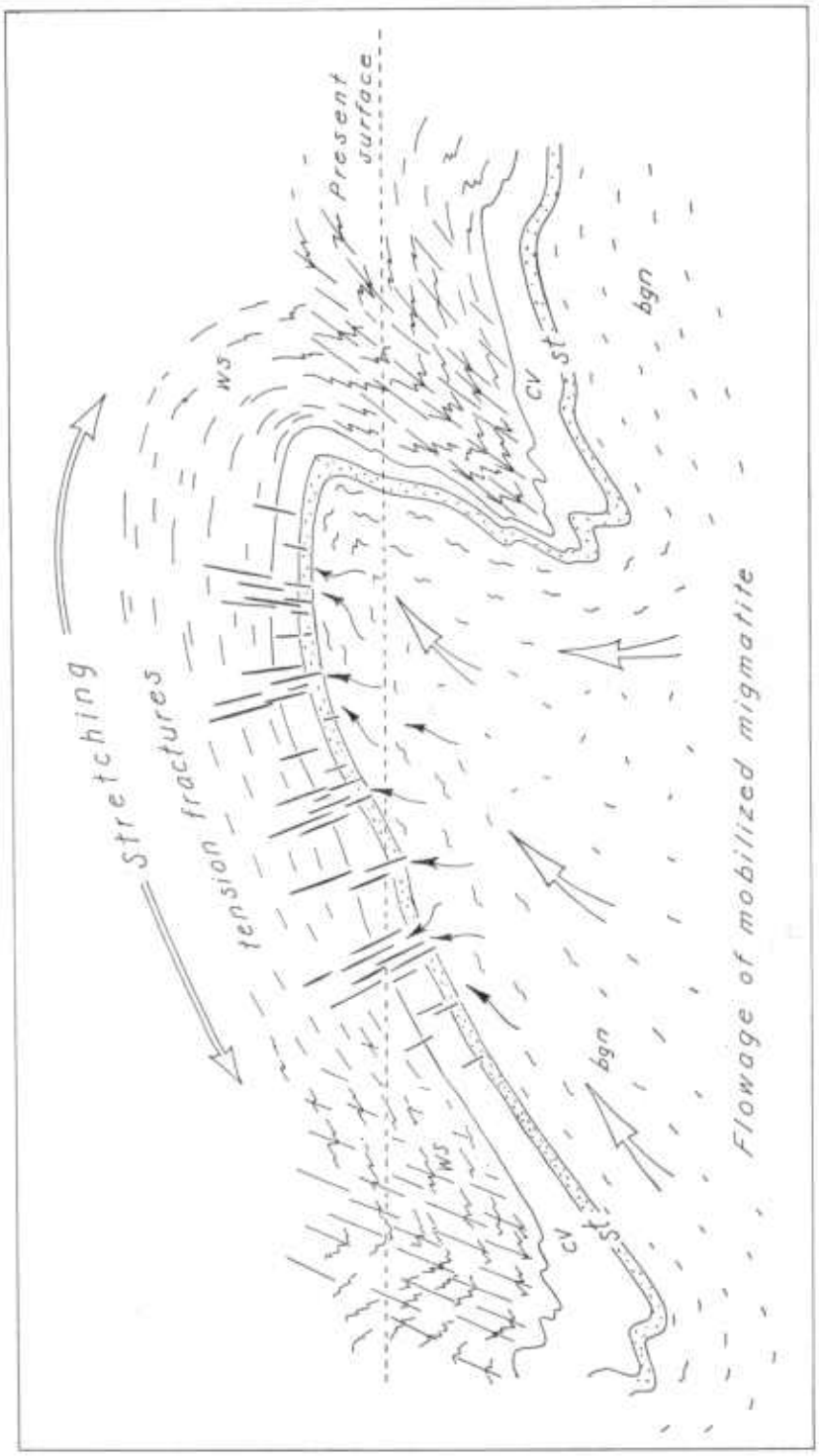


FIGURE 45. Schematic cross section, showing postulated origin of the pegmatites. The rock units are: Baltimore Gneiss, bgn; Setters Formation, st; Cockeysville Marble, cv; and Wissahickon Formation, ws. Open arrows in the gneiss portray flowage of mobilized gneiss and migmatite. Black arrows show anatectic magma and fluids rising from the gneiss and entering tension fractures (heavy black slashes) in the overlying metasedimentary rocks. The tension fractures developed where the metasedimentary mantle was stretched across the back of the rising gneiss dome.

Isolated slivers and large pod-like masses of serpentinite and soapstone situated in western Howard County and central Montgomery County have no visible connection with any parent gabbroic mass. Many of them, however, contain large schlieren-like masses of highly altered gabbroic rock.

Sills and small discordant masses of leuco-diorite and leuco-quartz diorite (Relay Quartz Diorite) intrude the Baltimore Gabbro Complex, and are probably genetically related to it. This is suggested by their close spatial relation to the complex, and by their petrographic similarity to the net veins of leucogabbro and diorite that were filter-pressed from the gabbro. Silicification and albitization took place during solidification of the leuco-quartz diorite, transforming parts of it to albite granite. All these leucocratic granitic and dioritic rocks were metamorphosed with the gabbro, and many of them are crushed and recrystallized to gneiss or locally smeared out into phyllonite.

The numerous other granitic intrusives in the Baltimore-Washington area are distinctive petrographically and chemically from the highly leucocratic Relay intrusives, and they are not spatially associated with the gabbro. They range from dark quartz diorite to light-colored quartz monzonite and pegmatitic granite. Intrusions of the dark hornblendic quartz diorite were earliest, and the successively younger plutons were progressively more leucocratic and contain more K-feldspar. Emplacement and solidification of these plutons began during the height of regional deformation and metamorphism, continued through a period of waning metamorphic intensity, and concluded under fairly static conditions. The synkinematic quartz diorite and granodiorite occur chiefly as concordant intrusions. Vestiges of primary magmatic texture survive in some of them, but many have been thoroughly granulated and recrystallized to orthogneiss, or even to quartzofeldspathic schist. The late-kinematic plutons are partly discordant, and consist of potassic granodiorite and quartz monzonite. They are more weakly foliated or massive, and show little evidence of deformation except locally along their borders. These late-kinematic rocks are cut by small dikes of granite pegmatite, derived from their own residual fluids.

Two co-magmatic groups may be distinquished: an early gabbroic series and a later granitic series. The former includes the Baltimore gabbro and related ultramafic rocks, and the small bodies of diorite, leuco-quartz diorite, and albite granite that are collectively called Relay Quartz Diorite. Comprising the granitic series are most of the other intrusive masses of quartz diorite, granodiorite, and quartz monzonite in the Baltimore-Washington area.

Rocks of the gabbroic series stem from the large volumes of gabbroic magma that invaded the Glenarm Series during the earliest stages of orogeny. The ultramafic rocks originally formed as crystal accumulates differentiated from the gabbro but orogenic deformation squeezed many of them from their place of origin, in some cases severing their connection with the gabbro complex. Small amounts of residual dioritic and quartz dioritic magma were also squeezed from the gabbro and injected other parts of the complex as small intrusive bodies that crystallized under synkinematic conditions. Quartz and albite metasomatism accompanied the final stages of crystallization.

Thayer (1963) calls attention to a distinctive magmatic association in which peridotite and gabbro predominate, soda-rich diorite and quartz diorite are subordinate, and late-stage sodic metasomatic alteration is characteristic. This association evidently represents a distinctive line of magmatic descent, which he calls the "alpine mafic magma stem". The gabbroic series at Baltimore appears to be a good example of this association, and has many features in common with other examples from Oregon and Cyprus, given by Thayer.

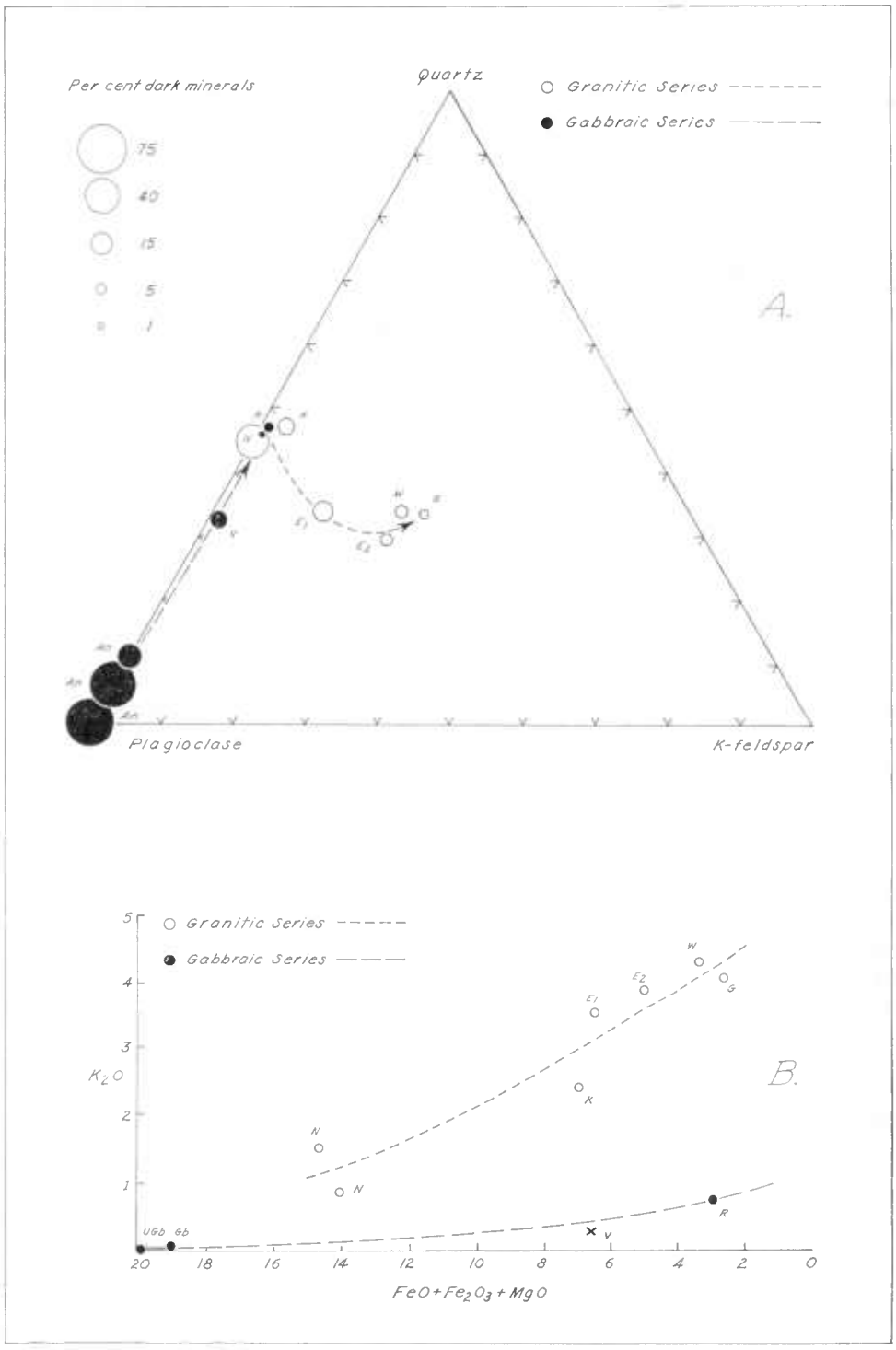

Per cent dark minerals
75
40
15
5
1
Quartz
Granitic Series
Gabbraic Series
A.
Plagioclase
K-feldspar
Granitic Series
Gabbraic Series
B.
K_2O
$FeO+Fe_2O_3+MgO$
UGb
Gb
N
K
W
G
E1
E2
R
V

FIG. 46

Rocks of the granitic series progress chronologically from dark hornblendic quartz diorites to granodiorites and quartz monzonites that are progressively lower in ferromagnesian constituents and richer in alkalies, especially potassium. This progression suggests differentiation from some common magmatic source. It seems unlikely, however, that Baltimore gabbro was the parent magma, because (1) the granitic plutons are unrelated spatially to the gabbro complex; (2) the exposed volume of granitic rock greatly exceeds the amount that could have been fractionated from the gabbro; and (3) the gabbroic and granitic series have divergent differentiation trends.

Data bearing on the differentiation trends are plotted in figure 46. The triangular diagram (fig. 46A) shows average modal composition for each of the granitic intrusives and for some typical rocks of the Baltimore Gabbro Complex. The long-dashed line connects rocks of the gabbroic series (including Relay quartz diorite and albite granite), and the short-dashed line connects members of the granitic series. Arrowheads point the direction of progressively younger rocks of each series. Although both series show a progressive decrease in ferromagnesian minerals (shown proportional to the size of the dots) their other trends differ significantly. Late members of the gabbroic series show a strong increase in quartz but hardly any enrichment in potassium feldspar. In the granitic series this trend is reversed: there is strong increase in potassium feldspar but a slight impoverishment in quartz. Not shown on the diagram is the change in plagioclase composition: plagioclase in rocks of the gabbroic series changes progressively from about An_{80} to An_0, whereas the granitic series plagioclases change from about An_{50} to An_{20}. This reflects more pronounced sodium enrichment for late members of the gabbroic series.

Meaningful chemical variation diagrams for the two plutonic series cannot be drawn for lack of sufficient data. One difference in their chemical trends is so pronounced, however, that even the existing data bring it out clearly. This is the potassium trend, shown in figure 46B. The variation diagram, in which K_2O is plotted against total ferromagnesian oxides ($FeO + Fe_2O_3 + MgO$), shows the strong potassium enrichment in the later, more leucocratic members of the granitic series, and the almost negligible potassium increase in corresponding members of the gabbroic series. It also shows that the granitic series as a whole is more potassic than even the most leucocratic member of the gabbroic series.

To sum up, the gabbroic series was pre-kinematic to synkinematic, and shows a trend toward strong enrichment of its late members in silica and sodium. The granitic series is synkinematic to late-kinematic, and the trend is toward potassium enrichment.

FIGURE 46. Plots of modal and chemical data for the plutonic rocks.

A. Diagram showing average modal composition of granitic and some of the gabbroic rocks: AM, metagabbro; v, quartz dioritic vein in metagabbro; R, Relay quartz diorite and albite granite (2 points); N, Norbeck quartz diorite; K, Kensington quartz diorite; E_1, Ellicott City granodiorite; E_2, Ellicott City porphyritic quartz monzonite; W, Woodstock quartz monzonite; G, Guilford quartz monzonite. Line of long dashes shows trend of rocks genetically related to the Baltimore Gabbro Complex; line of short dashes show trend for granitic plutons not related to the gabbro. Arrow indicates direction of progressively younger rocks in each series. Volume of the dots is proportional to the percentage of mafic minerals. Data from Tables 33, 35, 36, 38, 39, 40, 41, 43, 44.

B. Chemical variation diagram, with K_2O plotted against total ferromagnesian constituents, comparing the trends of the granitic and gabbroic series. Values for the granitic rocks are from Table 37. Gb is gabbro (average of 23 powders, from Williams, 1886, p. 39); UGb is uralite gabbro (average of 19 powders, from Williams, 1886, p. 37). X_V is the approximate composition of a quartz dioritic vein in gabbro, calculated from the mode (Table 35, col. 7).

Granitization was not an important process at the present exposed level except within the mantled gneiss domes. The granitic plutons in the Glenarm Series were magmatic bodies, intruded into rocks of medium metamorphic grade (garnet, staurolite, and kyanite zones). Parts of some of the newly solidified plutons were incipiently granitized by their own residual fluids rising from deeper parts of the same mass, but the Glenarm metasediments were mostly uneffected. Within the gneiss domes, however, there is widespread evidence of microcline granitization, migmatization, and plastic flowage. The domes rose where softened, migmatized basement gneiss was tectonically squeezed up into its Glenarm cover, much like diapiric intrusions. Anatectic fluids expelled from the migmatitic gneiss were evidently responsible for the swarms of granite pegmatites that cluster around the domes. Despite evidence of gneiss flowage and formation of pegmatites, however, it is doubtful if more than a small fraction of the gneiss ever became liquid at the level now exposed. This is shown by its textures and internal structures, and by the fact that the gneiss so rarely broke across its sedimentary mantle.

Thus, two wholly different styles of granitic plutonism were active in the Baltimore area during Early Paleozoic orogeny: (1) Basement gneiss was pervaded by potassic fluids, migmatized, and softened sufficiently to rise as elongate domical intrusions of granodioritic and quartz monzonitic gneiss. Anatexis at one locality gave rise to the cross-cutting, diapiric Gunpowder Granite, but elsewhere the gneiss moved chiefly by plastic flowage. (2) The granitic plutons in the Glenarm Series were emplaced as magma, first as quartz diorite and later as granodiorite and quartz monzonite.

These two types of granite-forming activity were going on simultaneously during part of the time. The migmatitic basement (Baltimore Gneiss) was reactivated and began to rise as gneiss domes while the Baltimore Gabbro Complex was still crystallizing as a flat sheet. The palingenetic activity and doming of the gneiss was very lengthy, and outlasted the solidification and metamorphism of the gabbro and its cortege of leucogranitic intrusives; it finally ended at about the time when the large pegmatite swarms developed along tension zones across the backs of the overturned domes (Fig. 45). These pegmatites, derived from fluids stewed from the gneiss, mark the end of the period of strong deformation and metamorphism because their igneous textures and structures are undisturbed except for local mild cataclasis. The early, quartz dioritic plutons also participated in the strong deformation and metamorphism, as shown by the fact that they are largely recrystallized to gneiss; thus, their emplacement overlaps the period when the gneiss domes were active.

The relative timing of these events, deduced from structural evidence, is substantiated by radiogenic dating, described farther on.

The origin of the granitic and gabbroic magmas, and their relation to the palingenetic activity in the Baltimore Gneiss, is still a matter of conjecture. Nevertheless the following scheme, which accord with the preceding facts and interpretations, is offered as a working hypothesis.

The gabbroic magma presumably originated near the mantle, and worked its way upward into the Glenarm Series during an early stage of slowly accelerating orogeny. Where large masses collected in deep chambers and crystallized slowly magmatic differentiation led to ultramafic rocks formed as crystal accumulates, and to small amounts of Na-rich dioritic and granitic rocks which evolved along the liquid line of descent (Bowen, 1928). Simple stratiform complexes might have resulted, had not strong orogenic deformation and regional metamorphism intervened. As it was, the siliceous residual liquids as well as mobile masses of ultramafic rock were tectonically squeezed

from their place of origin, and much evidence of original gravitational differentiation has been distorted or lost.

With the onset of regional metamorphism temperatures deep in the crust approached the minimum melting range of the sialic gneisses. The most mobile chemical elements, especially potassium, were sweated from the gneiss, some remaining behind in veinitic migmatite but the rest migrating upward into the gneisses at higher levels, as demonstrated in the Grenville Series (Engel and Engel, 1958). Thus the metasomatic influx of potassium into the Baltimore Gneiss may have resulted from degranitization at depth. Mobilization of Baltimore gneiss and growth of the mantled gneiss domes was aided by the activity of these potassic fluids, as well as by rising temperature and the beginning of strong regional compression. The favored occurrence of metasomatic microcline augen gneisses near the gneiss dome contacts suggests that the quartzite and marble at the base of the Glenarm Series formed a barrier to further upward migration of potassium.

Meanwhile, continued heating in the degranitized zone beneath led to fusion of the deep crustal gneisses. This anatectic magma, formed from rock that had already sweated out much of its potassium, had the composition of quartz diorite. Batches of the quartz diorite magma—early members of the granitic magma series—were squeezed upward along zones of weakness, invading the Baltimore Gneiss and Glenarm Series during the later stages of strong regional deformation and metamorphism. Subsequent batches of granitic magma, working upward much more slowly after regional compression had relaxed, had the opportunity to differentiate to granodiorite and quartz monzonite. Evidence from the Ellicott City pluton suggests that this differentiation involved a gradual upward transfer of K and Si toward the higher portions of deep magma columns, perhaps by thermodiffusion and vapor transport.

RADIOGENIC DATES AND SOME GEOLOGIC AGE PROBLEMS

INTRODUCTION

The development of radiogenic mineral dating has infused new life into the difficult task of deciphering geologic history in the Piedmont crystalline rocks. In the Maryland Piedmont geologists are benefiting from radiogenic age studies in the Baltimore-Washington area by G. R. Tilton, G. L. Davis, G. W. Wetherill, and their colleagues at the Carnegie Institution of Washington. This work has not solved all geologic age problems in the area—indeed, it has raised some new ones—but a long stride forward has been taken, and the way has been shown toward their eventual solution.

The published radiogenic ages for the Maryland Piedmont rocks are compiled in this chapter, and their bearing on the following four age problems is discussed: (1) the age of the Baltimore Gneiss; (2) the time of emplacement of the plutonic rocks; (3) the time of regional metamorphism; and (4) the age of the Glenarm Series. The reader is urged to consult the original references for details of the isotopic analyses and for the interpretations given by their authors.

This chapter has been improved by the constructive criticism of George Tilton. Doctor Tilton has also made available unpublished dates given in Tables 47 and 49.

BALTIMORE GNEISS

Baltimore Gneiss, consisting of strongly transformed gneisses and migmatite, crops out west and north of Baltimore in anticlinal domes mantled by metasedimentary rocks of the Glenarm Series (Fig. 12). Although the contact between the gneiss and Glenarm

mantle appears everywhere to be concordant it was once a surface of major unconformity (p. 56–57). Other Baltimore Gneiss, composed chiefly of layered granulite and amphibolite derived from stratified tuffaceous rocks, crops out in Baltimore City and eastern Howard County in a narrow strip overlapped by Coastal Plain sediments. This rock is bordered by concordant mica schist and intrusive Relay Quartz Diorite, and its stratigraphic relation to the other gneiss and basal Glenarm formations is uncertain; possibly it is younger (p. 30, 53–54). The published radiogenic age data are all from the gneiss in the mantled domes.

Radiogenic mineral ages from the Baltimore Gneiss demonstrate its Precambrian age and support the long-standing conclusion of geologists that it represents crystalline basement upon which the Glenarm Series was deposited. Other mineral ages from the gneiss provide evidence of its participation in the Paleozoic metamorphism. The data and original conclusions are given by Tilton *et al.* (1958; 1959; 1960) and Wasserburg *et al.* (1957). Their mineral ages from the gneiss in the Baltimore area are summarized in Table 47.

Zircon and microcline from the gneiss give isotopic ages ranging from 1190 to 940 m.y. (Table 47). Most of the dates cluster between 1000 and 1100 m.y., evidently recording a major metamorphic or plutonic event. Zircon dates, however, are open to criticism that the zircons may be a mixture of relict detrital grains, whose apparent age has nothing to do with the age of the rock. However the zircon ages from the Baltimore Gneiss are in good agreement with the 1-billion year zircon and other mineral ages from basement rocks throughout the Appalachians and Grenville Province (Tilton, *et al.*, 1960). Such a consistent age from many different rocks over such a vast area could hardly result from random mixtures or relict detrital grains; therefore it probably represents a period of crystallization or strong recrystallization (Tilton *et al.*, 1960).

Rubidium-strontium ages for microcline from three localities in the Baltimore Gneiss also average about 1100 million years (Table 47) and confirm the zircon dates. A relict detrital origin for these feldspars is clearly excluded, for they are in delicate interlocking contact with the other metamorphic minerals (Tilton *et al.*, 1958, Fig. 2). A possible criticism for the microcline ages, however, is that young microclines might give older apparent ages if an older generation of radiogenic strontium were taken up by cryptoperthitic plagioclase. The dated microclines were X-rayed, however, and found not to be perthitic. Moreover, common Sr in the microclines that have low Rb was found to have normal isotopic composition (G. W. Tilton, personal communication).

Therefore the zircon and Sr/Rb microcline ages show conclusively that the Baltimore Gneiss was a crystalline rock 1 billion years ago, when it participated in metamorphism and orogeny that affected a large portion of eastern North America (Tilton *et al.*, 1960). Subsequently it was uncovered by erosion, deeply buried beneath a thick cover of Glenarm sediments, and then metamorphosed once more during Paleozoic orogeny. At that time the gneiss was remigmatized and mobilized, punching upward as plastic masses that domed but rarely breached its metasedimentary mantle. The mica ages and the A/K feldspar ages (Table 47) stem from this Paleozoic activity, although they do not accurately date the doming or the strongest metamorphism.

It is difficult to separate the effects of Paleozoic migmatization and flowage in the gneiss from those which occurred about 1 billion years ago. Proof of Precambrian migmatization is as follows: The gneiss is so thoroughly granitized and distorted by plastic flowage that nearly all evidence of its origin has been effaced, whereas the overlying Glenarm metasedimentary rocks clearly preserve their original character. The

TABLE 47

MINERAL AGES FROM BALTIMORE GNEISS

GNEISS DOME		ROCK	MINERAL	AGE, IN MILLION YEARS						REF.
				Sr^{87}/Rb^{87}	A^{40}/K^{40}	Pb^{206}/U^{238}	Pb^{207}/U^{235}	Pb^{207}/Pb^{206}	Pb^{208}/Th^{232}	
Phoenix dome	(1)	Migmatitic biotite gneiss	Zircon			960	1020	1120	1100	A, B
	(1)		Microcline	1190						A, B
	(1)		Biotite	310	390					B
Woodstock dome	(2)	Veined biotite gneiss	Biotite	305	410					B
Towson dome	(3)	Biotite gneiss	Zircon			1040	1070	1120	940	A, B
	(3)		Microcline		309					A
	(3)		Biotite	305	339					A
Towson dome	(4)	Veined biotite gneiss	Biotite		319*					C
Towson dome	(5)	Microcline augen-gneiss	Microcline	1100						B
	(5)		Biotite	315	290					B
Towson dome	(6)	Microcline augen-gneiss	Microcline	1010						D

Locations:

(1) Piney Creek, along U.S. Highway 111, 0.5 mile southwest of Verona, Hereford quadrangle.
(2) Railroad cut, B & O Railroad, 300 yards west of Woodstock, Ellicott City quadrangle.
(3) Charles Street Extended, 200 yards northwest of Sheppard-Pratt Hospital gatehouse, boundary between Towson and Cockeysville quadrangles.
(4) Roadcut, Charles Street Ave., Towson, between Malvern and Chesapeake Aves,. Towson quadrangle.
(5) Cub Hill Road, 200 yards south of Cromwell Bridge, Towson quadrangle.
(6) Long Green Creek, 200 yards north of Hartley Mill (0.9 mile east of Glenarm), White Marsh quadrangle.

References:

(A) Tilton, Wetherill, Davis, and Hopson (1958, p. 1471).
(B) Tilton, Davis, Wetherill, Aldrich, and Jäger (1959, p. 171).
(C) Wasserburg, Pettijohn, and Lipson (1957, p. 356).
(D) G. R. Tilton, personal communication, December 10, 1959.

* Calculated with a different decay constant; subtract 15 m.y. to compare with the other A/K dates given here.

gneiss is so thoroughly transformed right up to the contact that one must conclude that it took place before the Glenarm Series was deposited. This is substantiated by the 1-billion year Sr/Rb microcline ages from the gneiss at three localities (Table 47). Here the migmatite and augen gneiss are Precambrian relics.

Evidence that the Baltimore Gneiss became mobile and locally palingenetic after Glenarm deposition is as follows:

(1) The domical form and internal structures of several of the mantled gneiss masses suggests upward movement of mobile material, rather than anticlines resulting from compressional folding of hard crystalline rock (p. 50–52).

(2) The lower part of the Setters Formation is granitized at several places at the

Towson dome and the Phoenix dome (p. 59). Attempts by G. R. Tilton and G. L. Davis to date microcline from some of the replacement dikes was unsuccessful, due to its low ratio of radiogenic to common strontium (personal communication).

(3) Partial anatexis of Baltimore Gneiss in the Towson dome resulted in a diapiric intrusion of migmatitic granite (Gunpowder Granite) that cuts across the Glenarm mantle (p. 46–47). Microcline from the granite gives a Sr/Rb age of 580 ± 130 m.y. The large uncertainty is due to the low ratio of radiogenic to common Sr.

Another post-Glenarm feature that is believed to reflect palingenesis in the gneiss are the 440 million-year-old pegmatites that occur along the flanks of the domes (p. 186–187). These are discussed farther on.

THE PLUTONIC ROCKS

Widespread plutonic activity along the eastern part of the Maryland Piedmont preceded, accompanied, and outlasted the strong early Paleozoic metamorphism. Two series of co-magmatic rocks are recognized: a gabbroic series and a later granitic series (p. 189–191). Mineral ages are published for rocks from the granitic series (Table 49) but none are available from the gabbroic series. In the Baltimore area there are also several swarms of large pegmatite dikes and sills, which appear to have a closer genetic relation to the gneiss domes than to either series of igneous intrusions. Mineral ages from these pegmatites are listed in Table 48.

The difficulty of dating plutonic intrusions by their radiogenic mineral ages, especially in metamorphic terrain, is well known. In the ideal case liquid magma is emplaced, solidifies and cools rapidly, and then remains undisturbed to the present time. The age of the pluton is then the time elapsed since the magma crystallized; the age of its minerals by the various isotopic methods should be in agreement. However, such a simple model rarely holds for granitic intrusions emplaced at relatively great depth under orogenic

TABLE 48

MINERAL AGES FROM PEGMATITES

Dike Swarms in the Baltimore Gneiss Dome Area

MINERAL	AGE, IN MILLION YEARS		REFERENCE
	$\frac{Sr^{87}}{Rb^{87}}$	$\frac{A^{40}}{K^{40}}$	
(1) Muscovite	450	360	Tilton *et al.* (1959, p. 172)
Muscovite		372	Wasserburg *et al.* (1957, p. 356)
(2) Microcline	440	280	Tilton *et al.* (1959, p. 172)
Microcline		367	Wasserburg *et al.* (1957, p. 356)
(3) Microcline	440	320	Tilton *et al.* (1959, p. 172)
Microcline		347	Wasserburg, *et al.* (1957, p. 356)
(4) Microcline	440	265	Tilton *et al.* (1959, p. 172)
Microcline		344	Wasserburg *et al.* (1957, p. 356)
(5) Muscovite	400	280	Tilton *et al.* (1959, p. 172)

Locations:

(1) Baltimore and Ohio Railroad tracks, 0.3 mile east of Daniels, Ellicott City quadrangle.
(2) Near Henryton Road, 0.7 mile east of Henryton, Sykesville, quadrangle.
(3) Glenarm Road, 0.4 mile north of Cromwell Bridge, Towson quadrangle.
(4) Notch Cliff crossing, old Maryland & Pennsylvania Railroad, 1½ miles southwest of Glenarm, Towson quadrangle.
(5) Falls Road at Clipper Mill Road, Baltimore City.

TABLE 49

MINERAL AGES FROM GRANITIC ROCKS

LOC.	PLUTON	ROCK	MINERAL	AGE, IN MILLION YEARS						REF.
				$\frac{Sr^{87}}{Rb^{87}}$	$\frac{A^{40}}{K^{40}}$	$\frac{Pb^{206}}{U^{238}}$	$\frac{Pb^{207}}{U^{235}}$	$\frac{Pb^{207}}{Pb^{206}}$	$\frac{Pb^{206}}{Th^{232}}$	
1	Guilford	Muscovite-biotite quartz monzonite	Biotite	295						B
			Muscovite	335						B
1		Late pegmatite dike	Microcline	375						D
			Muscovite	355						B
2		Late pegmatite dike	Microcline	370						B
3	Woodstock	Biotite quartz monzonite	Zircon			330	340	415	315	B
			Biotite	310	295					B
4	Ellicott City	Biotite granodiorite	Zircon			355	370	450	310	B
			Biotite	290	315					B
			Hornblende		300 ±10					H
5		Late pegmatite dike	Muscovite	360						B
6	Kensington	Biotite quartz diorite gneiss (A)	Zircon			370	395	550		E
			Biotite	305	380					E
6		Biotite quartz diorite gneiss (B)	Zircon			400	420	510	350	E
			Biotite		350					E
7	Norbeck	Biotite-hornblende quartz diorite gneiss	Zircon			495	510	570 ±50	485	F
			Hornblende		315 ±15					H
8	Gunpowder	Muscovite-biotite quartz monzonite gneiss	Microcline	580 ±130						G
9	Bear Island Complex		Zircon			568	(lead-alpha age)			J

Location:
(1) Middle Patuxent River, 1½ miles southwest of Guilford, Savage quadrangle.
(2) Quarry along U. S. 29 at Atholton, Savage quadrangle.
(3) Sylvan Dell quarry at Granite, Ellicott City quadrangle.
(4) Quarry along River Road, 0.3 mile southeast bridge at Ellicott City, Ellicott City quadrangle.
(5) Thistle, Ellicott City quadrangle.
(6) Broad Branch Road, Washington, D. C.
(7) Green Wood Knolls in Wheaton, Kensington quadrangle.
(8) Gunpowder Falls near Harford Road, Towson quadrangle.
(9) Bear Island, Falls Church quadrangle, Maryland-Virginia.

References:
(B) Tilton, Davis, Wetherill, Aldrich, and Jäger (1959, p. 171–172).
(D) G. R. Tilton, personal communication, Dec. 10, 1959.
(E) Davis, Tilton, Aldrich, and Wetherill (1958, p. 178).
(F) Davis, Tilton, Aldrich, Wetherill, and Bass (1960, p. 158).
(G) G. R. Tilton, personal communication, July 6, 1960.
(H) Hart (1961, p. 2999).
(J) Jaffe, Gottfried, Waring, and Worthing (1959, p. 115).

conditions. For the synorogenic plutons in Maryland some of the complicating factors are: (1) The magma probably was anatectic, and may have carried older unfused relict minerals. (2) Magmatic crystallization probably merged into protoclastic deformation and then into metamorphic recrystallization, as the newly solidified masses became

involved in the regional metamorphism. Some of the gneissic quartz diorite plutons, in particular, probably moved as a protoclastic mush for a long time, and then later were kneaded and recrystallized in an essentially solid state, although with sufficient "pore fluid" to still give the mass some mobility. The time when such a mass finally ceased to intrude its wall rock might postdate the crystallization of its original melt fraction by millions or perhaps tends of millions of years. (3) Heat loss from the solidified mass was very slow. Recrystallization or annealing of strained minerals, or loss of radiogenic daughter elements by diffusion probably continued over a very extended period of time. (4) The rocks were involved in recurring spasms of metamorphism. It is thus small wonder that the various isotopic mineral ages from a single granitic rock are generally smeared out over a long time interval.

In Maryland these complications are lessened for the late-kinematic plutons that lack a recrystallized gneissic texture, such as the Guilford, Woodstock, and core of the Ellicott City. Probably the most favorable rocks for dating in the Baltimore area are the undeformed intrusive granite pegmatites, since they did not have an involved protoclastic history, and because there is less possibility that unfused relics are included among their dated minerals. However even the pegmatites yield divergent mineral ages.

Granite Pegmatites, Baltimore Gneiss Dome Area

The swarms of large granite pegmatite sills and dikes which cluster near the mantled gneiss domes have been dated at five localities. Their mineral ages, from Wasserburg *et al.* (1957) and Tilton *et al.* (1959), are summarized in Table 48. The close spatial association of these swarms with the gneiss domes, and their lack of any apparent relationship to the granitic plutons has already been mentioned. It was suggested (p. 186–187) that they formed where pegmatitic fluids, expelled from the gneiss, migrated into expanding tension fractures where the metasedimentary mantle was being stretched across the backs of the rising domes. The fluids presumably emanated from deeper levels within the gneiss, where it was palingenetic.

The Sr/Rb ages of the large microclines and coarse books of muscovite from giant-textured zones in the pegmatites cluster near 440 million years (Table 48). The ability of microcline to retain its original Sr/Rb age despite later severe thermal and mechanical disturbance has been demonstrated for the Baltimore Gneiss and for rocks in other areas. Very coarse muscovites seem to retain their Sr/Rb age nearly as well. Tilton and colleagues (1959) therefore conclude that the 440 m.y. Sr/Rb ages date the primary crystallization of the Baltimore pegmatites as well as the time of their emplacement. This accords with the field and microscopic evidence that the pegmatites have not been significantly deformed or recrystallized.

The 440 million year old pegmatites serve as a useful reference point in estimating the age of other plutonic bodies in the area. Plutons which the pegmatites intrude must be older; this includes the Baltimore Gabbro, Relay Quartz Diorite, and Ilchester Gneiss. Moreover, since the pegmatites have not been significantly deformed or recrystallized, so that even their fine-grained aplitic or granophyric portions lack any trace of a new metamorphic foliation or lineation, they evidently were emplaced after the period of most intense penetrative deformation and metamorphism. Therefore the plutonic bodies that were involved in the severe metamorphism—those that are crushed, granulated, recrystallized into gneiss, schist, phyllonite, or crosscut by later cleavage—must be older than about 440 million years. In the Baltimore-Washington area this includes all the plutonic bodies except the Guilford and Woodstock plutons, possibly the core of

the Ellicott City pluton, and some of the small granite intrusives along the Potomac River near Bear Island.

The pegmatite dates also give information on the mantled gneiss domes, placing a minimum age limit on their period of greatest mobility. During active doming masses of mobilized gneiss punched obliquely upward, pushing, dragging, and stretching the Glenarm mantle, and causing the formation of kyanite along zones of tectonic overpressure. Had the pegmatite swarms been present at that time they could hardly have escaped being deformed and foliated, at least in some places. Therefore the period of gneiss mobilization and strong doming must predate 440 m.y.; this is also suggested by the 580 ± 130 m.y. age of microcline from the diapiric migmatic intrusion (Gunpowder Granite) at the Towson dome. If the pegmatite swarms are genetically related to the gneiss domes as postulated, then the doming may have climaxed and ended about 440 m.y. ago, just as rupturing and pegmatite formation began along zones of greatest stretching in the Glenarm mantle. This, according to the time scale given by Kulp, (1961), would be about middle or late Ordovician.

The Granitic Plutons

Radiogenic mineral ages from plutons in the Baltimore-Washington area (Table 49) do not give precise information on the time of their emplacement. They had complex crystallization and cooling histories, and the various isotopic ages from each pluton are smeared out over a wide time interval. Nevertheless these data, in conjunction with the geologic evidence of relative age, permit deductions about the plutonic sequence and its place in the geologic time scale.

In an earlier section, the sequence of granitic intrusions inferred to be: (1) dark biotite-hornblende quartz diorite (Norbeck, Georgetown), (2) biotite quartz diorite (Kensington), (3) biotite granodiorite (Ellicott City), (4) biotite quartz monzonite (Woodstock), and leucocratic muscovite-biotite quartz monzonite (Guilford). Emplacement of this series began during the height of regional metamorphism and deformation but lasted into the period when these effects had greatly diminished; hence the degree of deformation and recrystallization of the various plutons, as well as their crosscutting relations, help to establish the order of intrusion. Also, the granitic sequence shows progressive increase in K, Na, and K/Na, and decrease in Fe, Mg, and Ca (p. 190–191; Table 37), comparable to the chemical trend of progressively younger granitic intrusions in the Sierra Nevada (Bateman *et al.*, 1963) and other Cordilleran batholiths (Larsen, 1948; Taubeneck, 1957; 1958).

The various isotopic ages of the plutons overlap (Table 49), but they correspond to the intrusive sequence if one compares only isotopic ages of the same kind. Thus the zircon Pb^{206}/U^{238} ages, for example, become progressively younger through the sequence: Norbeck hornblende quartz diorite, Kensington biotite quartz diorite, Ellicott City granodiorite, Woodstock quartz monzonite. The same holds for the zircon Pb^{207}/U^{235}, Pb^{208}/Th^{232}, and Pb^{207}/Pb^{206} ages, and for the biotite A^{40}/K^{40} ages. The radiogenic ages therefore accord with the geologically inferred age sequence, but the overlap of the different kinds of isotopic dates blurs the picture.

The zircon ages are all discordant, and their relationship to the time when the magma initially crystallized (hereafter called the magmatic age) is hard to interpret. If the discordance were due only to admixture of older zircons which had survived anatexis the apparent ages would be too old. On the other hand, if the discordance were due solely to lead loss from the new magmatic zircons—caused perhaps by later meta-

morphism, or by continuous diffusion during slow cooling of the deeply buried mass—the apparent zircon ages would be too young. A combination of both factors—admixture of relict zircons but also post-magmatic lead loss—might yield a spread of discordant ages that overlaps the true magmatic age. The last possibility may well apply to the granitic rocks in Maryland, for they commonly contain "rounded" (relict?) as well as euhedral zircons, but they also had eventful post-magmatic histories.

Tilton *et al.*, (1959) believe that the Pb^{207}/Pb^{206} age may be the most meaningful in many cases, for it is the one that would show least error in the case of lead loss. Geologic evidence supports this choice, especially for the Kensington pluton. The strong deformation and metamorphism of the Kensington Quartz Diorite (p. 166–168) clearly marks it as older than the little-deformed 440 million-year old pegmatites; yet all of its isotopic zircon ages except Pb^{207}/Pb^{206} are considerably younger (Table 49). The magmatic age of the Kensington therefore appears to predate the zircon Pb^{207}/U^{235} and younger ages, and may approach the older Pb^{207}/Pb^{206} age.

The Ellicott City pluton is another whose zircon Pb^{207}/Pb^{206} age may best approximate the initial magmatic age. The granodioritic outer zone of the pluton, where the zircon sample was taken, crystallized under synkinematic conditions and participated in the regional deformation and metamorphism. This is shown by the coincidence of fabric symmetry in the granodiorite and surrounding metamorphic terrane (Cloos, 1947), and by the metamorphism of the pluton's early satellitic dikes (p. 174–175). After about 440 m.y. ago, however, the syntectonic metamorphism was no longer strong enough to impart foliation or lineation to the crystalline rocks (p. 198–199). The syntectonic crystallization of the granodiorite, therefore, seems to be dated more closely by the Pb^{207}/Pb^{206} age (450 m.y.) than by its other isotopic zircon ages (370, 355, 310 m.y.)

These and other relationships suggest that the magmatic ages of granitic plutons in the Baltimore-Washington area are no younger than their zircon Pb^{207}/U^{235} age, and that Pb^{207}/Pb^{206} may give the closest approximation. Perhaps this is not surprising, for the plutons were emplaced in terrane undergoing regional metamorphism of staurolite and kyanite grade, and the temperature of the country rock was probably in the range from 500–650°C (Turner, *in* Fyfe *et al.*, 1958, Fig. 107). Cooling of the plutons was therefore slow, and the zircons were probably held at relatively high temperatures for a long time after solidification—a circumstance ideal for the loss of daughter lead by diffusion.

The magmatic age of a pluton might vary in different parts of the same mass. The Ellicott City seems to be an example of a pluton that continued to grow long after its margin had crystallized. Granodioritic magma was first emplaced and had begun to solidify under tectonic conditions, but the pluton continued to enlarge and to change its composition toward quartz monzonite as tectonic activity waned; the core solidified as porphyritic quartz monzonite without protoclastic or metamorphic deformation (p. 173, 175). A significant age difference between the early rim and later core of the pluton might be expected. Therefore it is not surprising to find that a crosscutting pegmatite dike, derived from the pluton's residual fluids, gives a muscovite age (360 m.y., *see* Table 49) that is substantially younger than the apparent magmatic age (450 m.y.) of granodiorite from near the margin. This age difference may be exaggerated, however, by Sr loss from the muscovite or by other factors.

The granitic magma series in Maryland (p. 189–191) appears to range from Cambrian or early Ordovician to Devonian, based on correlation of the zircon and Sr/Rb feldspar ages with the geologic time scale given by Kulp (1961). The beginning of the granitic

sequence is marked by the Norbeck pluton, composed of dark, strongly deformed and recrystallized biotite-hornblende quartz diorite (Pl. 33, fig. 1). It gives isotopic zircon ages of 570 (±50), 510, 495, and 485 m.y. (Table 49). The minimum age of the granitic series can be placed at about 370 m.y.—the Sr/Rb microcline age of pegmatites from two dikes in the Guilford Quartz Monzonite (Table 49). The Guilford is the youngest of the plutons, and the pegmatites were derived from its residual magma. The date seems reliable because microcline is known to retain its Sr/Rb age with little change, even under rather rigorous conditions (Tilton *et al.*, 1958); in this case the microcline came from fresh, undeformed giant-textured pegmatites. Proof that the date is good comes from the fact that the two Guilford microcline ages, from pegmatites located over 2 miles apart, agree almost perfectly at 375 and 370 m.y.

It seems paradoxical that the mica ages (295, 335 m.y.) from Guilford quartz monzonite are younger than the pegmatites which cut it. This is also true for the Ellicott City pluton, in which biotite ages from the granodiorite are 290 and 315 m.y., but the age of coarse muscovite from the crosscutting pegmatite is 360 m.y. This curious relationship, in which the pegmatites give mineral ages older than those from the host rocks, is now known to be common; examples have also been described from Ontario (Wetherill *et al.*, 1960), Michigan (Tilton *et al.*, 1959), and California (Wasserburg *et al.*, 1959). It evidently reflects the fact that loss of radiogenic daughter elements by diffusion from the small micas of the granitic rocks is much greater than from the very coarse pegmatitic minerals. (*see* Tilton and Hart, p. 360–361). It is obvious that in this area, at least, the magmatic ages of the granitic plutons cannot be ascertained from dates given by fine-grained micas.

In summary, the granitic magma series began with syntectonic intrusions of hornblende-rich quartz diorite, possibly as early as about 570 m.y. ago (Cambrian) but no later than about 490 m.y. ago (Early Ordovician), and ended with leucocratic quartz monzonite about 370 m.y. ago (Devonian). Still earlier, however, large amounts of gabbro and ultramafic rocks were emplaced, followed by small related intrusions of leucodiorite, leuco-quartz diorite, and albite granite. These earlier rocks, which constitute a gabbroic magma series (p. 189–191), have been metamorphosed. Emplacement of the mafic rocks began prior to strong folding and metamorphism, and they are evidently a good deal older than the Norbeck and later granitic plutons. No mineral dates from this older group of rocks in the Baltimore-Washington area have been published.

PALEOZOIC METAMORPHISM

Regional metamorphism in the Maryland Piedmont is bracketed within the Paleozoic by geologic evidence. Metamorphosed Lower Cambrian rocks along the western edge of the Piedmont are unconformably overlain by unmetamorphosed Triassic sediments (Stose and Stose, 1946). The Precambrian basement gneiss nearby has been retrogressively metamorphosed to stilpnomelane-bearing phyllonite. Farther east in the Piedmont the Cambrian sequence dipping off Mine Ridge anticline is metamorphosed to biotite and garnet-zone schists, while the crosscutting Triassic diabase dikes are unaffected (Cloos and Hietanen, 1941). Unmetamorphosed dikes of Triassic diabase also cut strongly metamorphosed rocks of the Glenarm Series near Baltimore, whose age is probably latest Precambrian (this report).

The radiogenic dating confirms the Paleozoic age of the metamorphism, and helps to show that the climax of strong syntectonic metamorphism had passed by about late Ordovician. The intense deformation and high-temperature recrystallization that

converted gabbro to gneissic amphibolite, granitic rocks to gneiss and schist, and pelitic sediments to kyanite schist cannot be younger than late Ordovician, for the metamorphic rocks and structures are truncated by relatively undeformed intrusive pegmatites that are well dated at about 440 million years.

The geologic age relations are well shown along the Patapsco River and the B & O railroad tracks, just below the Ilchester tunnel. Here layered gabbro was intruded by granodiorite (Ilchester Gneiss), tilted on end, and metamorphosed to amphibolite and granodiorite gneiss. An intense metamorphic lineation has been imparted to the amphibolite and dike rocks, and all trace of their original textures was obliterated by granulation and recrystallization. Large pegmatites belonging to one of the 440 million year old dike swarms sharply crosscut the deformed and recrystallized rocks, but the pegmatite's original textural features, including graphic intergrowths and fine-grained hypidiomorphic groundmass, are unmarred by recrystallization. Minor cataclastic effects are developed locally in the pegmatite, but there is no trace of new lineation, foliation, or cleavage.

Relations such as these are found at other localities in the Baltimore area, clearly showing in each case that the 440 million year old pegmatites crystallized after the host rocks had been strongly metamorphosed.

The Baltimore gneiss domes grew concurrently with strong syntectonic regional metamorphism, as shown by the structural relations described earlier and by their relation to the metamorphic zoning. The kyanite zone closely encircles the domes, and the grade of metamorphism drops away from them in all directions. It was concluded (p. 53) that the kyanite formed in response to tectonic overpressures, developed where the domical masses of mobilized gneiss were squeezed upward into the Glenarm cover from deeper root zones in the basement. However the active rise of the domes, accompanied by the intense metamorphism along their fringes, must have waned or ceased by about 440 million years ago when the swarms of undeformed pegmatites were emplaced (p. 199).

The granitic plutons also furnish evidence that the strongest orogenic pulse came before about 440 m.y. ago. The synkinematic plutons—those having recrystallized gneissic textures and chiefly concordant contacts with the surrounding schists—have so far yielded zircon Pb^{207}/Pb^{206}, Pb^{207}/U^{235} ages and a microcline Sr/Rb age greater than 440 m.y., whereas the late-kinematic plutons with well-preserved igneous textures and less evidence of deformation have all younger mineral ages.

Thus evidence from the pegmatites, the gneiss domes, and the granitic plutons all indicate that strong syntectonic regional metamorphism occurred prior to about 440 million years ago (Middle or Late Ordovician).

The first mineral age studies in the Baltimore area showed that micas from the Baltimore Gneiss and Glenarm rocks consistently gave ages of 300–350 million years, and it was concluded that this marked the time of the main Paleozoic metamorphism (Wasserburg *et al.*, 1957; Tilton *et al.*, 1958). Subsequent work has shown that the micas and hornblendes from the granitic intrusions (Table 49) also yield dates mostly near 300 m.y., even though the other mineral ages and geologic relations show plainly that the main Paleozoic plutonic and metamorphic activity occurred earlier. Tilton *et al.* (1959, p. 173–174) discuss the possibility that daughter argon and strontium diffused from the micas during the interval from approximately 450 to 300 million years ago, and they conclude:

"It is believed that in the Paleozoic era (450 million years ago) there was intrusion of pegmatites (and probably granites). Age measurements on biotite indicate still another event 300 million years

ago whose geologic significance is less clear. Perhaps it is to be correlated with a later metamorphic episode that was less intense than the one 450 million years ago. In New England and the Maritime Provinces, orogenies are known to have occurred throughout a time interval of comparable span."

There is ample geologic evidence of milder and less pervasive tectonic and metamorphic activity following the main high-grade metamorphism, but the 300 m.y. mineral ages are not readily linked with any specific event. For example, the pegmatites and many of the granitic rocks locally show strain and mild cataclastic effects. In the pelitic schists there is widespread development of one or more discordant slip cleavages, sometimes with recrystallization of new mica along them. There is also a later widespread but patchy retrogressive metamorphism which formed chlorite and sericite from the higher-grade minerals, especially along zones of movement and shearing (p. 82, 97); Fisher, 1963; Reed and Jolly, 1963). These and other effects may have all played some part in the formation of new 300-million year minerals, or in the diffusion of radiogenic daughter elements from the older ones; however, available evidence suggests that they mark a sporadic succession of local events of decreasing intensity throughout an extended time interval, rather than one distinct pulse of renewed metamorphism about 300 million years ago.

Whatever the explanation, it is plain that soon afterward the temperatures dropped too low for recrystallization or significant loss of daughter elements by diffusion. Not only are mineral ages younger than about 300 m.y. lacking, but geologic evidence indicates that by this time (Middle Carboniferous) the Piedmont in the Central Appalachians was so deeply denuded by erosion that high-grade metamorphic rocks were beginning to be uncovered. Recent studies show that from mid-Ordovician time onward the Piedmont terrane was rising and actively shedding clastic detritus northwestward into the Appalachian geosyncline (Pelletier, 1958; Hunter, 1960; McIver, 1960; McBride, 1962; Yeakel, 1963; Meckel, 1963). The heavy-mineral assemblages and clastic-rock particles from these sediments show that the source area to the southeast furnished exclusively sedimentary and low-grade metamorphic debris until the late Paleozoic, when high-grade minerals began to appear. In the Pottsville Formation (Middle Pennsylvanian) garnet, staurolite, kyanite, sillimanite, and green hornblende became locally abundant as detrital minerals (Meckel, 1963), showing that the high-grade metamorphic rocks had become exposed.

AGE OF THE GLENARM SERIES

The sequence of metamorphosed sedimentary rocks referred to the Glenarm Series are probably at least 20,000 feet thick in the eastern and central parts of the Maryland Piedmont and therefore form an important chapter in the early history of the Central Appalachian geosyncline. Their age bears strongly on one's interpretation of the geosyncline, as well as on certain aspects of its deformation. A Cambro-Ordovician age for the series would support the orthodox view that geosynclinal growth was greatest during the early Paleozoic, with a thick, tectonically active eugeosynclinal portion developing along what is now the Piedmont, and a less active miogeosyncline with clean quartz sands, carbonates, and muds along a belt lying west of the present Blue Ridge (King, 1959). An older Glenarm age, on the other hand, might show that geosynclinal growth along the central part of the chain was already well advanced by latest Precambrian, as it was in the Southern Appalachians. Moreover, it would greatly reduce the width of the early Paleozoic part of the geosyncline and would deprive it of the zone of thickest graywacke sedimentation, ophiolite intrusion, strong regional meta-

morphism, and granite emplacement—the portion generally considered the eugeosyncline.

The solution to one of the most vexing structural problems in the region—the Martic overthrust—also hinges on the age of the Glenarm Series. A Precambrian age for these rocks, where they abut the Cambro-Ordovician sequence dipping off the Mine Ridge anticline in southeastern Pennsylvania, would necessitate a fault of very large displacement along their zone of contact (Knopf and Jonas, 1929a). An early Paleozoic age for the Glenarm, on the other hand, would require only minor faulting and perhaps a sedimentary facies change along this zone (Cloos and Hietanen, 1941; Swartz, 1948).

Although the Glenarm Series at its type locality was originally assigned to the Precambrian (Knopf and Jonas, 1923; 1929) opinion since 1935 has almost unanimously favored an early Paleozoic age (Miller, 1935; Mackin, 1935; Cloos and Hietanen, 1941; Swartz, 1948; Watson, 1957; McKinstry, 1961). A valuable review of the Glenarm controversy has been given by Swartz (1948). Most of the authors who favor a Paleozoic age for the Glenarm have felt that the regional stratigraphic and structural relations in southeastern Pennsylvania are satisfied best by correlating the Setters, Cockeysville, and Wissahickon Formations with the Lower Cambrian quartzites, the Cambro-Ordovician carbonates, and the Ordovician shales, respectively. However Cloos (Cloos and Hietanen, 1941, p. 186–193) pointed out the resemblance and possible identity of the Wissahickon and Antietam (Lower Cambrian) schists where they are nearly in contact at the Martic Line, and cautioned against equating the Wissahickon with the Martinsburg (Ordovician) shales until more evidence is available.

In the Baltimore-Washington area the minimum age of the Glenarm Series is fixed by the radiogenically dated plutonic rocks that intrude it. It was concluded that the granitic series of intrusions began with hornblende-rich quartz diorite, possibly as early as 570 but no later than 490 million years ago, and was closely followed by most of the other gneissic, synkinematic members of the series. All these had been emplaced, strongly deformed, and recrystallized well before the intrusion of the granite pegmatite dike swarms about 440 million years ago. Therefore the minimum age of the Glenarm, which was already deeply buried and undergoing metamorphism when these intrusions began, must be older than about 500 million years.

However, even before the granitic series there was widespread emplacement of gabbro and ultramafic rocks and related small bodies of leucodiorite and albite granite. There are no radiogenic dates for these rocks but they were evidently much earlier. Gabbroic magma first invaded the upper Wissahickon while the sediments were still being deposited (p. 155, 187). Possibly somewhat later the Baltimore Gabbro was intruded into lower Glenarm strata in a semihorizontal sheet many thousands of feet thick, and much of it had crystallized and differentiated before being caught up in the strong orogenic deformation. The gabbro mass and its differentiates were strongly folded and metamorphosed during the rise of the Baltimore gneiss domes. The earliest granitic plutons—those for which we have radiogenic dates—were emplaced at the height or perhaps the later stages of the strong syntectonic metamorphism. These relations are summarized in an earlier section (p. 187–193).

Similar series of mafic and granitic intrusions, are typically found in other orogenic belts, and are somewhat separated in time. In general, the mafic and ultramafic rocks first appear during geosynclinal sedimentation and the early stages of regional deformation, whereas the granitic intrusions—perhaps generated by crustal anatexis—typically appear after compressional deformation and regional metamorphism are well

advanced or finished. In northeastern Oregon, where the plutonic sequence is very similar to that in Maryland, the mafic and granitic series may be separated by at least 70 million years. Alpine-type gabbro-peridotite complexes, with closely related bodies of quartz diorite and albite granite, are Early or Middle Triassic (Thayer, 1963, p. C82, C84), whereas the batholiths and stocks of tonalite and granodiorite are Early to Middle Cretaceous (Taubeneck, 1955; Thayer, 1963, p. C84). On the reasonable assumption that the gabbroic and granitic series in the Maryland Piedmont were separated by a comparable or perhaps even shorter time interval it seems likely that the gabbroic intrusions are no younger than about 550 m.y., and they could be much older.*

Thus the age of the Glenarm Series, based on the plutonic rocks that cut the Wissahickon and older formations, can be no younger than Cambrian—estimated by Kulp (1961) to have ended approximately 500 m.y. ago—and in all likelihood no younger than Early Cambrian.

The maximum possible age of the series is about 1 billion years, for Glenarm strata rest unconformably upon basement gneiss of that age. However, the upper part of the Glenarm is probably not much older than the earliest mafic intrusions, for there is evidence that some of the mafic sheets were disrupted by sliding of Wissahickon sediments that were still unconsolidated (p. 155).

The Wissahickon Formation can no longer be correlated with the Martinsburg Formation or other Ordovician strata, as formerly supposed (Miller, 1935; Mackin, 1935; McKinstry, 1961). Not only is the Wissahickon older, but it now appears that the highest part of the formation or the rocks that once covered it may have formed part of the terrane that furnished clastic detritus to the Martinsburg trough. This is reasoned as follows. The long period of carbonate sedimentation in the Central Appalachian seas during the Early Paleozoic came to an end in the late Middle Ordovician, when muds and then graywacke silts and sands were flushed into the Martinsburg trough. McBride (1962, p. 87) states that "The transition of the lower [shale] member into the upper [sandy] member of the Martinsburg records an increase in the frequency of the turbidity currents and the thickness of their deposits, both of which probably reflect an increasing rate of uplift in the source area." The source area, according to McBride and most previous authors, lay chiefly to the southeast, including parts of the present Piedmont. The uplift to which McBride alludes, in the Maryland segment of the Piedmont, may be correlated with the Baltimore anticlinorium. It is the major anticlinal structure in Maryland east of the South Mountain (Blue Ridge) uplift, and its most active period of growth—judging from the gneiss domes that rose along its core—was coming to an end about 440 million years ago (p. 199). New highlands were doubtless rising along the eastern part of the Piedmont above this growing structure, providing a ready source for some of the Martinsburg detritus. The Glenarm sediments were undergoing metamorphism and granitic intrusion in the core of the anticlinorium, at about the time when detritus destined for the Martinsburg trough was beginning to be eroded from rising terrane near its top.

Radiogenic dating forms the basis for concluding that the Glenarm is pre-Ordovician, but further conclusions must still depend on stratigraphic evidence. Although an early

* Since this report was written Dr. Stanley R. Hart (Carnegie Inst. Wash, Yearbook 62, p. 276, 1963) has reported pyroxene and plagioclase K/A ages of 702 and 580 m.y. in unrecrystallized olivine gabbro from the Baltimore Gabbro Complex. These dates suggest an early Cambrian or latest Precambrian age for the gabbro; however, Dr. Hart warns that the possibility of excess argon in either mineral cannot be ruled out, and could lead to apparent ages that are too old.

Cambrian age is not precluded by the radiometric data the regional stratigraphic relations seem to rule it out. McKinstry (1961, p. 575) and others have pointed out the difficulty of correlating the Glenarm with any part of the known Cambrian section.

A Late Precambrian age for the Glenarm Series has been held to be equally improbable, because sedimentary strata are lacking beneath the known Lower Cambrian rocks in Pennsylvania and Maryland. Along Mine Ridge anticline in southeastern Pennsylvania Lower Cambrian Chickies Quartzite and Hellam Conglomerate lie directly on the much older crystalline basement (Knopf and Jonas, 1929a; Cloos and Hietanen, 1941). In Maryland the Lower Cambrian Weverton Quartzite along the South Mountain anticlinorium is separated from gneissic basement by only the Catoctin Greenstone (Whitaker, 1955). Thus, with no evidence for any Upper Precambrian strata, it has seemed more logical to correlate the Glenarm with some part of the Paleozoic section.

This reasoning begins to lose force, however, if stratigraphic relations farther south in the Appalachians are considered. In the western part of the Virginia Piedmont the Lynchburg Formation comprises a thick sequence of metamorphosed Upper Precambrian clastic sediments. It rests unconformably on basement gneiss but lies beneath the Catoctin Greenstone. Thinly interbedded pelitic and arenaceous rocks, some with graded bedding, constitute the bulk of the formation. Its thickness is estimated at approximately 10,000 feet in the Lynchburg quadrangle (Brown, 1958), and perhaps up to 20,000 feet farther north along the Rockfish River (Bloomer, 1950, p. 765). When the Lynchburg is traced westward onto the Blue Ridge anticlinorium it thins rapidly to only a few hundred feet of graywacke and volcanic rocks (Swift Run Formation) and then pinches out (Bloomer, 1950; Bloomer and Werner, 1955). This has been attributed to overlap onto an eastward-sloping basement.

Thus, the stratigraphic section along parts of the Blue Ridge in Virginia compare closely with those just west of the Piedmont in Maryland and Pennsylvania: Lower Cambrian quartzite or Catoctin greenstone rest unconformably upon basement gneiss. In Virginia, however, it has been possible to recognize the rapidly thickening mass of Upper Precambrian clastic rocks going eastward into the Piedmont.

Still farther south, in the Great Smoky Mountains along the North Carolina-Tennessee border, the Ocoee Series forms an Upper Precambrian clastic sequence at least 30,000 feet thick (King *et al.*, 1958). The Ocoee lies unconformably on a basement of earlier Precambrian gneiss, and on the northwest side of the mountains it is overlain by the basal sandstone of the Chilhowee group (Lower Cambrian or Precambrian (?)). Although large-scale thrust faulting has greatly complicated stratigraphic relations within the Ocoee it has been possible to subdivide it into three groups. The thickest of these, the Great Smoky Group (about 25,000 feet), consists in large part of argillaceous and silty rocks and poorly sorted feldspathic sandstones with abundant graded bedding (King *et al.*, 1958), slump structures, and sole markings. The same features also characterize the Great Smoky in southern Tennessee (Hurst and Schlee, 1962) and northern Georgia (Hurst, 1955; Mellen, 1956).

Thus, voluminous clastic sedimentation during the Late Precambrian is well established from central Virginia southward. Moreover, the general character of the Precambrian sediments, from northern Georgia to Virginia, are quite similar: exceptionally thick sequences of pelite, siltstone and feldspathic graywacke, large portions of which show evidence of deposition from turbidity currents. In Maryland the Wissahickon metasediments have similar thickness, lithology, and primary structures, suggesting that they were deposited in a comparable environment.

A Precambrian age for the Glenarm Series cannot be excluded, therefore, on the grounds that known Precambrian strata are lacking. Upper Precambrian metasediments of comparable thickness and lithology occur in the Virginia Piedmont, approximately along strike, and for a long way to the south. The absence of Upper Precambrian sediments in Pennsylvania and Maryland west of the Martic Line is not a compelling objection, for this lies across strike from the Glenarm, and beyond a zone of possible major faulting. On the basis of stratigraphic relations in Virginia, a thick section of Precambrian strata in the Maryland Piedmont might be expected to thin rapidly westward and disappear.

A more difficult problem arises if the Glenarm Series is Paleozoic: what then becomes of the section of Precambrian Lynchburg metasediments, 10,000–20,000 feet thick in Virginia, that strike directly into the Maryland Piedmont?

In conclusion, the minimum age of the Glenarm Series in Maryland is fixed by the intrusive igneous rocks: radiogenic mineral ages from early members of the granitic series show that the Glenarm is no younger than Cambrian, and the still earlier mafic intrusives suggest that an Early Cambrian minimum age is more likely. The Glenarm metasediments do not correlate readily with the known Cambrian section, but do appear to correspond to the thick clastic sequences of Late Precambrian age in the Virginia Piedmont and farther south. A Late Precambrian age for the Glenarm Series is therefore favored.

SUMMARY

Geologic evidence and radiometric dating show the following age sequence in the Maryland Piedmont:

(1) Strong Precambrian orogeny at about 1000–1100 m.y., transforming earlier rocks to gneiss and migmatite.

(2) Deposition of Glenarm sediments on the eroded gneiss during the Late Precambrian and possibly Early Cambrian.

(3) Intrusion of mafic magma, which began during the time of Flysch (Wissahickon) sedimentation and lasted into the period of strong orogenic deformation and metamorphism.

(4) Reactivation of the Precambrian basement, which rose as anticlinal domes beneath its mantle of Glenarm sediments. The doming and incipient palingenesis of the old basement gneiss was approximately coincident with the strong syntectonic regional metamorphism; this reached a climax and then waned after the formation of pegmatites about 440 m.y. ago (about Late Ordovician).

(5) Highlands rising in the Piedmont shed clastic detritus westward into the Martinsburg trough, beginning in the late Middle Ordovician. This event might be correlated with the growth of the Baltimore anticlinorium with the gneiss domes rising along its core, for the gneiss dome activity culminated about 440 m.y. ago.

(6) Emplacement of plutons belonging to a granitic magma series. This began with intrusion of quartz diorite under syntectonic conditions no later than about 500 m.y. ago (about Early Ordovician), continued with granodiorite, and concluded with quartz monzonite emplaced no later than about 370 m.y. ago (Devonian), after regional metamorphism and deformation had greatly waned.

(7) Temperatures remained sufficiently high for argon to diffuse from micas, feldspars, and hornblende in the crystalline rocks until about 300 m.y. ago (Middle Carboniferous). The Piedmont terrane was being actively uplifted and eroded, and by Middle Pennsylvanian there were medium- to high-grade metamorphic rocks exposed.

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STRUCTURAL GEOLOGY OF HOWARD AND MONTGOMERY COUNTIES

BY

ERNST CLOOS

INTRODUCTION

General Remarks

Structural geology deals with the relative position of rocks in the ground and with the changes these rocks have undergone since they were deposited or intruded. Additional changes are treated by C. A. Hopson in the chapters on Crystalline Rocks and Metamorphism, and our chapters overlap just as the subject matter does. Structural changes cause changes in rocks, and changes within rocks are accompanied by changes in structure or the appearance of new structures. Since the two parts are written by two people with different viewpoints there is some overlap and at times there may even be some contradiction in the interpretations.

In the following I will proceed from large to small structures and try to provide a picture of the phenomena.

Structural Environment

Figure 1 shows the location of the two counties adjacent to the overlap of Coastal-Plain sediments onto the Piedmont crystalline rocks, and Figure 47 shows the position of the counties in relation to the major Appalachian folds.

The Appalachian folded area extends about 115 miles from the Coastal-Plain overlap between Washington and Baltimore northwestward to the western boundary of Allegany County west of Cumberland. In this section are four first-order anticlinal axes (marked I on Fig. 47) in which older formations now outcrop at the surface between younger formations. These zones are anticlinoria which contain many smaller folds, called second-order (II) on Figure 47.

The first-order anticlinoria are the Cumberland, Folz, South Mountain, and Baltimore. They are remarkably evenly spaced and of rather comparable width. Southeast of each of these highs occur abnormally deep depressions, filled with much younger formations. The southeastern half of Howard and Montgomery Counties are on the Baltimore anticlinorium, and the western edge of Montgomery County barely touches the Sugarloaf anticlinorium, which is probably a second-order fold (Fig. 47).

Between the Baltimore anticlinorium and Sugarloaf it is difficult to distinguish large-scale structural units, mainly because the stratigraphic correlations are uncertain. Much smaller structures are common, and folds of all sizes occur in profusion.

Size of Folds

The size of a fold may be measured as the wave length between axes or between flanks of folds; it may also be defined in terms of the depth of exposure in the anticline or of the thickness of deposition in a syncline. This could be called amplitude. Either measurement would give an indication of size.

Several orders of magnitude can be distinguished.

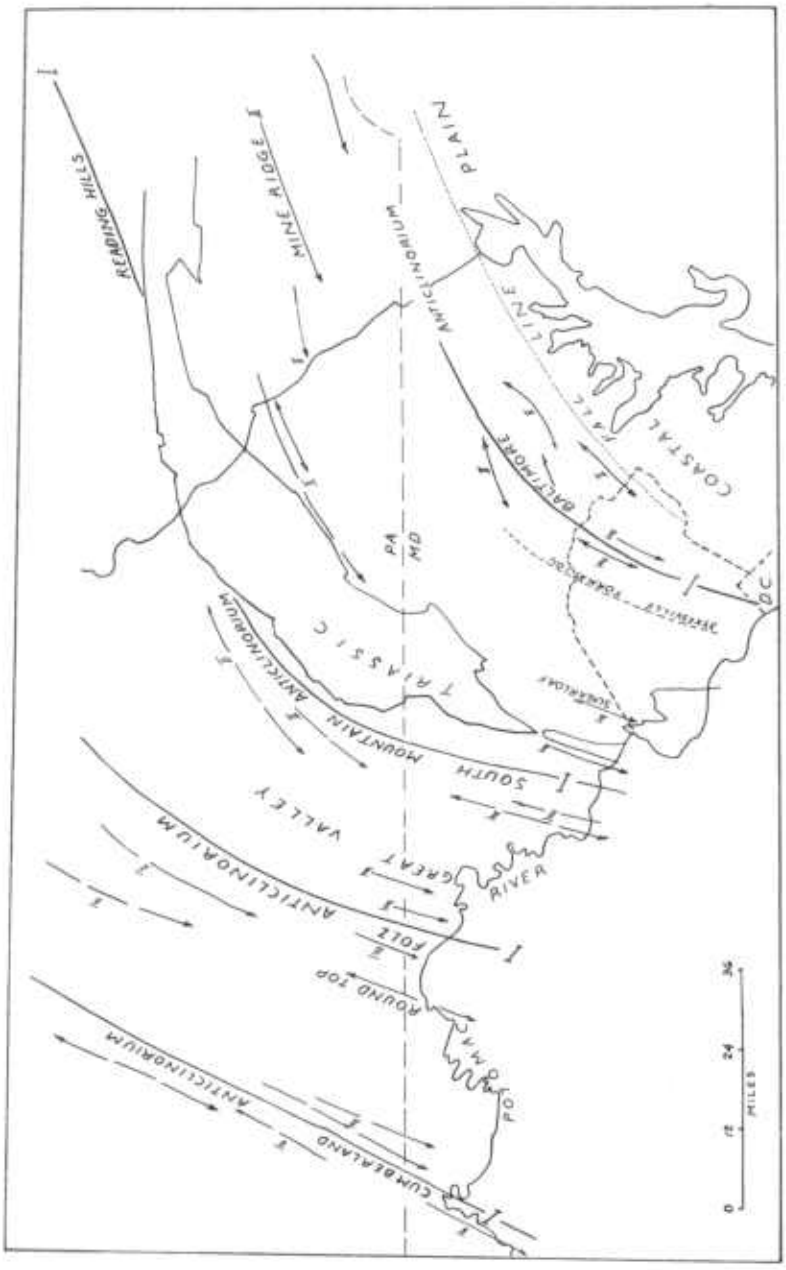


FIGURE 47. Map of Maryland and southern Pennsylvania showing location of Howard and Montgomery Counties (in south-center) in relation to major fold systems of Appalachian area. I: first-order folds; II: second-order folds.

First-order folds are the four anticlinoria marked I in Figure 47. Each consists of several folds with parallel trends which may each uphold a mountain such as Elk Ridge, South Mountain, and Catoctin Mountain in the South Mountain anticlinorium (second-order folds), or such as the smaller Woodstock dome, Chattolanee, Towson, and Clarksville anticlines in the Baltimore anticlinorium (third-order folds) (Fig. 47; Pl. 7).

First- and second-order folds are determined by mapping of stratigraphic units such as key beds or formations (Pl. 7). The Setters quartzite is an outstanding key bed for such mapping, whereas the Cockeysville Marble is missing in too many places. Within the area of Wissahickon schist it is very difficult to outline first- or second-order folds because of the absence of key beds.

The Baltimore anticlinorium, underlying the eastern half of the counties, is well shown in Figure 12 and on Plate 1. In eastern Howard County four anticlines can be mapped: The Woodstock dome and three anticlines to the southwest. Setters quartzite surrounds most of them, and the Baltimore gneiss forms their cores. Above these domes is the Wissahickon schist in which larger folds cannot be mapped. Within the schist are the Sykesville and Laurel Formations, which form one thick layer that outlines the Baltimore anticlinorium as a thick competent unit on its two flanks. The two formations are joined in the District of Columbia and south of the Potomac River.

Third- and fourth-order folds range from a mile to less than 2000 feet in wave length. They can be mapped only on large-scale maps such as that of the Bear Island area or as portions of larger folds along Seneca Creek. They are too large to be seen in one normal outcrop and yet too small to appear ordinarily on a County map. They are the ideal subject for a plane-table survey and detailed work.

Fifth-order folds are those which may be seen and possibly photographed in very good outcrops, such as those in the Potomac River Gorge. Wave lengths of hundreds of feet are common.

On the flanks or in the crests of fifth-order folds appear still smaller folds which may have wave lengths of tens of feet or less, and they may be readily seen in the field. Such folds are common along the Potomac River Gorge and also along Seneca and Patuxent rivers (Pl. 7).

Folds of smaller wave lengths are more numerous, and they may be so small that they can be seen only under the microscope, especially in mica schists and phyllites, where each of their flanks may be made up of only one or two grains of mica.

Between all these orders of magnitude are gradations. Folds vary with competence of the materials involved, the intensity of deformation, the presence or absence of mineral orientations and species, and the magnitude and character of tectonic transports.

MEANING OF FOLDS

Folds and their shapes and sizes have been discussed since the beginning of geological observation, and the relation between thickness of folded layers and sizes of folds and competence of folded material were discussed by Albert Heim (1878) and Bailey Willis (1893) who tried to imitate Appalachian folds experimentally.

There is, however, a lack of knowledge of folds and folding and orders of magnitudes in relation to the physical properties of the folded material. The average textbook on structural geology lists the appearance of folds, their shapes and nomenclature, but says little about orders of magnitude. The most detailed discussion of folds and folding phenomena is by Turner and Weiss (1963) who list no less than 186 items in their index on folds, folding, and points to consider in their interpretation.

The Setters Quartzite, for instance, is a competent key bed and makes prominent second-order folds (Pl. 7; Fig. 12), but within the quartzite are folds of 1-foot wave lengths (Fig. 51) in the quartzite members and folds of $\frac{1}{10}$ inch wave length in the micaceous interlayers. Thus, even competent quartzite beds can be crenulated, and competence alone is no absolute measure.

Folds are convenient indicators of movements and forces and permit determination of tectonic trends and directions of transport. In a fold strata are bent or sheared, and bending takes place by movement of particles, minerals, or components in a plane of motion, with or without formation of additional structures such as cleavages, lineations, or fractures.

Folds indicate the plan and symmetry of deformation for small or large areas in deformed terranes, and, since the deformation plan remains constant as long as the forces and movements remain constant, it is possible to follow a deformation plan through geologic time and detect breaks in continuity and also to find the limits of continuity in space. All folds indicate deformation, and small folds may be more significant than large ones.

It is one of the most astonishing facts that in the Maryland section of the Appalachians the axes of first-order folds are in general parallel with those of the small folds and that axial trends remain consistent from the base of the Cambrian to the top of the Carboniferous. In so large an area and so thick a geological section deviations must be expected, and they are abundant, but major breaks are strangely absent.

CO-ORDINATES AND DEFORMATION PLANS

It has become customary to refer deformation to a co-ordinate system and to describe crustal deformation in terms of a plan within such a co-ordinate system.

The customary co-ordinates *a*, *b*, and *c* are here used; *b* becomes *B* if it represents observed directions. Usually it is the fold axis. Normal to *B* is the *ac* plane which is the deformation plane in which movement has occurred.

Axis *a* is defined as the direction of tectonic transport and may vary with the size of the area under consideration. For instance, if strata are bent in a fold, they will slide over each other in the two limbs. The direction of sliding is then the transport at the boundary between layers and is often recognized as striation on bedding planes. If, on the other hand, the tectonic transport of the entire Appalachian folded area between Hagerstown and Cumberland is described, *a* would be approximately horizontal or would dip slightly to the northwest across the folds. It is therefore essential to define the size of the area to be described by an *a* axis.

Generally *a* axes are far more variable than *b* or *B* axes because of the changes in direction of up-and-down movement which can, however, still be perpendicular to a constant *b* or fold axis *B*. Axis *c* is normal to the *ab* plane.

Other elements that determine a deformation plan are such structures as cleavages, bedding, lineations and fractures. Their mutual geometrical relations make up the deformation plan and establish the symmetry of a deformation which in turn permits conclusions on the symmetries of motion which produced the observed structures (Turner and Weiss, 1963).

LARGE AND SMALL STRUCTURES

Many geological maps show anticlines and synclines either as outcrop pattern, as for instance the Baltimore County map (Knopf and Jonas, 1929) or as lines and outcrop

patterns, as the Allegany County map of the Maryland Department of Geology, Mines, and Water Resources. Frequently these are thought to be the only significant structures. In addition, faults and possibly attitude of formations (strikes and dips) are shown.

In the Cecil County report of 1902 (Bascom, 1902, p. 103) five pages deal with structural geology of the crystalline rocks in a terrane where folds, cleavages, and an infinite array of structures are well exposed along the Susquehanna River and its tributaries. At that time the British geologists covered their maps with structure symbols, and detailed descriptions filled their reports.

Garrett County had also been mapped then (Maryland Geological Survey, 1902). The discussion of structures takes up 18 pages and is a description of the major synclines and anticlines in that county.

The new geologic map of Garrett County (Amsden, 1953) shows the major first-order anticlinoria and some symbols for attitude of bedding in addition to much other new information. The new geologic report (Amsden, 1954) contains six pages on structures and without details.

The report on Baltimore County (Knopf and Jonas, 1929) has seven pages on structural geology describing the general geology as seen on the geological map which shows no structure symbols. Cleavage or schistosity do not occur in the index in spite of the fact that the entire county is underlain by crystalline schists!

Much progress has been made in the use of so-called "minor structures". Some structures are as typical for a formation as are its composition, color, or grain size. As recently stated by Ramsay (1963, p. 145):

> "Careful mapping of these structures shows that they are, without exception, ordered and systematic although [as will be discussed later] the geometric arrangements are often complex as a result of repeated phases of deformation. Their importance was realized by some of the earliest workers, in particular Clough who methodically recorded the orientation of linear structures and the strike of minor fold axial planes on his maps (see Peach and others, 1907, 1910)."

Uses of the detailed data are numerous: Stratigraphic thicknesses cannot be measured in folded areas unless it is certain that sequences are not repeated. For this it is essential to determine dip and top and bottom of strata. Many such measurements are needed before a section can be determined with confidence. For such measurements cross-bedding, graded bedding, cleavage and bedding relations can be used.

For determinations of deformation symmetries, overturning of folds, and movement directions a structural inventory is needed including the orientations of bedding, cleavages and lineations.

The need for detailed observations has been summarized elegantly by Gilbert Wilson (1961, p. 424) as follows:

> "The object of this work is to try and present, to the ordinary field geologist, the importance and significance of the many small scale structures which can be seen in the rocks. Such small scale structures, to my mind, are those which can readily be seen in the field, but are too small to record (other than diagrammatically) on a map of 1:10,000 or lesser scale. As Professor Eugene Wegmann would say, they are of the scale of the exposure or the hand specimen. The significance of these structures is four-fold: They may be used in the elucidation of the geometry of the larger scale structures and the stratigraphical succession of the beds; they may be used as 'tectonic weather-cocks' for indicating the directions and sense of the local movements which have affected the rocks; they may give an indication of the distribution of the stresses which were responsible for the deformation; and they may indicate a time sequence of different phases of deformation, in which case the structures assist in deciphering the tectonic history."

In keeping with this viewpoint the Washington County report (Cloos and others, 1951) includes a 40-page discussion of structural geology. The general structures are discussed as well as folds of all orders of magnitude, cleavage, thrusting, boudinage, striation, drag folds, lineations, and joints. Since the area is almost entirely west of the metamorphic rocks of the Piedmont, there is no discussion of rock fabrics or metamorphism. Unfortunately this report is already out of date because there is no discussion of primary sedimentary structures. Systematic observations on such structures have contributed greatly to our knowledge of the early stages of the formation of the Appalachian geosyncline and the directions of currents which deposited the sediments. During the past 10 years great progress has been made in the study of paleo-currents (Potter and Pettijohn, 1963).

FABRIC DATA

We have not undertaken a fabric analysis of all stratigraphic units in the two counties, but where such data existed and seemed relevant they are presented. Fabric analyses are based on determinations of orientations and are presented in statistical diagrams. For mica orientations 100 grains are usually satisfactory, but quartz-axes measurements should include at least 200 grains. All diagrams are projections into the lower hemisphere of a Schmidt net for microscopic and megascopic measurements.

STRUCTURAL INVENTORY AND DEFINITIONS

GENERAL REMARKS

All pre-Triassic rocks in Howard and Montgomery Counties are intensely folded, dislocated, and metamorphosed. In addition, only a few formations can serve as key beds. We also suspect primary changes of sedimentary facies from west to east as well as from bottom to top of the section. These factors make it difficult to correlate units, to trace them across the map, and to establish a stratigraphic sequence. It is therefore necessary to make use of all possible observations especially on the petrography and petrology as described by C. A. Hopson and the detailed structures as outlined below.

STRUCTURE SYMBOLS ON COUNTY MAPS

The geological map of Howard County (Cloos and Broedel, 1940) shows the following structural symbols:

Strike and dip of primary structures; "flow planes" in igneous rocks; "bedding" in sediments

Horizontal projection and plunge of "flow lines" in igneous rocks and "stretching" in crystalline schists

Strike and dip of cleavage planes or schistosity

Strike and dip of "fracture cleavage"

Faults.

On the Montgomery County map (Cloos and Cooke, 1953) there are no symbols for primary igneous rocks because they were not needed. There are symbols for bedding, cleavage and schistosity, lineation, fold axes, and faults. The term lineation is much preferred to "stretching" as used in the older map. Fold axes are also lineations but deserve special designation. The symbols for primary bedding and secondary cleavage have been reversed because a committee of the U. S. Geological Survey attempted a more standardized usage of symbols for greater convenience, and we adopted the committee's recommendation. For simplicity, not all observed structures are included on the map.

Structures are essentially planar or linear, and the two kinds are pictured in the map symbols. In addition one may separate primary from secondary structures with possible emphasis on folds. This system will cover most of the possibilities and permit fairly complete presentation. With these data a deformation plan can be constructed at most localities and for many types of rocks without using different symbols for each rock type.

Table 50 presents the number of symbols shown on the two county maps and the number of field measurements taken. The total number of symbols is 1386, and the number of measurements 3531. The distribution of these measurements and symbols reflects (1) size of areas underlain by formations; (2) distribution of outcrops and, since the best outcrops occur in stream valleys, the drainage pattern; (3) resistance of rock types against weathering; (4) structures of the formations such as folds and/or prevalence of cleavage; (5) lithologic characteristics; (6) degree of metamorphism; and (7) subjectivity of the observer who becomes enthused over one kind of formation which shows folding better than another, or who is attracted by the excellent Potomac River outcrops and pays less attention to the deeply weathered, generally dusty, railroad cuts. As a case in point, the Bear Island area is small but beautifully exposed, and more than 10 per cent of all measurements were taken in this small area. Bear Island furnished almost all the readings for bedding (162 out of 172), but it shows only 33 cleavage measurements as compared to 439 for the counties in the Wissahickon schist alone. The consolation lies in the fact that the good exposures along the Potomac River supply far more fundamentally important information than the remaining area.

The Wissahickon Formation west of Sykesville shows bedding more readily than does the schist east of the Sykesville. The Kensington Quartz Diorite covers an area only a little larger than the Ellicott City Granodiorite but shows lineations very much better.

TABLE 50

STRUCTURE SYMBOLS ON GEOLOGIC MAPS OF HOWARD AND MONTGOMERY COUNTIES

SYMBOLS FOR:	BEDDING	CLEAVAGE, FOLIATION, SCHISTOSITY	LINEATION	FOLD AXES
Harpers Phyllite	6 (28)	8 (26)	5 (5)	— (3)
Wissahickon Fm. west of Sykesville	105 (172)	159 (439)	96 (160)	14 (93)
Wissahickon Fm. at Bear Island	(162)	(33)	(26)	(24)
Ijamsville Phyllite	14 (33)	96 (191)	25 (29)	2 (14)
Sykesville Formation	—	195 (568)	60 (137)	2?
Laurel Formation		20 (72)	5 (12)	
Wissahickon east of Sykesville	78 (135)	188 (547)	47 (114)	2 (7)
Cockeysville and Setters (only Howard County)	28 (70)	5 (18)	1 (20)	
Baltimore Gneiss (only Howard Cty)	—	51 (35)	8 (23)	
Ellicott City Granodiorite		13	2	
Kensington Quartzdiorite		51 (156)	25 (49)	
Mafic rocks		48 (73)	12 (12)	
At Bear Island		— (111)	— (51)	
Serpentinite		13 (36)	2 (2)	
Total symbols and measurements	231 (438)	847 (2305)	288 (647)	20 (141)

Total map symbols	1,386
Total measurements	3,531

Cleavage is obviously the most prominent structure, and lineation is next (2305 and 647 readings respectively). Also included are factors of spacing of measurements and of space on the map. At some well-exposed area like Bear Island only very few symbols can be plotted on the map.

During field work no attempt was made to sample in a particular manner, but we tried to distribute symbols as evenly as possible on the finished map. This is probably not the best method, and a more systematic sampling method would be desirable.

LIST OF STRUCTURES

Cloos (1937, p. 37–40) prepared a list of structures found in the crystalline rocks, but usages of terms have changed, and new observations have been made. In addition not all the structures of that list appear in the counties. In Table 51 is a new list of structures noted in Howard and Montgomery Counties.

PRIMARY AND SECONDARY STRUCTURES

The distinction between primary and secondary structures is useful in any attempt to unravel the history of a rock but becomes too confining where more than two structures or groups of structures occur in rocks of varying ages.

For instance primary stratification can also include a primary linear structure reflecting currents during deposition. Superimposed on these can be a cleavage and a fold axis which transect bedding and old lineation, and these can in turn be cut by a second fold axis or lineation, and another cleavage. In that case there would be two primary structures and four secondary ones which represent two or more superpositions.

On the other hand primary structures are those which date back to the formation of the rock and thus deserve greater emphasis than the later ones. For instance current directions in a sediment may be equally or more important to the interpretation of an area than all superimposed later structures.

Wilson (1961, p. 439) uses the terms "original sedimentary structures" and "secondary structures of a tectonic origin" in an attempt to separate structures of different ages.

Patterson and Weiss (1961, p. 870, 876) use the terms "initial" or "inherited" and

TABLE 51

LIST OF STRUCTURES IN HOWARD AND MONTGOMERY COUNTIES

Primary structures
Bedding, stratification in sedimentary and metamorphic rocks (S_1)
Foliation in igneous rocks
Flowage lineation in igneous rocks
Secondary structures
Cleavages
Bedding schistosity (S_2 parallel S_1)
Flow cleavage, axial plane cleavage (S_3)
Fracture cleavage (S_4)
Other cleavages (S_5, S_6, etc.)
Lineations
Fold axes
Crenulation axes, "crenulation lineation," puckering axes, etc.
Intersection of surfaces such as S_1, S_2 and S_3 (B, β)
Rodding, rolling, elongate bodies such as vein quartz hinges in small folds
Mineral orientations, single minerals or clusters
Orientation of fragments, "inclusions" in rows, streaks, etc.

"imposed" fabrics. Together inherited and imposed fabrics can result in a new composite fabric which did not exist prior to the event responsible for the imposed fabric.

The terms "primary" and "secondary" are retained here even though the distinction is not entirely satisfactory. Recently the sequence of surfaces and lines have been simply numbered by S_1, S_2, S_3, etc. or L_1, L_2, L_3, etc., indicating their succession. This system is useful and is in part applied here, but it does not make the important distinction between the early premetamorphic history of a rock and its alteration through metamorphism, which may indeed take place many millions of years later.

In addition the numbering of structures in sequence alone is inadequate because the reader cannot draw his own conclusions on the validity of the authors' observations and correlations, and a number cannot substitute for a description. Numbering is useful if descriptions and phenomena are put in sequence and labelled so that the reader knows the reason for the author's interpretation of the sequence of events, but it is not advisable to use numbers for large areas that contain different rock units. For instance: bedding in schist is labelled S_1; S_1 is cut by a cleavage S_2 which in turn is parallel and formed simultaneously with a foliation in an intrusive granitic rock that should be labelled S_1 because it is the first structure in the igneous rock. In this case S_2 in the schist would be of the same orientation and age as S_1 in the granite.

In Howard and Montgomery Counties the Wissahickon schist next to the Ellicott City Granodiorite shows bedding (equals S_1). Recrystallization of minerals in the bedding planes has resulted in a bedding-plane schistosity S_2 parallel to S_1. A good cleavage (S_3) cuts across S_1 and S_2 and is contemporaneous and parallel with primary foliation in the granodiorite, which should be called an S_1. If the granodiorite and its schist mantle are considered together in a broader deformation plan, S_3 in the schist is the equivalent of S_1 in the granodiorite. The same relation applies to the lineations. Thus, labelling alone does not serve the purpose.

SYSTEMATIC DESCRIPTION OF STRUCTURES

BEDDING, STRATIFICATION

Bedding can be seen in most of the sedimentary rocks but not in all outcrops. It can be recognized by lithologic changes, color banding, changes of grain size or mineral composition, or by differences in alteration. In all pre-Triassic sedimentary rocks bedding tends to be obscured by superimposed structures.

Bedding is distinct at many places in the Setters Formation, Cockeysville Marble, Wissahickon Formation, Ijamsville and Harpers Phyllites. In the crystalline rocks it becomes more readily visible as metamorphism decreases from east to west. The outstanding and best-exposed region is the Potomac River Gorge in Montgomery County, but the Patapsco River and other stream valleys permit study of bedding at many places. The post-crystalline rocks from the Triassic upward are well bedded.

Bedding is absent or very indistinct and rare in the Sykesville and Laurel Formations. Because of the character of the formations it probably never was distinct.

Bedding is very questionable in the most intensely altered Baltimore Gneiss, which is well foliated and banded. It is, however, not proven that this banding was once bedding, even though it has been so interpreted by some authors (Knopf and Jonas, 1929).

Excellent bedding in the Wissahickon Formation is shown in Figures 1 and 2 of Plate 36 and in many figures by Hopson from Bear Island. Graded bedding and soft-sediment deformation are also described by Hopson.

FOLIATION IN IGNEOUS ROCKS

The term foliation is here used for a primary planar structure which may be due to flowage, settling, or crystallization processes. It is not meant to suggest a mode of origin but rather to describe a phenomenon. Foliated igneous rocks are the Ellicott City Granodiorite, Relay Quartz Diorite, Kensington Quartz Diorite Gneiss, the Norbeck Quartz Diorite, the Baltimore Gabbro Complex and portions of the Bear Island Granite and the Georgetown Complex. The other igneous rocks show a very faint foliation at a few places or are massive.

The Ellicott City Granodiorite foliation has been described (Cloos, 1933; 1947). In the outcrop it is easily recognized as a parallel orientation of dark inclusions and minerals in planes with a consistent regional orientation.

A statistical analysis of mica orientations under the microscope is shown in Figure 48 for 125 biotite grains. (*See also* Cloos, 1947, Fig. 4.) The maximum number of mica cleavage poles coincides with the foliation pole. A weak girdle is suggested.

Two and a half feet from the granodiorite contact the biotite orientation differs in that a girdle is shown but with a maximum coinciding with the foliation pole (Fig. 49). Foliation is parallel with the granodiorite contact and with the wall-rock cleavage. Lineations in granodiorite and schist are also parallel.

Most intensely foliated is the Kensington Quartz Diorite, but the foliation is at least partly due to later metamorphism.

Foliation in the Gabbro Complex is not uniformly distributed. It is in part segregation banding and partly a secondary orientation due to metamorphism. (*See* Hopson, section on Baltimore Gabbro Complex.)

The Relay Quartz Diorite is also foliated but occurs only in the southeast corner of Howard County, and it was not possible to study its regional pattern in any detail.

The Bear Island Granite is foliated in some places, and an example has been described

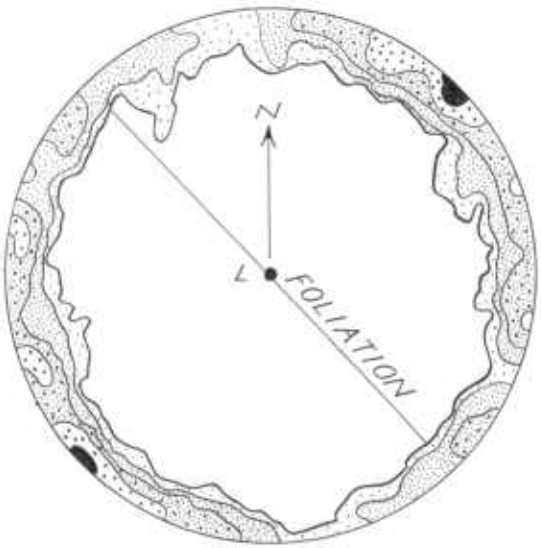


FIGURE 48. Statistical orientation diagram of poles of 125 biotite grains in Ellicott City Quartz Diorite. Contours 1-2-3-4-5-(6-7) per cent. L: lineation. Lower hemisphere. Diagram is horizontal with North at the top. A distinct foliation trends northwest, but a girdle is suggested with L as vertical axis (after E. Cloos, 1937). Center of pluton.

FIGURE 49. Statistical orientation diagram of poles of 200 biotite grains in Ellicott City Quartz Diorite, 2½ feet from contact. Contours 1-2-4-6-9- per cent. L: lineation. Lower hemisphere. Diagram horizontal. North at top. A distinct foliation parallel to that in Figure 48 but a complete girdle indicates a prominent vertical axis (after E. Cloos, 1947).

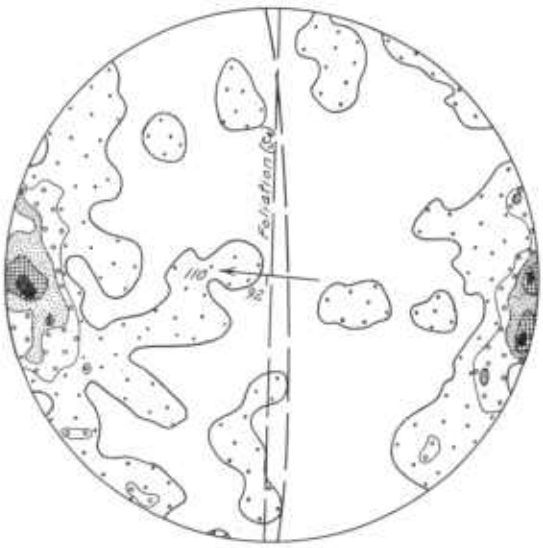


FIGURE 50. Mica diagram of Bear Island Granodiorite near discordant contact with schist. 100 grains. Contours 1-3-5-7-9(10) per cent. The maxima are poles of foliation which is parallel to the contact. Diagram vertical. 110: azimuth reading 20° south of east (after E. Cloos, 1954).

by Cloos (1954, p. 81). The foliation is mostly weak and not readily seen. At some places, especially near contacts with schist, fragments and mica grains are aligned. Figure 50 shows the accumulation of 100 muscovite cleavage poles in the pole of the megascopic foliation in a discordant granite dike at Bear Island. The muscovite orientation is distinct and parallel with the contact which cuts across bedding and bedding schistosity of the adjacent schist. (*See also* Cloos, 1954, Figs. 2 and 6.)

FLOWAGE LINEATION

In some of the igneous rocks a primary igneous lineation due to mineral alignment or inclusions has in the past been called flow lines on the Howard County map legend (Cloos and Broedel, 1940) or, more generally, lineation (Cloos, 1946; Cloos and Cooke, 1953). The same rocks that are foliated are also lineated, especially the Ellicott City Granodiorite, Kensington Quartz Diorite, and portions of the Baltimore Gabbro. The lineation in the Ellicott City Granodiorite seems to be a primary structure, but this is not certain for the Kensington Quartz Diorite and Baltimore Gabbro (Cloos, 1947, p. 1–4). It is due to the parallel orientation of elongate inclusions and of mineral clusters. The Relay Quartz Diorite, Guilford Quartz Monzonite and Bear Island Granite have not furnished any data on lineations.

ORIENTATION OF PRIMARY STRUCTURES

Bedding in sedimentary rocks is folded mostly on northeast–southwest-trending axes, except in the vicinity of the gneiss domes in the Baltimore anticlinorium. Its orientation is discussed under Regional Orientation of structures. The geologic maps of the counties include symbols for bedding attitudes. In the igneous rocks foliation and lineation are parallel to the contacts in small or narrow bodies and deviate from this only in the very much larger Baltimore Gabbro Complex. Since the contacts are generally parallel to a wall-rock cleavage, foliation and lineation in the Ellicott City Granodiorite and Kensington Quartz diorite are parallel to a dominant wall-rock cleavage adjacent to it. A noteworthy exception is the Bear Island Granite which is a discordant intrusive as shown in Figures 1 and 2 of Plate 37.

Other igneous contacts are also discordant, as for instance the Guilford Quartz Monzonite which is rather massive so that internal structures are not obviously parallel with contacts. In the discordant pegmatite dikes of the Patapsco River Valley below Ilchester, a very striking compositional banding parallels discordant contacts.

CLEAVAGES

General Remarks

The terminology of secondary planar structures approaches chaos. The terms used differ in accordance with an author's background, taste, and observations. Terms such as cleavage, schistosity, foliation and *S* plane are used interchangeably by different authors. Frequently a genetic implication is added by a descriptive prefix such as flow cleavage, fracture cleavage, shear cleavage, strain-slip cleavage, pseudocleavage, axial-plane cleavage, bedding foliation, crenulation cleavage, original cleavage, false cleavage and others.

In the following I will use the term cleavage as equal to S_2, S_3, S_4, etc., for secondary planar surfaces. Inasmuch as cleavage describes a surface along which the rock can be cleaved or is cleavable, the cleavability is a property which applies to the entire rock and cannot be confused with a direction of joints, except where joints follow cleavage. A cleavage can parallel bedding, but bedding is a primary surface and may well be called S_1 or *SS*.

Bedding Schistosity and Axial-plane Cleavage

In the metamorphic rocks of Howard and Montgomery Counties the primary bedding plane has become a plane in which intense recrystallization has taken place. The bedding is more distinct because mica grains larger than original clastic grains have grown parallel

to bedding, which is thus emphasized. This cleavage is referred to as "bedding schistosity" by Hopson and Fisher because it is due to recrystallization, possibly prior to folding, or at least at an early stage of more or less static-mimetic crystallization.

In spite of the fact that this schistosity is not a new secondary direction, but parallel to S_1 or SS, it is here called S_2 because it is due to a secondary, post-bedding process. S_2 is then parallel to S_1.

In intensely folded schists bedding schistosity can grade into axial-plane cleavage (S_3), or it may be eliminated.

Figure 51 shows a fold in the Setters Formation. Bedding is visible as light and dark

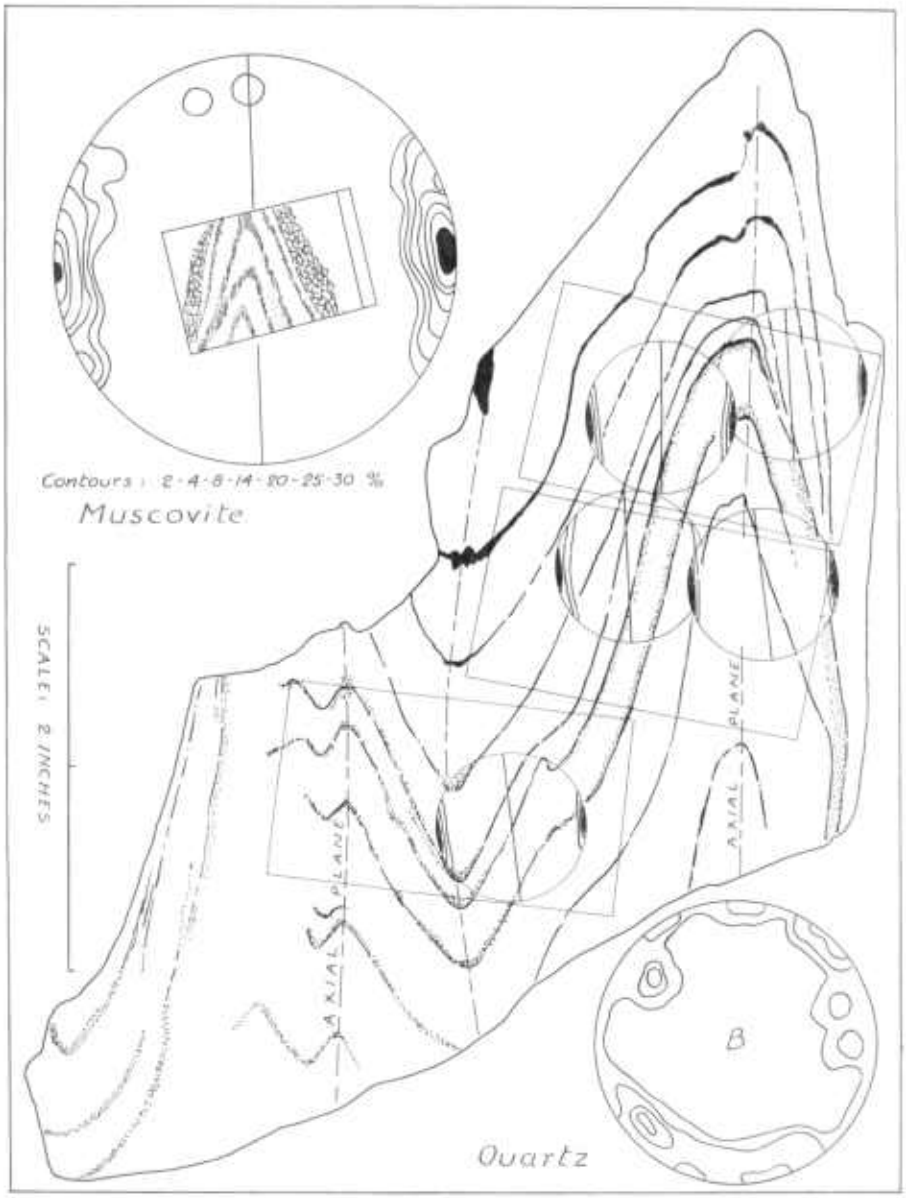


FIGURE 51. Drawing of fold in Setters quartzite (by R. Fiske). Girdle not suggested by mica (upper left) but by quartz orientation (lower right). Wave length of fold 3.5 inches.

banding due to varying amounts of mica, feldspar or quartz and due to changes in grain sizes in these layers. Five thin sections were cut normal to the fold axis in the plane shown in Figure 51. Barely visible in the drawing is an axial-plane cleavage which is very distinct in thin sections (Pl. 38, fig. 1) as a rigid orientation of muscovite blades. It is also prominent in the measurement of several hundred muscovite orientations of Figure 51. A single maximum appears in six muscovite diagrams which are distributed over the entire fold. The plane of the maximum of mica poles is the axial plane of the fold.

Bedding schistosity is a prominent structure plane at many places, but it can be overpowered by or can grade into axial-plane cleavage (S_3). The gradation of bedding-plane schistosity into axial-plane cleavage is shown in Figure 1 of Plate 38 and in Figure 52. Axial-plane cleavage is the prominent structure strongly emphasized by muscovite blades in parallel orientation. Bedding becomes visible in the accumulation of larger muscovite blades of the same freshness and kind; the muscovite is larger, even though it is not parallel with bedding. In addition there is a change of quartz-grain size across bedding planes. In the crest of the fold, muscovite is not parallel with bedding, and there is no bedding schistosity but only an axial-plane cleavage.

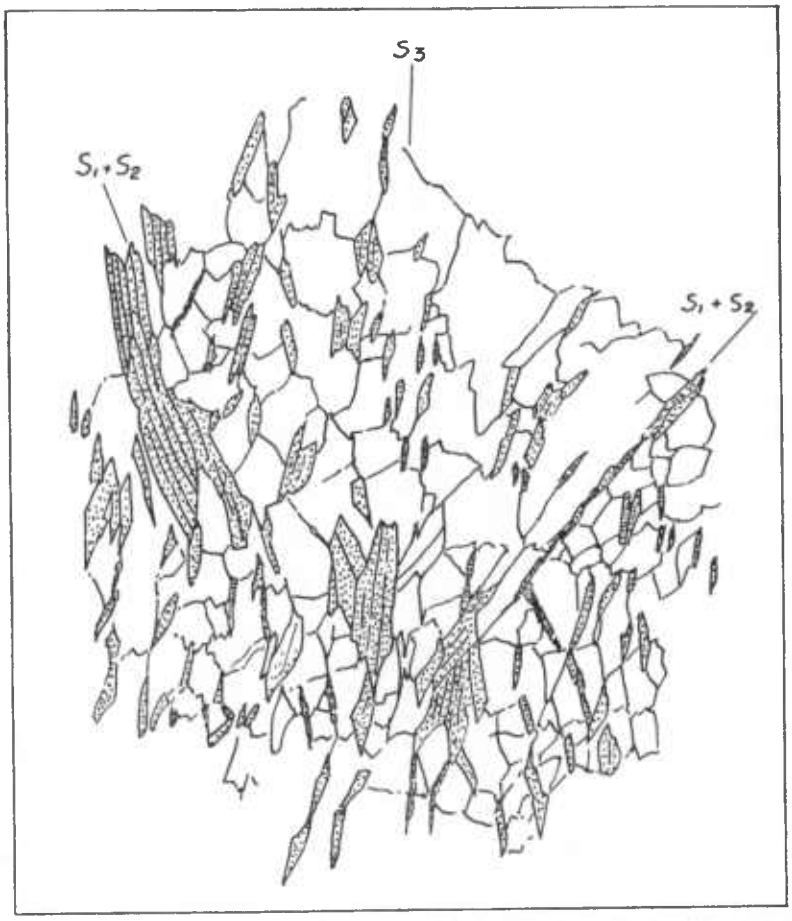


FIGURE 52. Detail of fold in Figure 51. Drawing of thin section. Bedding schistosity is indicated by large muscovite grains parallel to S_1 in limb of fold. In crest S_3 cuts across bedding which has disappeared. Generally, larger grains in plane of bedding, smaller grains in axial-plane cleavage (S_3).

There is no indication of two intersecting cleavages or of deformation of S_2 by S_3. There is no bending or distortion of earlier muscovite in later transecting shear planes. If one looked only at the crest of the fold, bedding-plane schistosity would be overlooked. This is also suggested in the diagrams of Figure 51 in which the predominance of axial-plane muscovites is striking. If the bedding schistosity were distinct the diagrams would show girdles for mica poles with the fold axis at the center of the diagrams.

Foliation in Baltimore Gneiss

The Baltimore Gneiss is well foliated almost everywhere, as described by Hopson. In the paragneiss the foliation is probably primary bedding (S_1). Elsewhere the lamination, banding, or foliation is probably secondary. Parallel to banding and layering is a mineral orientation probably equivalent to bedding schistosity (S_2). It is folded with the layers, and only rarely is there a suggestion of an axial-plane cleavage in the crests of folds.

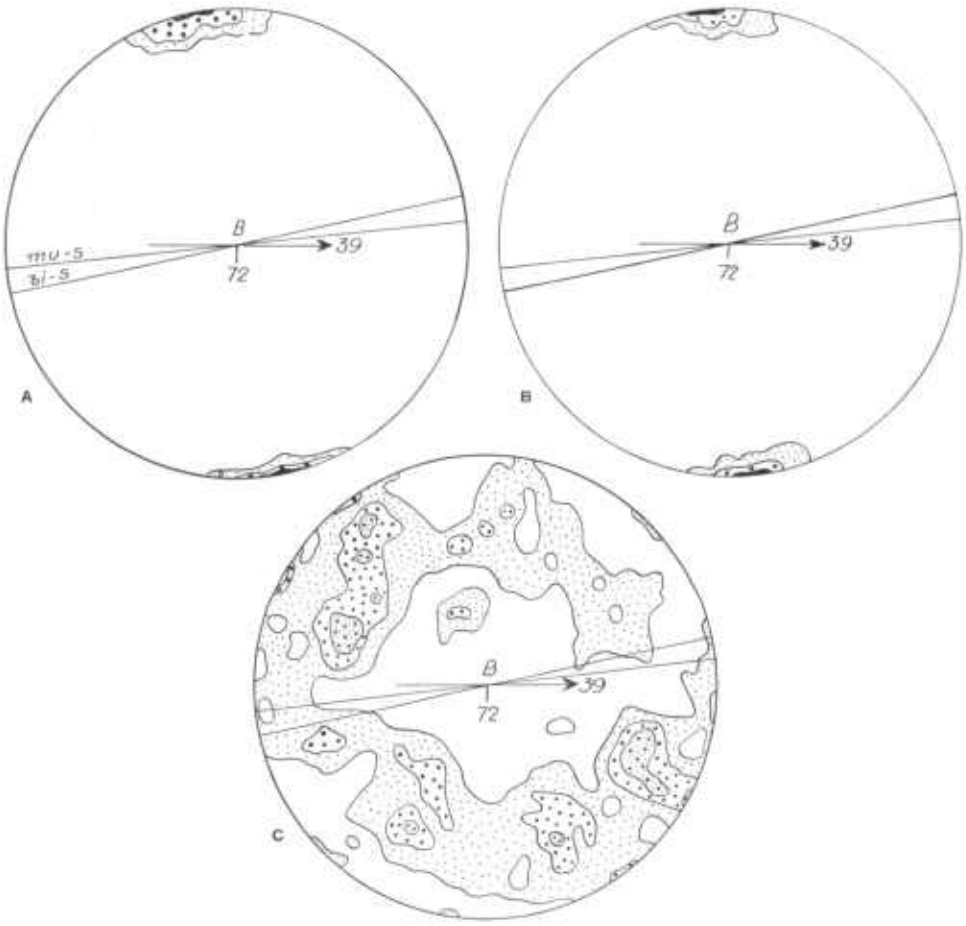


FIGURE 53. Statistical orientation diagrams of mica and quartz (after Stefansson, 1953). Locality: Near Mayfield 100 feet southeast of intersection of Frederick road with Triadelphia Turnpike (Plate 7). Baltimore gneiss, on east side of asymmetrically overturned limb of anticlinal dome.

A. 100 biotite poles, B. 38 muscovite poles. In both diagrams one strong maximum is the pole of foliation. No girdle suggested in mica diagrams. Contours for A and B 10-20-30 per cent, C. Quartz orientations, 400 grains. A girdle is suggested. Contours 1-2-3 per cent.

The regularity and degree of mica orientation in the foliation varies with the type of Baltimore Gneiss; the fine-grained layers show the best orientation. Coarse-grained gneiss and crenulated varieties tend to be much less regular than fine-grained ones, and the degree of orientation is well shown in an analysis of gneiss fabric carried out by Karl Stefansson (1943). His measurements include the orientations of 12,300 quartz grains, 2,950 biotites, and 1,028 muscovites. There are 77 orientation diagrams of Baltimore Gneiss and two of Setters quartzite.

All quartz diagrams and many of the mica orientations show distinct girdles with well-defined *B* axes, some of which are shown in Figures 53–56 for Howard and Montgomery Counties. None of the quartz orientations reflect the foliation. All mica diagrams show maxima for foliation, and some indicate the same *B* axis that is shown by the quartz orientation.

Figures 53 and 54 show the strict orientations of biotite and muscovite in the poles of foliation (S_2). A girdle is barely suggested in Figure 54B but is confirmed in the quartz diagram (Fig. 54 C). Very different are Figures 55 and 56 from localities 21 and

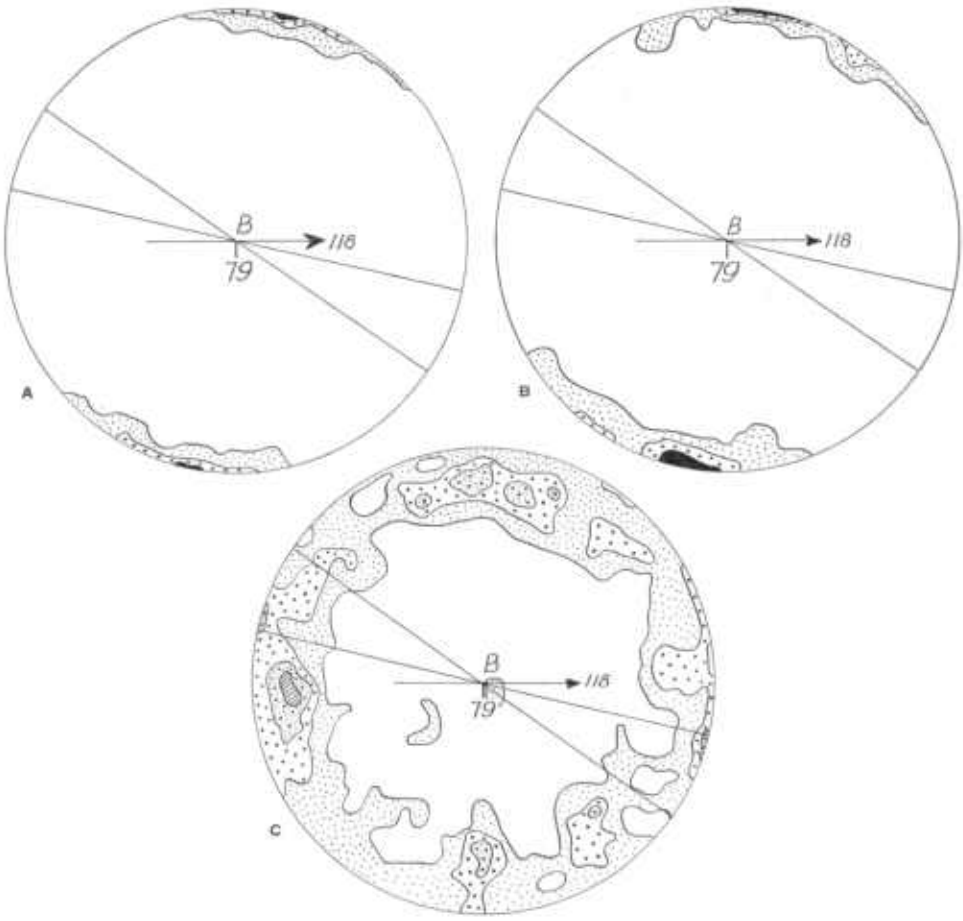


FIGURE 54. Orientation diagrams of mica and quartz (after Stefansson, 1943). Locality: 150 feet southeast of road intersection at Nichols, southwest edge of Clarksville dome. Baltimore gneiss.

A. 100 biotite poles, contours 6-14-22 per cent. B. 55 muscovite poles, contours 4-12-20 per cent. In both diagrams one strong maximum is the pole of foliation. Girdle suggested only in quartz diagram. C. 400 grains, contours 1-2-3-4 per cent.

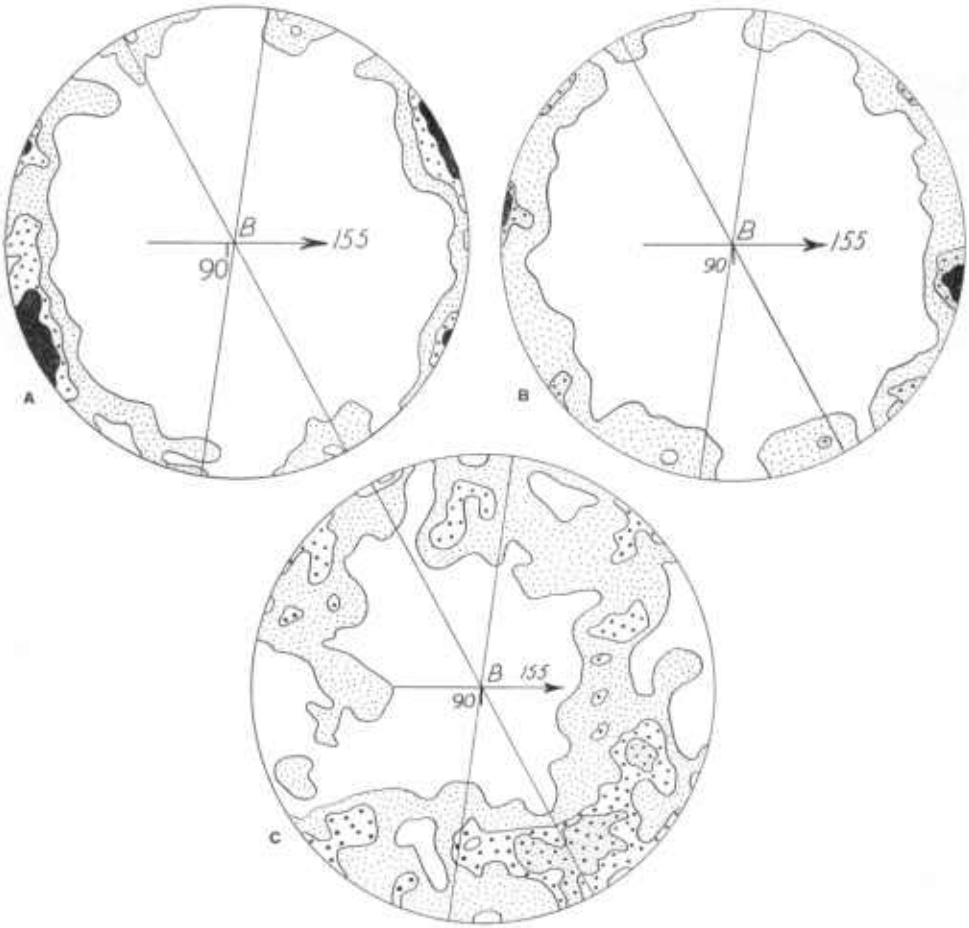


FIGURE 55. Orientation diagram of mica and quartz (after Stefansson, 1943). Locality: Old Frederick Road 3 miles Northeast of Mayfield. Baltimore gneiss, north end of Clarksville dome.

A. 100 biotite poles, contours 2-5-8 per cent. B. 100 muscovite poles, contours 2-6-10 per cent. C. 400 quartz axes. All three diagrams show distinct girdle and different foliation maxima for biotite and muscovite.

27 in the Clarksville dome. (*See* Figure 57.) Mica-cleavage poles are unevenly distributed along the periphery of the circle and suggest a girdle and *B* axis, which is also well defined in the quartz diagram. Several cleavage planes at varying angles are suggested.

Figure 57 is a portion of the county maps showing only the outlines of some formations and the foliations as measured in the field. Foliation in the Woodstock dome forms a foliation cupola or dome overturned asymmetrically toward the south and southeast. The Setters north of Alberton is overturned in a sharp fold. The southern domes have become long northeast-trending asymmetrical anticlines with a uniform dip to the west in the western dome and steep dip in the Clarksville dome. It is probably not accidental that the more distinct *S* maxima and foliations occur near the western limb of the domes, whereas the girdles occur in the centers of the domes.

I agree with the concept outlined by Hopson of a rising dome-shaped but asymmetrical diapir which raised the overlying formations—Setters, Cockeysville and Wissahickon—and probably the Sykesville. During doming and associated deformation and metamorphism, the upper layers slipped over one another resulting in a strongly layered

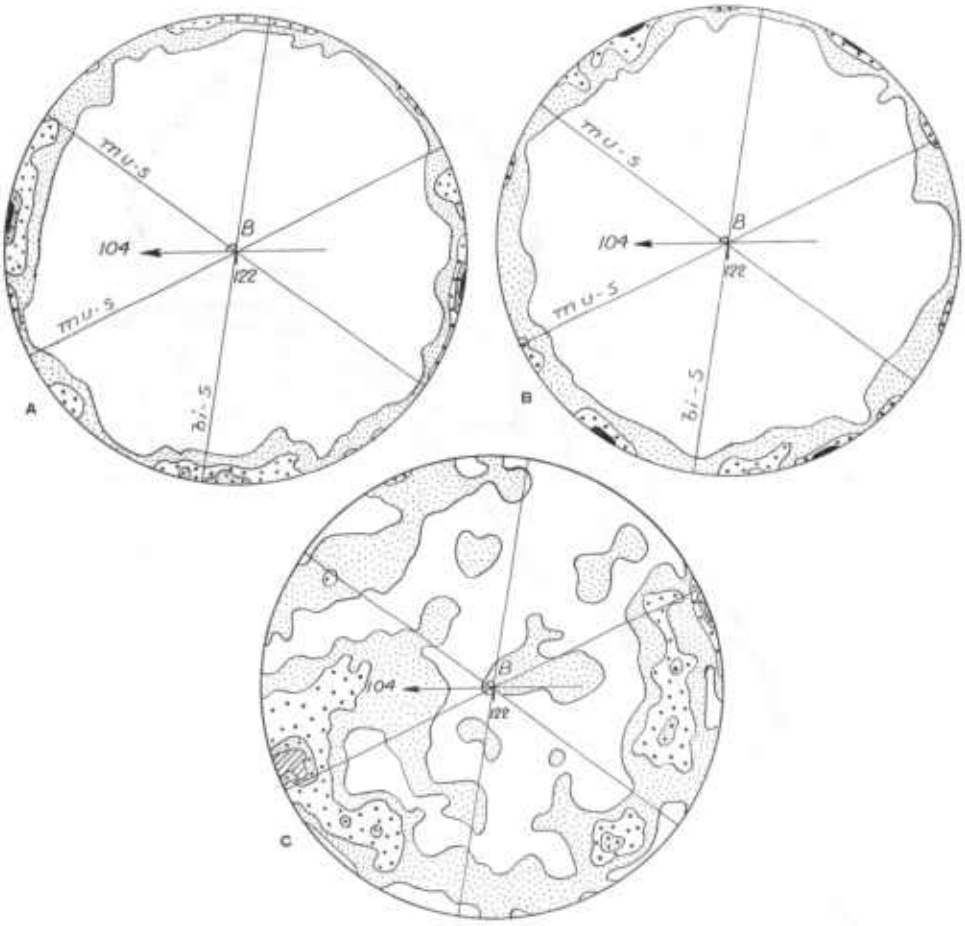


FIGURE 56. Orientation diagrams of mica and quartz (after Stefansson, 1943). Locality: 1½ miles southeast of Clarksville. Baltimore gneiss. Center of Clarksville dome.

A: 100 biotite poles, contours 2-5-8-11 per cent; B: 100 muscovite poles, contours 2-6-10 per cent; C: 400 quartz axes, contours 1-2-3-4 per cent. A girdle is distinct in all three diagrams. Maxima for biotite and muscovite are not identical.

and sharply emphasized foliation (S_2). Toward the center of the axes there was a great deal of crenulation and intersection of foliation planes as expressed in mica girdles and several cleavage maxima.

A similar gradation occurs in the Patapsco River section from northwest to southeast across the Woodstock dome. Here strong S planes are shown within the first mile of the gneiss (horizontal), followed by girdle diagrams, crenulation and intersecting foliation planes in the center and finally, at the overturned lower contact of the gneiss, again sharp S planes. Overturning along the southwest sides is well proven stratigraphically because the Setters dips under the gneiss. The lineations related to this regional picture are discussed below and are shown in Figure 67.

Cleavage in the Glenarm Series

The Setters Formation is usually well bedded and is commercially used for flagstone, demonstrating the smooth bedding planes which are decorated with tourmaline crystals. Parallel with that surface is an ever-present orientation of muscovite which determines

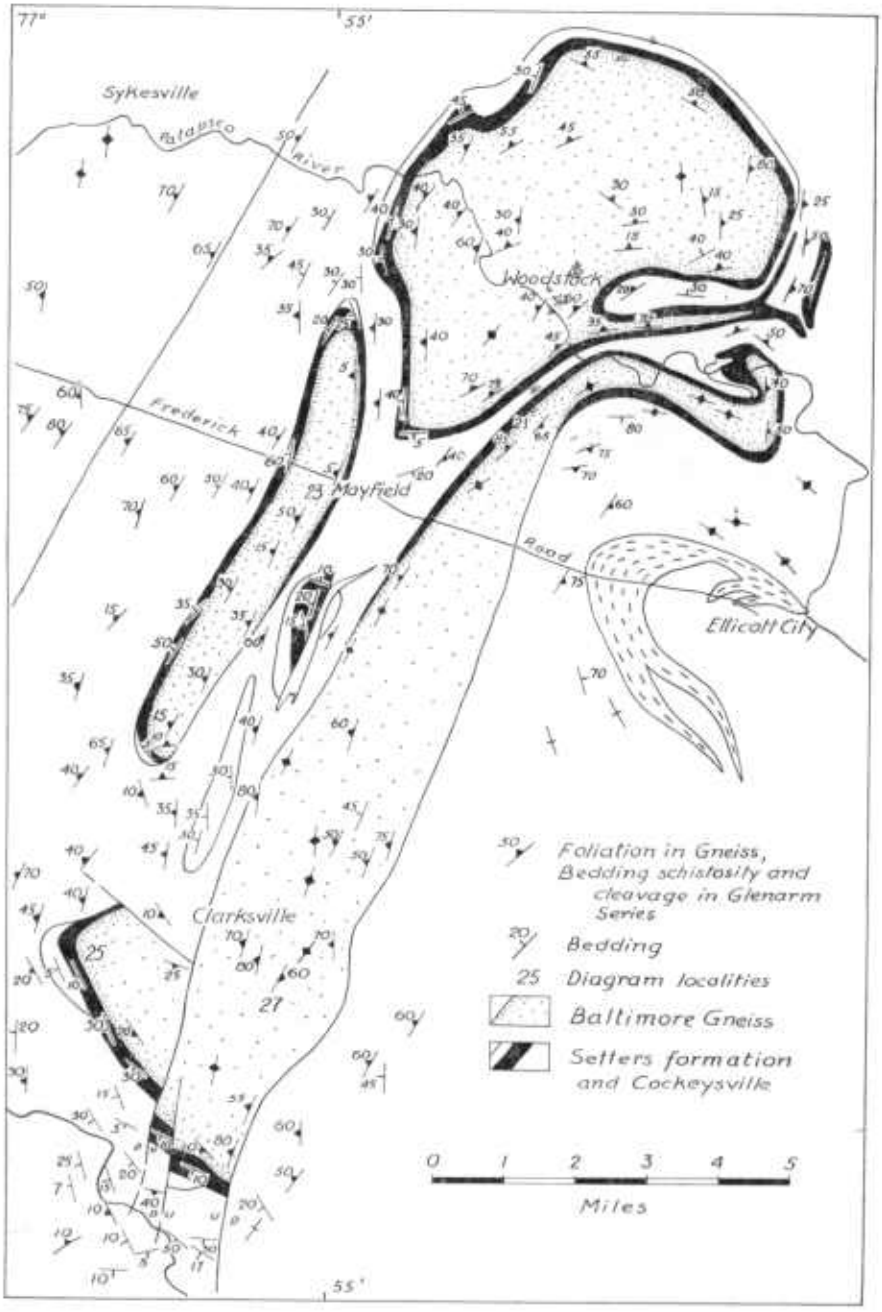


FIGURE 57. Foliation map of gneiss domes in Howard and Montgomery Counties. The domes are asymmetrically overturned to the east or southeast as shown by the constant dip of foliations to the west or north. The Clarksville dome is partially faulted, and only its northeast and southern portions show the Setters above the gneiss in normal contact. The Mayfield dome is an elongate asymmetrical anticline. The Woodstock domes have risen upward and southward over the adjacent Setters, Cockeysville and Wissahickon Formations.

the bedding-plane schistosity (S_2) and makes the bedding planes a sparkling surface with innumerable light-reflecting mica crystals. Where used in patios or similar paving, this reflection may be intense enough to be annoying on a bright summer morning or evening when the sun is low.

The structures of the Cockeysville Marble have been described by Choquette (1960, p. 1027–1052) for Baltimore County with only little reference to Howard and Montgomery Counties. Additional information is supplied by Hopson.

With few exceptions, the formation shows compositional layering which is probably the original bedding. Since the carbonate rocks are more readily deformed, bedding is at many places intensely crenulated, distorted, and intensely folded. Many of these folds resemble flowage folds without distinct constant symmetries or axial directions. They can be seen in many buildings where the marble has been used as building stone (Fig. 58).

In larger exposures, bedding is distinct as coarse layering paralleled by a bedding schistosity. "In places, however, the metamorphic foliation clearly transects bedding" (Choquette, 1960, p. 1035). At other places bedding is disrupted and obliterated by an axial-plane cleavage (S_3).

Bedding-plane schistosity and cleavage (S_2 and S_3) are the prominent structures of the Wissahickon Formation. In the eastern half of the counties, where metamorphism is most intense and the grains are larger, bedding is paralleled by crystallized layers of mica (S_2) but is also cut by a prominent cleavage (S_3) which may obliterate bedding in incompetent layers.

To the west the formation grades into phyllite, the grains are much smaller, cleavage surfaces are smoother and closely spaced, but bedding becomes more frequent, less obscure and at times dominant. At some localities where the Wissahickon Formation becomes an alternation of sandy and phyllite beds, a transecting cleavage (S_3) is limited to the phyllite. The gradation from east to west is clearly visible at Bear Island and along the Potomac River between Washington and the mouth of Seneca Creek, and the intensity of cleavage (S_2 and S_3) is closely related to the composition of the formation the less competent layers show the better cleavage.

Figure 1 of Plate 36 shows well-bedded Wissahickon schist below Anglers Inn, Potomac River, Montgomery County. Sandy and micaceous beds (S_1) are beautifully folded, or crenulated in the less competent layers, and an axial-plane cleavage (S_3) cuts across bedding and is filled by quartz veins, stringers, or rows of lenses. In the photograph, only the micaceous layers show the cleavage well; the lighter sandy beds appear mostly fractured (upper right). Micaceous layers have in some places "intruded" the disrupted sandy ones in the direction of the axial-plane cleavage. Under the microscope a cleavage can be seen also in the more competent layers as a quartz-grain orientation and parallelism of small mica flakes.

A similar example is shown in Plate 39. A laminated mica schist is intensely folded. Plate 39 is a photomicrograph of a thin section perpendicular to the fold and crenulation axes. Bedding (S_1) is very distinct as layers of quartz-rich and micaceous layers. Two cleavages are indicated by mica orientations: bedding-plane schistosity (S_2) and a transecting axial-plane cleavage (S_3). Many small micas are parallel with bedding and follow around the fold. Larger grains cut across and are especially concentrated in the lower half of the photograph where movement seems to have been more intense and where the lower limb of the fold is partly sheared out. In the upper half where the beds are merely flexed the mica is much more distinctly oriented parallel with bedding. In

FIGURE 58. Tracing of folds of Cockeysville Marble. North entrance to Latrobe Hall, Johns Hopkins University. The folds were traced and then reduced and redrafted.

Mica-rich layers show primary bedding. An axial-plane cleavage (S_3) is visible in these layers, and an additional widely spaced cleavage is mostly in the axial plane or near it. Mineralization follows this latter cleavage.

the fold crest mica orientation is more confused; it is partly parallel to S_1 and partly in S_3, or at any direction between the two.

Bedding is lacking in the Sykesville and Laurel Formations except for a vague suggestion of zones with more pebbles or fragments or with more micaceous inclusions. It is impossible to measure bedding attitudes with certainty. On the Howard County map there are still symbols for primary orientations which should be changed because the foliation is secondary.

A foliation due to orientation of lenticular fragments and mica orientations is common and distinct enough to be seen and measured in most outcrops. The intensity varies greatly from place to place, and the foliation is more distinct on vertical crosscutting surfaces than on horizontal ones.

Figures 1 and 2 of Plate 42 show the mica orientation which makes the foliation visible (*See* also Pl. 23, fig. 2; Pl. 25, figs. 1, 2). Figure 1, of Plate 42 is a view normal to the foliation and vertical, thus also parallel to the lineation; Figure 2 of Plate 42 is a view normal to foliation and lineation. A comparison of the two figures shows that the mica orientation is less rigid in Figure 2 but that the quartz pods are elongated in the direction of the foliation. A muscovite diagram (Fig. 59) of the section of Figure 2 of Plate 42 shows a maximum in the foliation plane but a complete girdle in addition. The muscovite is parallel with the lineation but only parallel in places with the foliation. Chlorite is more rigorously oriented than muscovite.

Foliation wraps around quartz pebbles and inclusions. Four mica diagrams of the Laurel Gneiss published by Chapman (1942, p. 1315–1316) show a distinct foliation and a suggestion for near-horizontal *B* axes trending northeast.

The Ijamsville Phyllite is fine-grained and shows good cleavage. At many localities slate has been quarried for roofing. The cleavage is at many places a "slaty cleavage" and generally an axial-plane cleavage; it is equivalent to S_3. I have not seen a strong bedding schistosity.

Harpers Formation is shown only in the extreme western portion of Montgomery County next to Sugarloaf anticline (Frederick County). These phyllites show only one very prominent axial-plane cleavage which is probably the equivalent of S_3 in the Wissahickon sequences to the east. It cuts across bedding as shown in Figure 2 of plate 38.

OTHER CLEAVAGES

General Remarks

At many places and especially in highly micaceous rocks an additional cleavage surface transects bedding, bedding schistosity (S_2), and axial-plane or flow cleavage (S_3). In highly schistose rocks such as the Baltimore Gneiss or the entire eastern half of

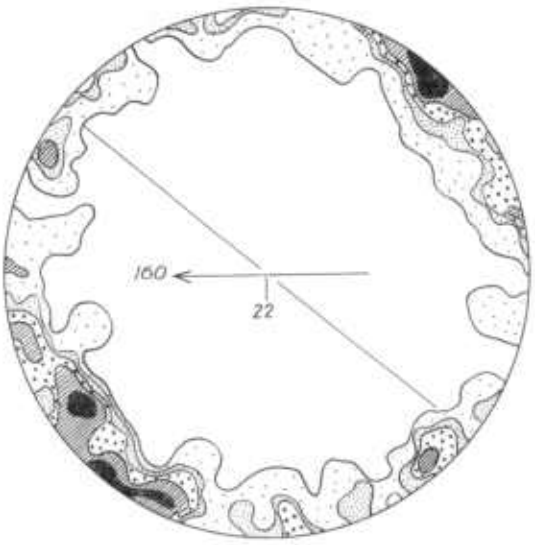


FIGURE 59. Mica diagram of Sykesville formation. Sample shown in Figure 2 of Plate 42; cut normal to foliation and lineation. A good foliation is shown in addition to a complete girdle normal to the photograph. Note that the foliation is not as obvious as in Figure 1 of Plate 42. Contours: 1-2-3-5-(7-9) per cent.

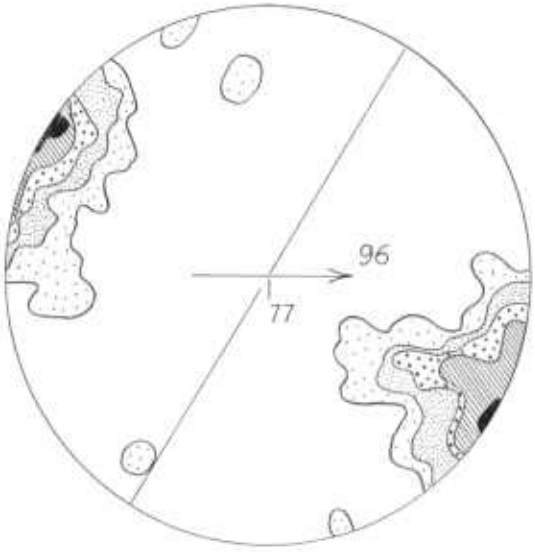


FIGURE 60. Mica diagram normal to that of Figure 59 of same sample. The girdle of Figure 59 is only faintly suggested by lobe of contours toward the center. The center is a blank spot where mica cleavages cannot be measured. The broadening of the maximum along the periphery is not a girdle but is due to mica orientation around large quartz grains. Contours: 1-3-5-(7-10) per cent.

the counties these surfaces are less prominent than they are west of the Sykesville Formation. A similar cleavage has been frequently described, and we have, somewhat loosely, called it fracture cleavage. It is rarely followed by extensive recrystallization.

It is easily seen as a lineation on S_1, S_2, or S_3, particularly in phyllite or slates, and has been called slip cleavage, fracture cleavage, strain-slip cleavage, or pseudocleavage because it is a widely observed structure but obviously less prominent than the earlier cleavages. A summary of cleavage nomenclature and a description of the many varieties has been given by Stevens (1959).

Appearance of S_4 and Microshears

As a rule S_4 cuts older surfaces at large angles and almost always as a system of several planes between which the older cleavage is rotated and broken. In its simplest form the earlier planes are kinked; they may also be bent, and S_4 may grade into the position of an axial plane of small sharp folds. Figure 61 shows the principle involved.

The phenomenon is very well known and has been described from all parts of the world. It is pictured by Lindström (1961, p. 55) from northern Sweden, by Wilson (1961, p. 461) from Wales, and by Albert Heim (1878) from Switzerland. It has been called accordion folding (DeSitter, 1956, p. 216–217) and is also well known as chevron folding. More detailed discussions have been given by DeSitter (1956) and by Turner and Weiss (1963).

Stevens found additional cleavage surfaces in Maryland which he named microshears (1959, p. 85). These are usually microscopic conjugate shear planes which become visible only with proper illumination when the planes stand out in relief. There can be three pairs of these surfaces which intersect in the *a*, *b*, and *c* axes of the co-ordinate system.

Although not shown on the county maps, S_4 is very common in Howard and Montgomery Counties, and microshears also occur but were not systematically studied.

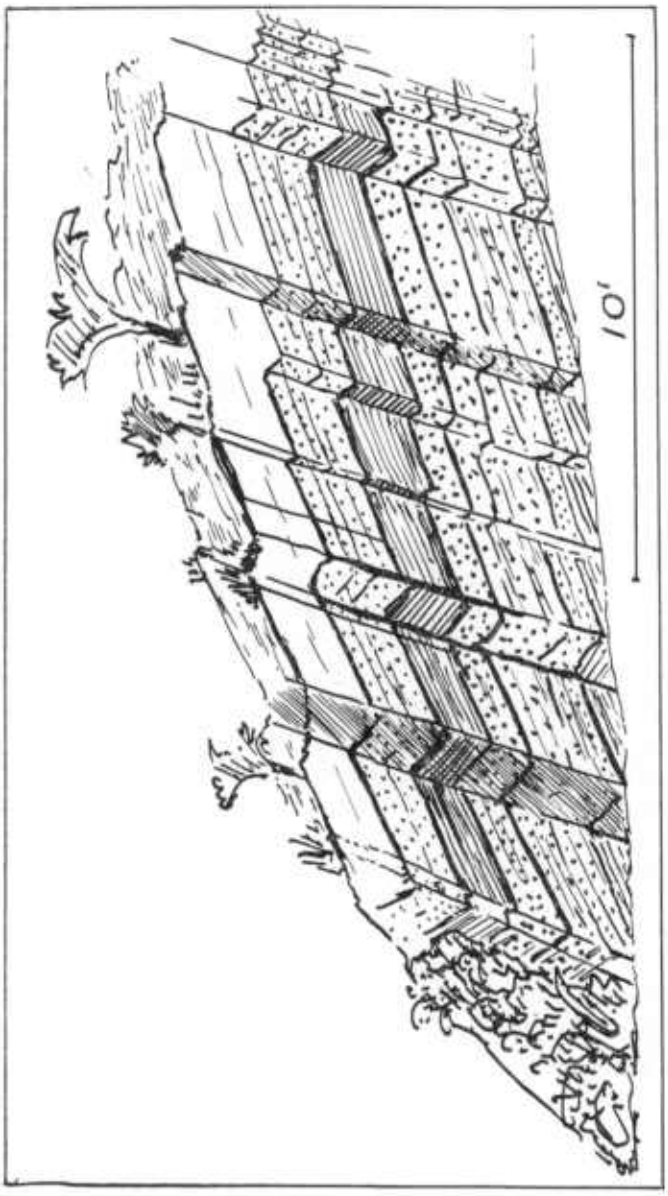


FIGURE 61. Sketch of "accordion" folding or "fracture cleavage" of S_4. South of Hoffmanns Mill, Carroll County. Bedding and S_2 are kinked in zones up to 1 foot wide and between surfaces S_1 and S_2 dip in the opposite direction. Height of outcrop about 10 feet.

SUMMARY ON CLEAVAGES

In Howard and Montgomery Counties all rocks except the youngest intrusions are schistose. The sediments show bedding-plane schistosity (S_2) which becomes less distinct toward the western extremity of Montgomery County. An axial-plane cleavage (S_3) is common and is accompanied by mineral orientation that overwhelms S_2 at many fold crests; this is the most obvious and widespread cleavage. In the Sykesville and Laurel Formations it is the only one.

A crosscutting strain-slip cleavage (S_4) is very common with little mineral orientation and accompanied by accordion folding. Still later microshears occur also and are as a rule conjugate systems which are probably related to deformation and to the same deformational symmetry as the folds and associated cleavages.

LINEATIONS

General Remarks

Lineation has been defined (Cloos, 1946, p. 1) as any kind of linear structure, without regard for origin or orientation with respect to other structures. So broad a term must be more precisely defined because there are many different kinds of linear structures.

A review of the literature furnishes a multitude of terms for linear structures. For example Wilson (1961) mentions the following lineations: fold axes, *b* lineations, *B*, rucking, corrugations, stretching, rolling, drag folds, mullions, rods, intersections of planes, mineral orientations, slicing of beds into strips parallel to *b*, and boudinage. Johnson (1957; 1960) found the following lineations in Scotland: striation, lineation, linear alignment, rods, small folds, corrugations, mineral orientations, axes of microfolds, intersections of planes, slickensides, mullions and *b* axes. Kizaki (1962) describes lineations from a small island in the Antarctic as follows: axes of undulations, fold axes, mineral orientations, *B* equal *b* parallel fabric axes, and *a* lineations. Lindström (1961) found the following lineations in Sweden: β (intersection of planes), small folds, phyllite wrinkles, striations, stretched pebbles, mullions, rods, mineral parallelism and fold axes.

The list of examples could be extended indefinitely. Some authors do not describe observed lineations but label them with reference to a tectonic co-ordinate system. This is an orderly method which, however, deprives the reader of knowing what has been seen. If for instance crenulations on a surface are described it matters little whether they are called small folds, corrugations, axes of microfolds, fine puckerings, Fältelung, Rillung, undulations, or phyllite wrinkles because most geologists are familiar with the phenomena and have observed them. If, on the other hand, the direction is simply called *b* or *B*, no reader knows what has been seen and must accept the author's interpretation as valid.

It would seem most desirable to describe the phenomena in some detail, to illustrate them, and then to interpret them in terms of a co-ordinate system.

List of Lineations

Most of the crystalline rocks show lineations. The structure varies greatly with intensity of deformation, its orientation within the deformation plan, the presence of certain minerals or inclusions, and the amount of recrystallization. Lineations can be seen in almost every outcrop.

The following lineations have been observed:

A. Fold axes
B. Crenulation axes, "crenulation lineation," puckering axes", etc.
C. Intersection of surfaces such as bedding (S_1) or bedding schistosity (S_2) and a cleavage (S_3, S_4, S_5)
D. Rodding, rolling, elongate bodies such as vein quartz in hinges of folds
E. Mineral orientations, single minerals or clusters (Baltimore Gneiss, Sykesville Formation, Ijamsville)
F. Orientation of fragments or "inclusions" in rows, streaks, or trains
G. Microscopic lineations.

DESCRIPTION OF LINEATIONS

General Remarks

A detailed description of lineations is here given because some lineations occur only in one rock type or formation, whereas others occur only in one constant relation to the deformation pattern. For instance, elongate blebs in volcanic rocks are almost without exception down-dip of cleavage planes in the a direction. On the other hand crenulations, intersections between S_1, S_2, and S_3, invariably parallel fold axes and the $b = B$ co-ordinate. Furthermore, the lineation patterns of Howard and Montgomery Counties may contribute to the widely discussed question of whether all lineations are B axes or whether some may occur in a.

Fold Axes

Fold axes are common in all layered or foliated rocks such as the Baltimore Gneiss; Setters, Cockeysville, and Wissahickon Formations; Ijamsville and Harpers Phyllites. They are absent in the Sykesville and Laurel Formations, because these are not layered rocks. The fold axes parallel the general trend of the area and are logically chosen as the B co-ordinate of the deformation plan.

Crenulation Axes

Crenulations are essentially small folds and occur mostly in fine-grained rocks. Cleavage or bedding surfaces may be covered with silky crenulations. They may be due to cleavage intersections or may occur without a second cleavage. They have been observed by many and called small folds, Runzelung, crenulations, wrinkles, and other names. Figure 1 of Plate 40 shows a cleavage surface covered with striations parallel with the scale. In coarser-grained rocks such small crenulations may be less obvious, as in Figure 2 of Plate 40, which is a cleavage surface in a fine-grained quartzite.

Crenulated Wissahickon schist is well exposed west of Gaithersburg along Seneca Creek, at Bear Island and in the Potomac River gorge below, and at innumerable other localities in both counties. Almost all crenulations parallel the fold axes or B direction.

Cleavage Intersections

The intersection of two cleavage planes is a line which may be barely visible or simply constructed from the attitudes of two planes that were measured in the field. Such intersections are very often visible in outcrops, particularly if they are emphasized by slight displacements on one or both planes. They can be observed everywhere and grade into crenulations and small folds. In Howard and Montgomery Counties they

occur in crests of folds or at all places where S_1 or S_2 are cut by S_3. They have been seen in the Baltimore Gneiss; Setters, Cockeysville, Wissahickon, Ijamsville and Harpers Formations, but not in the Sykesville and Laurel Formations.

This lineation is also parallel to the fold axes B if the intersection is due to S_1 or S_2 and S_3. Fracture cleavage S_4 also intersects S_1, S_2, and S_3 in direction B but not as consistently. The other S planes do not intersect parallel to B but rather in other directions and are much less conspicuous. It has become customary to call intersections of surfaces β and the lineation thus produced a β lineation. In this case β is due to intersection of S_1, S_2, and S_3 and is parallel to B.

Rodding

The classical locality for rods and rodding is at Benthutig in northern Scotland (Peach and Horne, 1907, p. 603; Wilson, 1953, p. 131–138; 1961, p. 514), where quartz has filled the hinges of small folds forming rods up to 60 cm long and 1 cm in diameter. "Some of the dip-surfaces of exposure look as if they were covered with a layer of parallel white bamboo canes" (Wilson, 1961, p. 575).

The rods in Montgomery County are not so spectacular but are just as well formed and, at a smaller scale, just as obvious. Figures 1 and 2 of Plate 41 show rods picked from the Wissahickon schist of Seneca Creek, Montgomery County. Here beds in the schist have been folded into disconnected rolls which are the remaining crests of small folds cast in vein quartz. At the ends the quartz rods become thin and spindle-shaped. Such rods can be seen along Seneca Creek south of Seneca and near Boyds, Gaithersburg, and Damascus. They are therefore fairly well distributed and quite common.

Rods are invariably parallel to fold axes B and β.

Mineral Orientations

Mineral orientations can be determined readily by measuring orientations of quartz and especially mica. Mineral parallelism is accompanied by fold axes, crenulations, or rods, which make the lineation more obvious. Mineral orientations alone can also indicate linear directions either as single elongate minerals like tourmaline or hornblende, or as elongate mica grains, but more often as groups of minerals, clusters, and patches. Such a lineation is less obvious and generally must be searched for.

The most obvious mineral orientation is the subparallel alignment of chlorite patches in the Catoctin Greenstone or in the purple Ijamsville phyllite east of Sugarloaf, just within Montgomery County. In phyllites, the elongated patches are up to 1 inch long, about ¼ inch wide, and paper thin.

Similar orientations occur in the Sykesville Formation and in the Kensington Quartz Diorite. Patches of biotite, chlorite or quartz form lenticular bodies which are very regularly oriented and are more distinct in the Kensington Quartz Diorite than in the Sykesville Formation.

Figure 1 of Plate 42 shows a photomicrograph of quartz clusters in the Sykesville Formation. The thin section is perpendicular to the cleavage and parallel with the lineation. Quartz spindles, whose length is about 4 times its thickness, consist of about 20–30 slightly undulatory recrystallized quartz grains. In the same slide, micas are strongly aligned in the cleavage planes which flow around the quartz clusters.

In slides normal to the lineation but also perpendicular to the cleavage the quartz clusters appear less elongated in the cleavage, and the micas are faintly aligned (Pl. 42, Fig. 2). The quartz clusters are probably elongated small quartz pebbles. Figure 1 of

Plate 43 shows such pebbles in a conglomerate layer of Wissahickon Formation, and a comparison of the quartz clusters in the Sykesville Formation with the quartz pebbles in the conglomerate at Seneca Creek suggests that the clusters in Sykesville are also pebbles.

Figures 62 and 63 show these pebbles as seen parallel and normal to the lineation. (Compare with Figures 1 and 2 of Plate 42.) In Figure 62 the clusters are lenticular with long drawn-out pointed ends parallel with the cleavage and wrapped in mica. Figure 63 shows a section normal to the lineation, and the clusters are only slightly longer parallel with the cleavage than across it.

A study of quartz orientation shows that clusters are made up of fragments of larger grains. Adjacent areas show almost parallel orientations, and the former grains can be readily reconstructed as shown in Figure 63.

The plots of quartz axes (arrows in Figures 62 and 63) show no maxima that can be related to a deformation plan; they are accidental accumulations of points of original super grains which are on the point of disintegration. One such plot is Figure 64 in which maxima appear where larger grain fragments accumulate accidentally. The orientations of quartz axes are shown in Figures 62 and 63 as arrows with plunge.

A plot of 100 quartz grains (Fig. 64) from "pebbles" shows a rather even distribution of points in the diagram and only a few feeble accumulations due to disintegrating larger grains.

FIGURE 62. Sketch of thin section parallel to lineation in Sykesville Formation (compare Fig. 1 of Pl. 42). Quartz clusters were coarser-grained pebbles which are broken, stretched, and wrapped in mica. Arrows show directions of quartz axes. Average diameter of pebbles is about 1 to 2 cm.

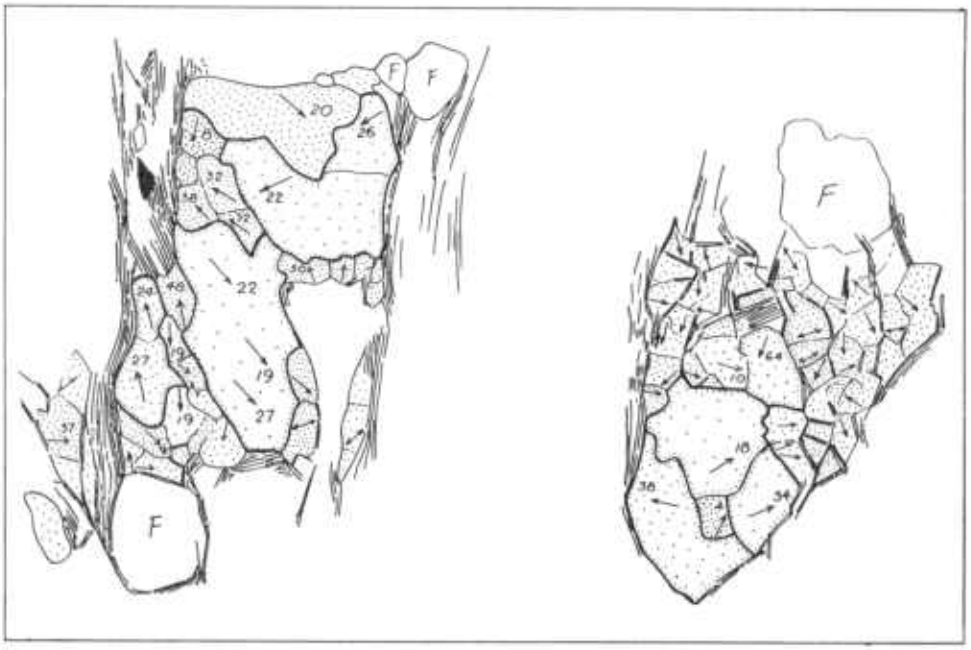


FIGURE 63. Sketch of thin section normal to lineation and Figure 62. Same hand specimen, distance between Figures 62 and 63 about 2 cm.

Quartz pebbles appear almost equidimensional, only slightly elongated parallel to cleavage. Larger grains are shown in equal patterns and surrounded by heavy lines. Arrows are orientations of quartz axes.

The quartz clusters are interpreted as pebbles which were elongated in the direction of the lineation and whose fragmental grains did not become reoriented tectonically.

The alignment of hornblende in amphibolites at Bear Island and in the Baltimore Gabbro Complex results in very distinct lineations at many localities.

Figure 2 of Plate 43 shows a surface normal to an amphibolite lineation at Bear Island. Figure 1 of Plate 44 shows the same specimen in a cut parallel with the lineation. In the former a faint orientation suggests a feebly foliated structure from the upper left to lower right of the photograph; in the latter streaks are obvious and resemble striation. The individual light areas are feldspar clusters between mafic minerals.

Thin sections of the same sample are shown in Figure 2 of Plate 44 and in Figure 1 of Plate 45. The first photomicrograph shows a feeble orientation with large light and dark patches normal to the lineation. The second picture shows the long light and dark streaks parallel with the lineation. Obviously the lineation is due to mineral orientations.

Mineral orientations which result in distinct lineations are common near Cabin John, Chain Bridge, and Bear Island in the Potomac River, along the shores of the Triadelphia Reservoir, Patuxent River, near Olney and west of Brookeville mostly in the Sykesville Formation.

Statistical Analysis of Mineral Orientations

A feeble or questionable lineation may become more obvious if the orientations of many individual minerals such as mica are measured and plotted in a diagram. Such diagrams are Figures 48–51, and 53–56 of mica orientation in cleavages. It is often possible to learn more about the degree of orientation in a diagram than in the field. Figure

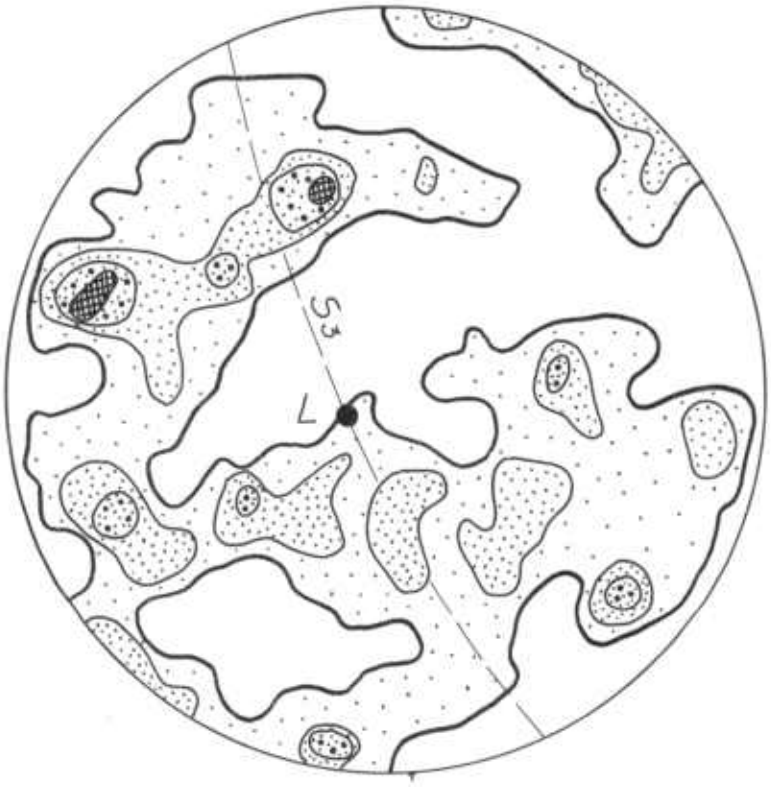


FIGURE 64. Quartz axes orientation diagram of pebble quartz. The maxima are accidental accumulations of grains that have fallen apart but are not newly oriented. S_3 indicates foliation, L is lineation. Same sample as Figures 62 and 63. Contours 1-2-3-4, 300 grains measured. Sykesville Formation east of Rockville.

49 for instance shows an axis in the center of the diagram which coincides with a macroscopic but not very obvious lineation in the Ellicott City Granodiorite. Not far from the granodiorite, in addition to a good cleavage, the Wissahickon Schist shows a suggestion of a girdle and girdle axis.

In the Sykesville Formation a cleavage is shown in the mica diagram (Fig. 59) cut normal to the lineation. The lineation is here the axis of a complete girdle. This lineation is visible in the field as orientation of clusters of mica or quartz and also as the orientation of individual mica grains.

Orientation of Fragments

There is no sharp distinction between mineral orientations and alignment of fragments, and the small pebbles of the Sykesville Formation are included here. Fragments are any pieces of material not identical with the composition, texture, or color of the host rock. They are most abundant in the Sykesville and Laurel Formations. Such inclusions are elongated, spindle-shaped and subparallel in a lineation. Large boulder-shaped fragments of the Sykesville or Laurel do not become oriented; at best they may be sheared at the edges and slightly flattened (Pl. 24, Fig. 1).

Elongated inclusions in the Sykesville Formation are most common near Chain Bridge, at Cabin John, Bear Island, in the Potomac River, and near Brighton, Olney, and Brookville.

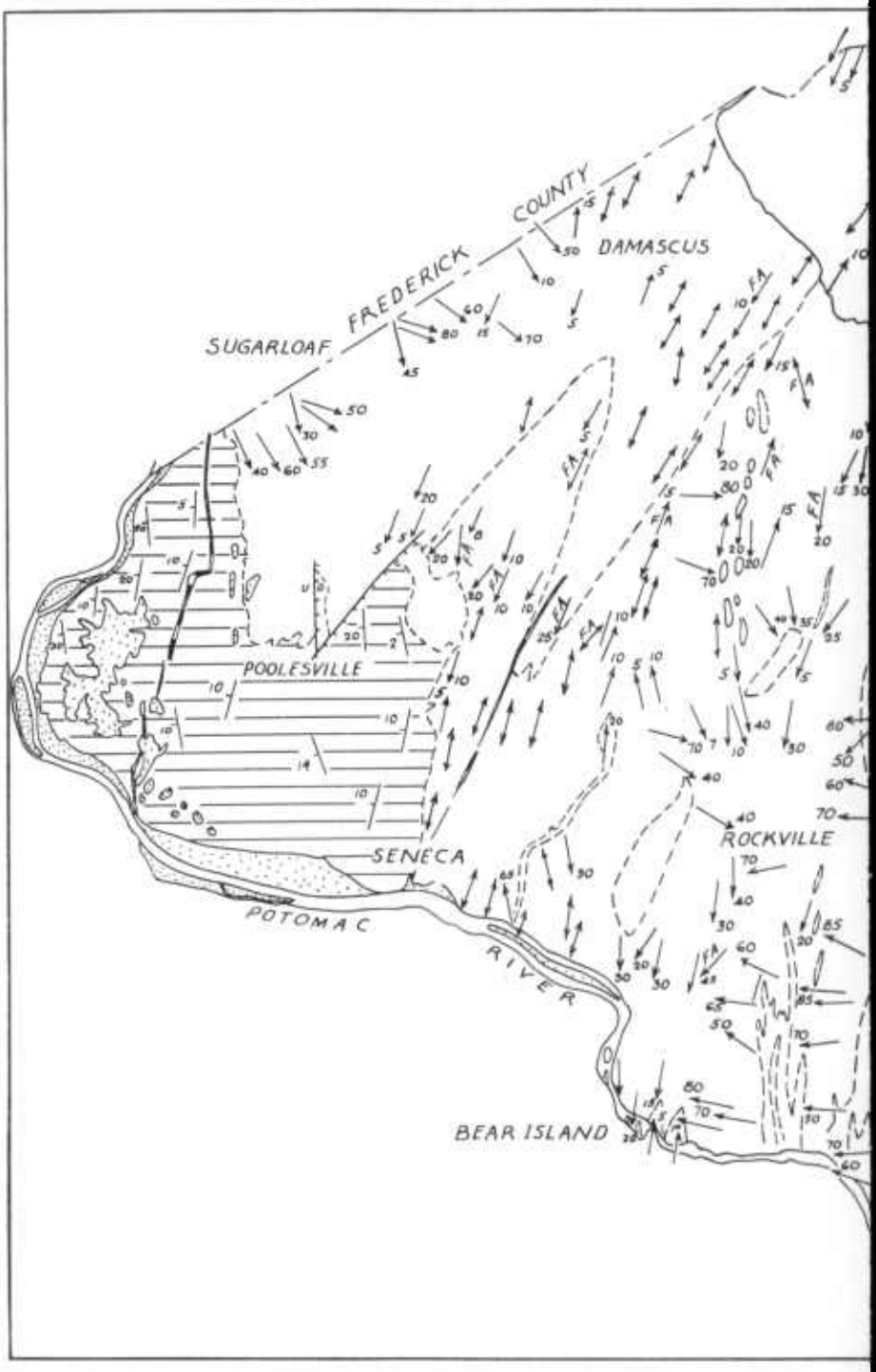
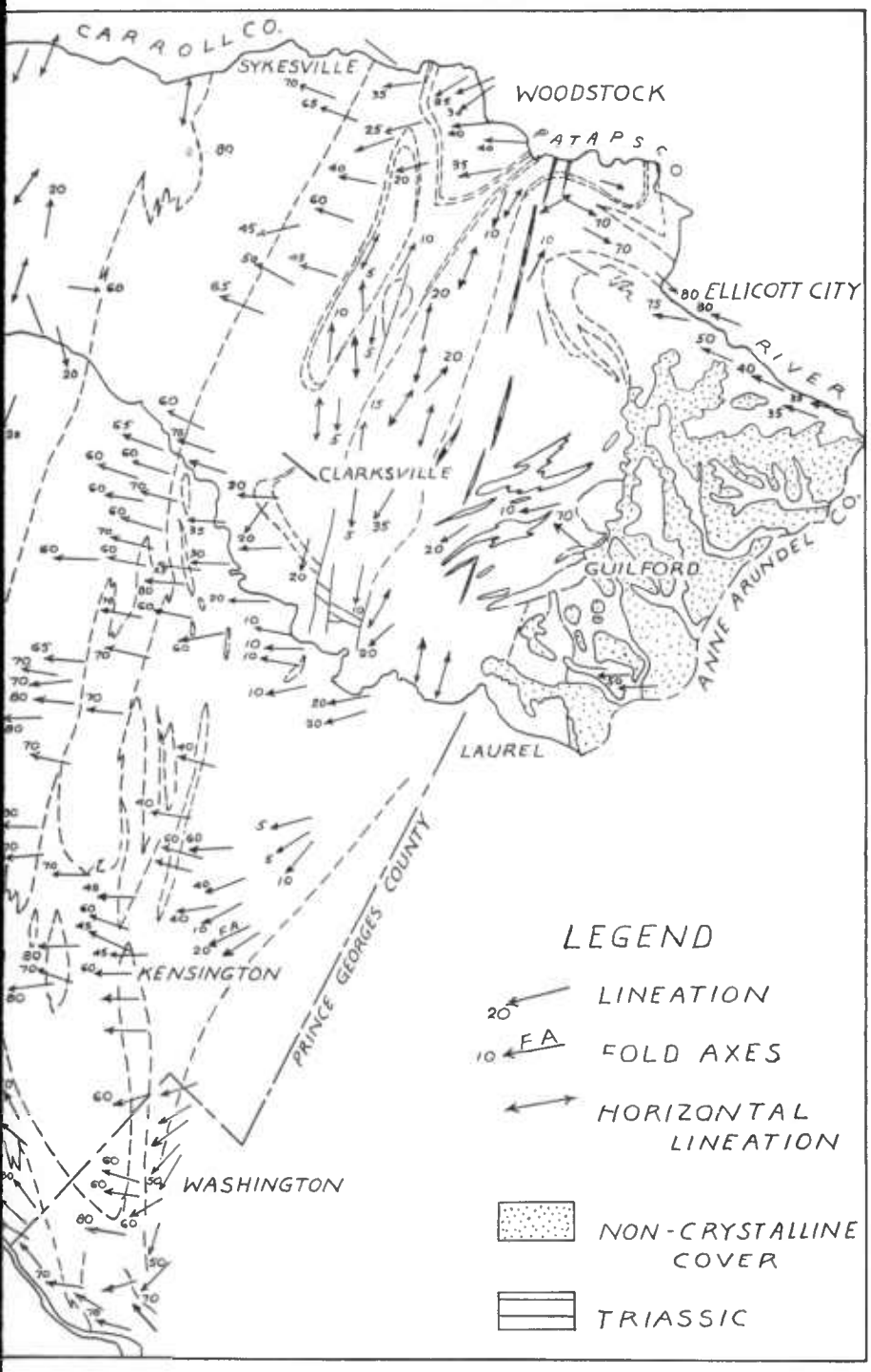


FIGURE 65. Lineation map of Howard and Montgomery Counties


CARROLL CO.
SYKESVILLE
WOODSTOCK
PATAPSCO
ELLICOTT CITY
RIVER
CLARKSVILLE
GUILFORD
ANNE ARUNDEL CO.
LAUREL
PRINCE GEORGES COUNTY
KENSINGTON
WASHINGTON
LEGEND
LINEATION
FOLD AXES
HORIZONTAL LINEATION
NON-CRYSTALLINE COVER
TRIASSIC

Co-ordinates and Lineations

If the fold axes are taken as reference axes and called *B* it is rather simple to describe lineation orientations in terms of this generally recognized co-ordinate. Normal to *B* is the *ac* plane, and *a* is the direction of tectonic transport. Parallel to *B* are then also crenulation axes, cleavage intersections (β) and rodding or rolling of beds. Not parallel to *B* are the mineral orientations and orientations of inclusions in the Ellicott City Granodiorite, the Sykesville Formation, the Laurel Gneiss, and the blebs in some of the phyllites in western Montgomery County. The orientation is not limited to formations but follows a regional pattern.

Regional Pattern of Lineations

The geologic maps of both counties show arrows for lineations. Figure 65 is a summary for easy reference. In order to separate the orientations statistically, the lineations measured in Montgomery County, except those along the Potomac River, were plotted into the lower half of an equal-area net which shows north at the top (Fig. 66). The river area was plotted separately because the exposures are so unusually good there that a disproportionate number of measurements from this one area would overshadow the more general picture, because, with the exception of the river, exposures are reasonably evenly scattered.

The Montgomery County plot (Fig. 66) is based on 323 lineation measurements as follows: 48 fold axes; 150 crenulations, intersections (β) and rodding; and 125 orientations of minerals or inclusions. The results can also be seen in the geological map (Cloos and Cooke, 1953) and in Figure 65. Fold axes trend northeast parallel with the general Appalachian trend. Their plunge ranges from 45° SW. to 40° NE., which is to be expected in an intensely folded area with a number of heterogeneous formations and igneous intrusions.

Parallel with fold axes are crenulations, intersections (β) and rodding and "rolled beds". These lineations are related to the folding process and can be thought of as an intensification of folding up to and including the disruption of bedding and the formation of an axial-plane cleavage.

Completely different are the orientations of minerals and of groups of minerals and inclusions. This lineation is about perpendicular to the fold axes and dips steeply between 90° and 45° normal to the regional trend (Fig. 66).

The orientation alone suggests a different origin from that of folding, bending, rolling, shearing, or intersection of planes. If fold axes are perpendicular to tectonic transport, these lineations are possibly in the direction of transport, or *a*.

Figure 65 shows lineation orientations at 291 localities in Howard and Montgomery Counties. Many more orientations have been measured but cannot be shown because they are crowded in the small, well-exposed areas. A similar map was shown in the upper right-hand corner of the geological map of Montgomery County (Cloos and Cooke, 1953). The map confirms the statistical plot (Fig. 66) but also shows the regional distribution of the two orientations parallel and more or less normal to the fold axes and *B*.

In order to limit different kinds of lineations and their relation to cleavage and bedding several more selective diagrams were reconstructed. In Figure 66 two major orientations are obvious, one parallel to the fold axes and a second one perpendicular to the axes. Figure 68 is a plot of 103 lineation measurements in the Sykesville Formation and 35 in Kensington Quartz Diorite. Seven schist lineations of the adjacent area were added. The rigid

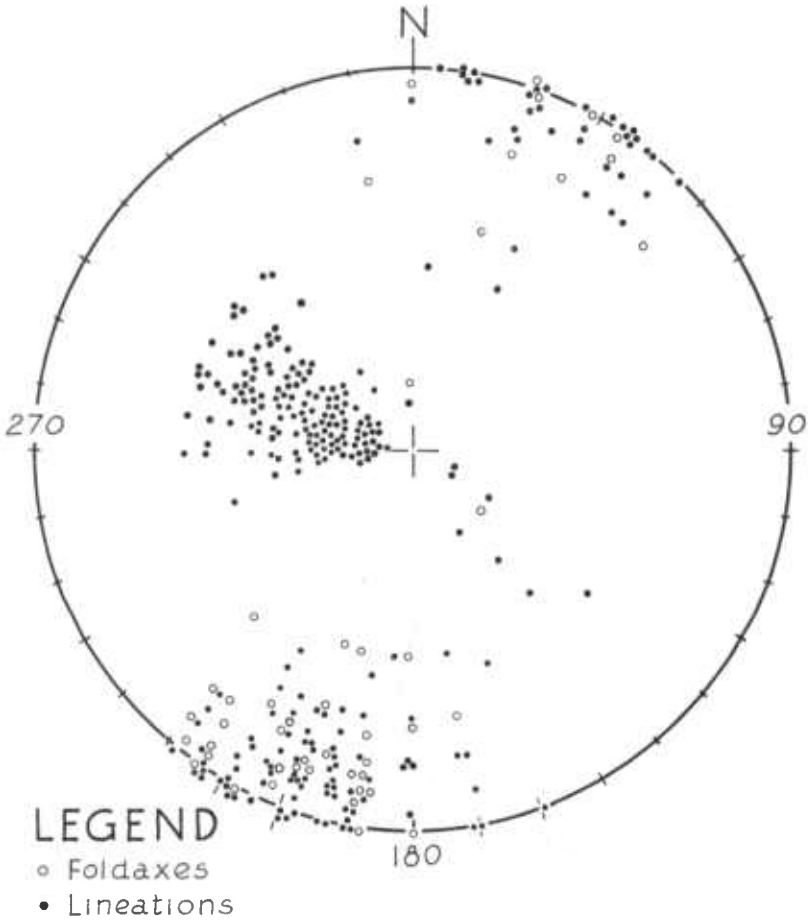


FIGURE 66. Lineations and fold axes in Montgomery County. There are two obvious accumulations of points: steeply northwest and parallel to fold axes. For distributions of orientations see Figure 65.

down-dip orientation is striking even if five Sykesville lineations dip about 50° S. and SW. The Kensington Quartz Diorite is still more regularly lineated.

Figure 69 shows a plot of the Bear Island area, where there is a strong lineation in amphibolite (46 readings), a lineation in schist (17) and fold axes. The fold axes trend northeast-southwest and plunge less than 45°. The amphibolite lineation ranges from almost vertical to horizontal and plunges into the southeast quadrant and south. Schist lineation is essentially parallel to fold axes. There is no west plunge comparable to that in the Sykesville or Kensington areas.

Figures 68 and 69 demonstrate the need for well-defined sampling principles and the selection of homogeneous areas or formations.

The Sykesville lineation is not crenulation or β, there are no fold axes, and it resembles most closely that of the South Mountain volcanic rocks. Its maximum orientation is down-dip of cleavage. In the Potomac River Gorge near Washington, however, the lineation swerves southward. Clearly this area is different from the central belt in Montgomery County. At Bear Island the amphibolite lineation is scattered because the amphibolite is a folded sill with very intense dislocations. However, the lineation in the amphibolite behaves like B axes in the adjacent schist.

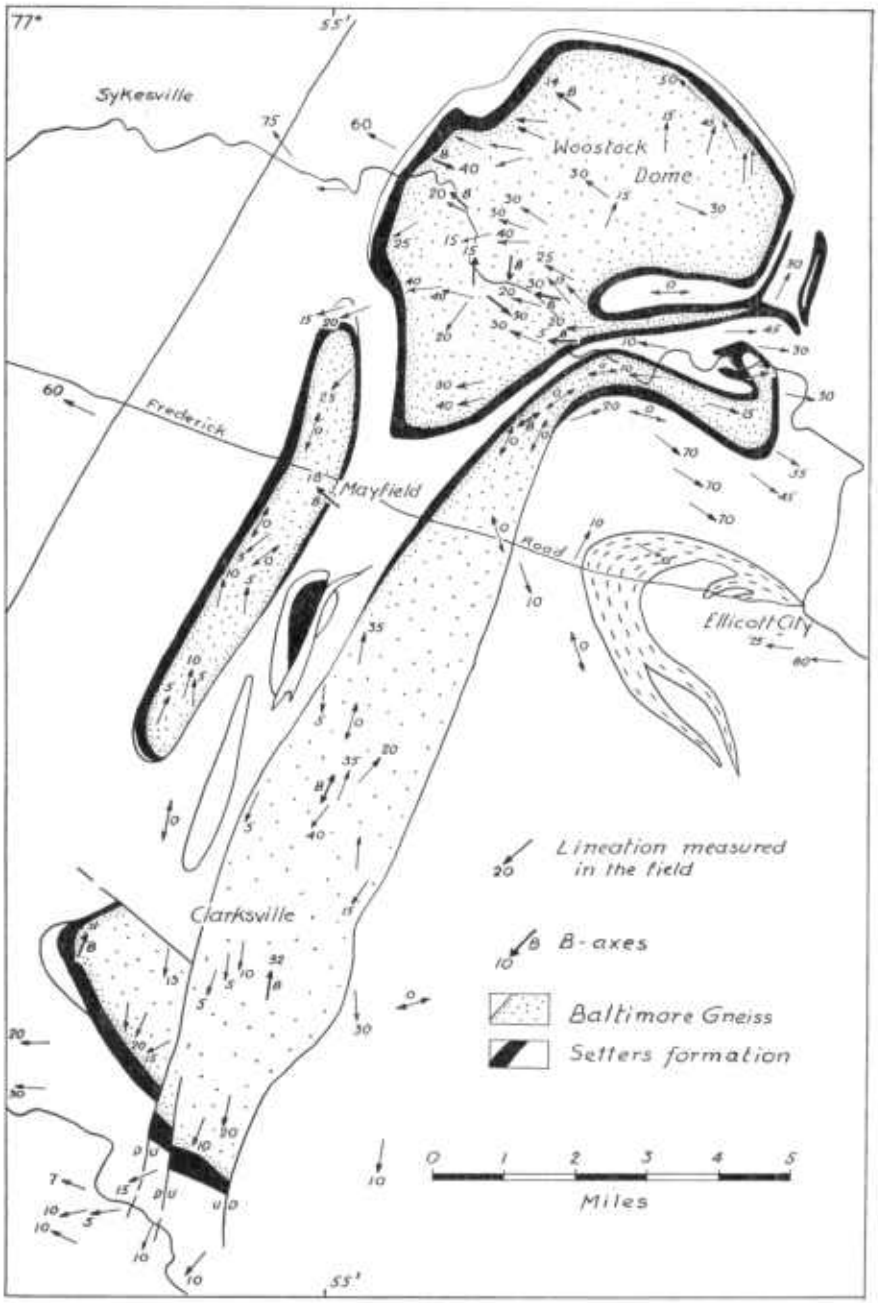


FIGURE 67. Lineation map of the Baltimore Gneiss dome area in Howard County. Black: Setters Formation.

Near the Baltimore Gneiss domes orientations are still more confusing and probably deflected by doming and overturning to the east and south. A far more rigid separation of lineations and more deliberate sampling is necessary before these puzzling orientations can be satisfactorily explained.

In summary: The counties are divided into three parts: an eastern area with steeply dipping, down-dip lineations; a central area with *B* lineations parallel with the regional

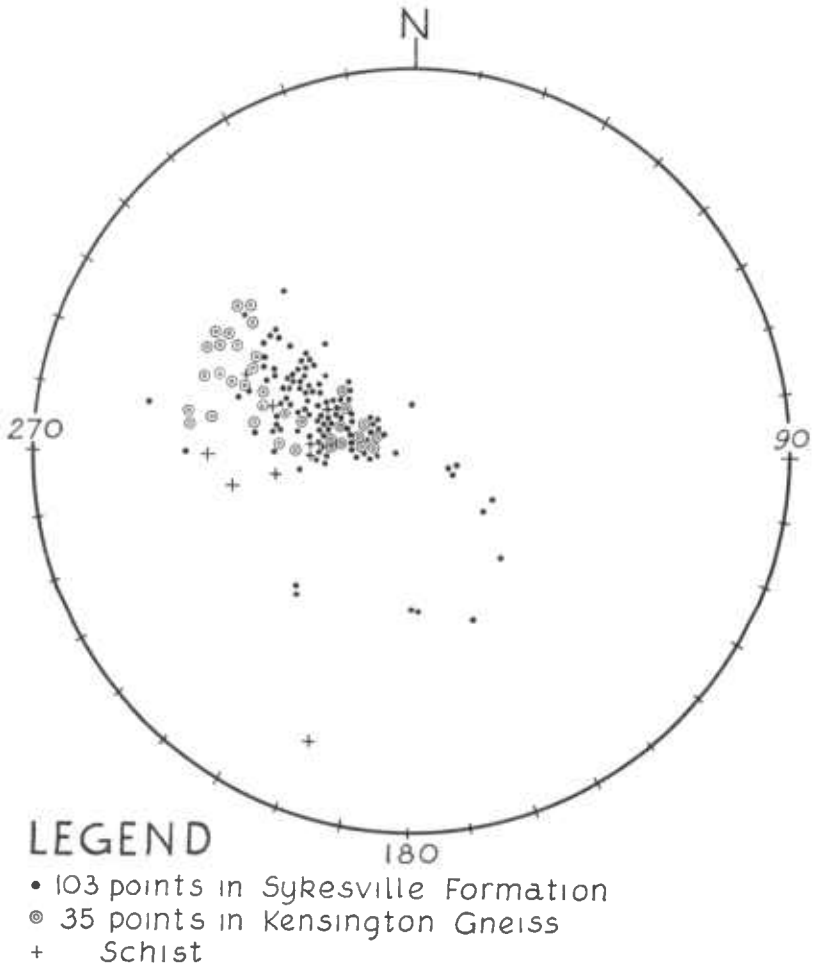


FIGURE 68. Statistical orientation diagram of lineation in Sykesville Formation and Kensington Quartz Diorite.

trend and mostly near horizontal; and along the western border of Montgomery County a narrow zone with down-dip lineations toward the east, also across the fold axes. In the eastern area, gentle lineations occur on the crests of the Baltimore Gneiss domes, where these parallel the general trend of the Baltimore anticlinorium. The break between lineation orientations is also a break in types of lineations observed and almost coincides with the western edge of the Sykesville Formation between Bear Island and 2 miles west of Sykesville along the South Branch of the Patapsco River. To the west *B* lineations occur; to the east mineral orientations, spindle-shaped inclusions and generally *a* lineations prevail.

INTERPRETATION OF LINEATIONS

The Lineation Problem

The problem is very clearly pointed out by Ramsay (1963, p. 146):

"Then developed a controversy concerning lineation: was it in *a*, parallel to the movement direction of the rocks during deformation (views held by Anderson, 1948, 1952; and Kvale, 1953) or in *b*, at right

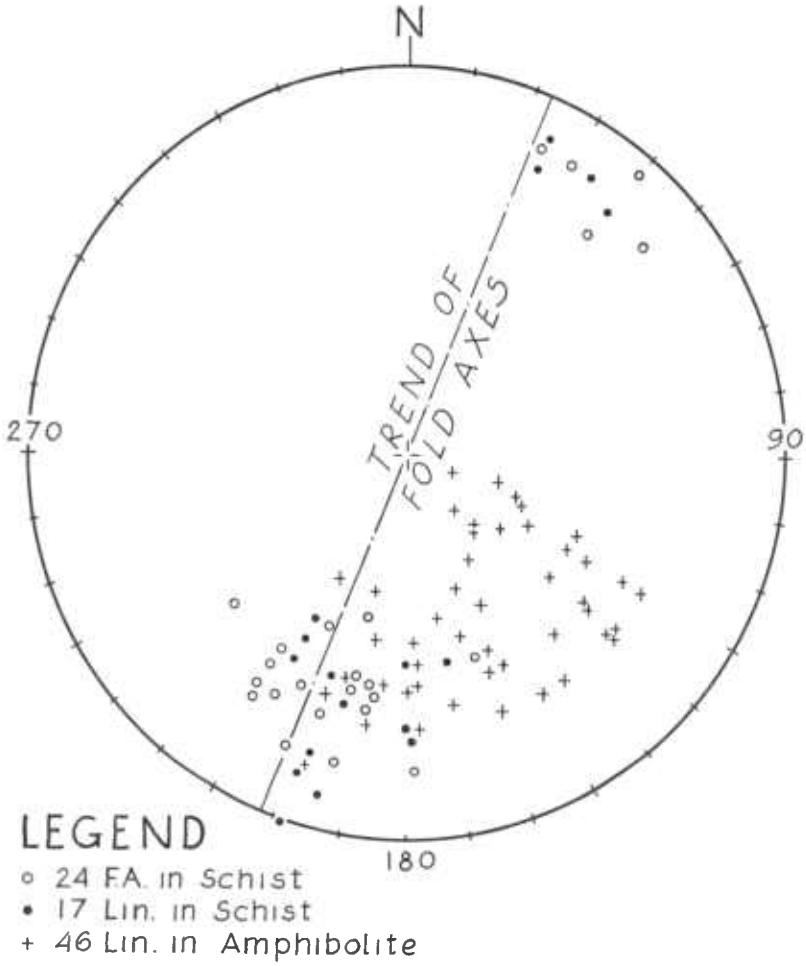


FIGURE 69. Statistical orientation diagram of lineation at Bear Island, in schist and amphibolite

angles to the movement direction, as suggested by Phillips and Wilson? Studies of the way in which early linear structures have been deformed by folds developed during later deformation (Ramsay, 1960) have suggested that the fold axes and the linear structures parallel to them may form at any angle to the movement direction; they may be parallel to *b* or to *a*, what is in fact much more common, in some direction between *a* and *b* in the schistosity plane."

This is not the place for a detailed account of the controversy or the problems involved. The lineations in Howard and Montgomery Counties are, however, apparently within the area of the lineation controversy and may be a good subject for a major detailed investigation and contribution. The problem differs from that of the Highlands of Scotland in that there are few, if any, indications in the Maryland Piedmont of superposition of deformational patterns and distortion of older deformation plans by younger ones. The Maryland pattern is thus less complicated and therefore more suitable for studies.

Interpretation

In the South Mountain anticlinorium a strong lineation parallel to *a* is closely related to the folding process which produced a profusion of fold axes, *B* axes, and crenulations.

The symmetry is monoclinic, and the *ac* plane is the movement and symmetry plane and contains the *a* lineation. (E. Cloos, 1947; 1950; 1951; 1953; 1957; Nickelsen, 1956, p. 258, Fig. 4f.)

A similar deformation plan occurs along the Frederick County line in Montgomery County where fold axes *B* and β are nearly horizontal and a lineation dips steeply to the east down-dip of cleavage planes (S_3).

In the Sykesville Formation a down-dip lineation is either vertical or dips steeply east or west. From here eastward a lineation dips more gently westward in the Wissahickon schist and down to the Baltimore Gneiss (Figs. 65–67). The orientation becomes less uniform, however, in the layered rocks than in the Sykesville Formation. Figure 67 shows the gneiss domes and their mantle. A west-dipping lineation is very regular in the Woodstock dome and south of the Clarksville dome.

In the Mayfield dome lineations and fold axes are parallel and trend northeast. One *B* axis (girdle of quartz diagram, Fig. 53) trends across the fold axes, the long axis of the dome, and parallel with the lineation to the west.

In the Clarksville dome the lineation is parallel with the dome trend, including the right-angle turn northwest of Ellicott City.

The process which resulted in this confused pattern may have been the result of updoming as described by Hopson.

The Baltimore Gneiss domes rose under considerable load, stretching their covers. If this occurred at sufficient depths fractures would be unlikely, but the structures grew by flowage. During the continuing rise of the domes the Sykesville Formation crept down the steepening flank; a strong cleavage and lineation developed. Where intense stretching and slippage occurred a down-dip lineation comparable to slickensides appeared. Where compression and crowding prevailed crenulation, folding, rolling and rodding in *B* occur. Between two such areas there are no sharp borders, and *B* axes may be deflected, bent and rotated into *a*. Such gradational regions seem to be south of the Clarksville dome, at the north end of the Clarksville dome (Fig. 65), and in the general core of the Baltimore anticlinorium where it plunges south between the south end of the Clarksville dome and Washington. Here, the *B* axes which plunge southward south of Clarksville are deflected to the west in the schist above, perhaps because the gneiss dome was moved eastward and the schist axes were dragged along its south edge. Typically, the syntectonic Kensington Quartz Diorite is strongly lineated, rigidly oriented, shows no drag and may have lubricated the entire system, acting as a large "slickenside" during the process.

If the domes had risen symmetrically an *a* lineation would have to point down the southeastern flank in the Laurel Formation and the gabbro complex. This is not the case, and as can be seen at the Patapsco River near Woodstock the domes are asymmetrically overturned to the southeast.

This asymmetry may have caused the prevailing asymmetrical dip of the lineations to the west, even along the east side of the domes where the limbs are overturned.

FAULTS

The geological map of Howard County shows faults only in the vicinity of the Baltimore Gneiss domes. Figures 57 and 67 show faults at the south end of the Clarksville dome and along the east side of the Mayfield dome. At the southern termination of the Clarksville dome the faults can be mapped, but they are not exposed. The east flank of the Clarksville dome lacks the normal cover of Setters and Cockeysville Formations, the foliation is steep on both sides of the contact and the fault is probably also steep.

The east flank of the Mayfield dome seems overturned, and the fault along its east side may well be a reverse fault along which the dome was pushed onto and above the crest of the anticline between the Clarksville and Mayfield domes.

Around the Woodstock dome there must also be faults in the intricate area along the Patapsco River and northeast of Alberton. It was impossible to map faults here, and none are shown. The intense overturning of the southeast flank of the Woodstock dome may well have been accomplished in part at least by reverse faulting.

The Montgomery County map shows faults only where the Triassic is faulted against the crystalline basement. Faults can be easily mapped at Buck Lodge, and to the west the remaining border of the Triassic is an unconformity. Fisher (1963) discusses some faulting in the Potomac River Gorge and maps two small faults.

As a whole faulting is not obvious, probably not easily seen in the field, and it does not seem necessary to assume major faults in the interpretation of the geology.

The relation between the Sugarloaf anticlinorium and the phyllite to the east may require some faulting.

JOINTS

Common Directions

All rocks are jointed, but the joints have not been systematically studied. The following directions are common:

1. Normal to fold axes in the *ac* plane. Attitudes change with the position of the fold axes. Stevens (1959) discussed them as cross joints with the designation S_5. Fisher (1963) has observed three major directions: one perpendicular to fold axes (*ac* joints), one parallel (*bc*) and a nearly horizontal direction.

2. Normal to intense lineations other than fold axes. In the Sykesville and Laurel Formations joints occur normal to that lineation and dip gently to the east or are flat where the lineation is very steep.

3. Parallel to cleavage, foliation or schistosity, in the Baltimore Gneiss parallel to banding. In well-bedded rocks joints are common parallel to bedding.

Bedding joints are mostly taken for granted, and if bedding planes are well exposed, as for instance in massive Setters quartzite, one thinks of the surface as a bedding plane rather than as a joint. Jointing parallel to earlier planar structures is common, however, and not always sufficiently recognized.

4. Diagonal directions are common on a small scale as shown in Figure 2 of Plate 41 in quartz rods or on a much larger scale in Figure 70.

Igneous rocks such as the Ellicott City Granodiorite are well jointed, especially normal to lineation and foliation. Since these structures are steep the joints dip gently.

In the Baltimore Gabbro joints are well displayed in the Patapsco River Gorge, and where lineation is very intense the joints are smooth and closely spaced.

Units such as the diabase sill at Boyds, the serpentinites near Gaithersburg and Hunting Hill or the Guilford Quartz Monzonite are very poorly jointed.

Joints at Annapolis Rock

Annapolis Rock is a large, lenticular quartz vein, about 100 feet or more thick and several hundred feet long parallel to the regional strike of the schist and the common trend of innumerable small quartz veins. It is located just north of the Howard-Montgomery County line on Md 94 (Ellicott Rd.) in western Howard County (Pl. 7). The

quartz which has been quarried in a small opening, is very intensely fractured. The joints are sharp, closely spaced and seem to be arranged in several conspicuous directions. Since the joints seemed somewhat confusing 450 attitudes measured all around the quarry wall were plotted in two statistical diagrams, Figures 70 A and B.

The following prominent directions occur in both diagrams:

1. Northeast-dipping cross joints (*ac*), probably normal to the long axis of the lenticular and possibly spindle-shapped quartz pod.
2. Strike joints parallel to the trend of fold axes (*bc*) and dipping southeast at almost the same angle.
3. Steeply dipping conjugate joints, symmetrically arranged on both sides of the axial trend.

The shape of the quartz body probably influences the joints, which are limited to the quartz pod, but not the regional directions. The shape of the quartz pod is not known except that the longest axis trends about 20°–30° NE.

Discussion of Joints

Joints are late structural elements which transect all earlier ones. They cut across folded bedding or cleavage and are geometrically related to general trends of fold axes and lineations. They are regionally significant and are locally conditioned by differences in competence of formations or shapes of rock bodies such as igneous intrusions or quartz pods. Joint systems may be limited to a diabase dike or granite body. Since they are very numerous and varied they are confusing and have not yet been studied thoroughly enough.

The quartz pod at Annapolis Rock shows regional and local joints well. The regional pattern has frequently been recognized and described. It shows that the deformation pattern of the Piedmont area has not changed radically from the beginning of deformation by folding, through the formation of cleavages, to the appearance of joints which have been formed in larger units as larger structure elements. Fisher (1963, p. 116) describes the joints of the Potomac River gorge as "one of the youngest structures in the area." In many geological terranes they have been recognized as young structures.

DIKES

Dikes are numerous in the two counties and in the entire Piedmont. They occur as somewhat irregular quartz veins, as much larger mappable pegmatites, as granitic dikes in wall rock near granitic plutons such as the Guilford Quartz Monzonite, or as the latest intrusions of Triassic diabase.

Dikes are filled joints, and they can indicate tectonic directions. The pegmatites in the Patapsco River valley below Ellicott City are large, subparallel bodies which seem to fill tension joints perpendicular to the gabbro lineation. At Bear Island granitic dikes are common; some are parallel, and others are normal to fold axes. Still others are very irregular vein fillings.

Quartz veins are overwhelmingly concordant, lenticular bodies that parallel regional cleavage directions both on a small and large scale. In the folded Wissahickon at Anglers Inn (Pl. 36, Fig. 1) they parallel the cleavage; on the Montgomery County map they are shown as large and concordant south of Rockville. Locally they are tension joint fillings along movement surfaces or in fold crests.

The largest pegmatites are shown on the geological county maps, for instance near the Patapsco River or east of Mayfield in Howard County. They are discordant and may

be 100 feet thick; they may be followed along the strike for several miles. As a rule they follow a prominent local tectonic direction and fill tension fractures.

The very prominent Triassic diabase dikes can be traced straight across country for tens of miles. They followed major tectonic fractures long after complete consolidation of the entire Piedmont province and Appalachian folding. They trend N. 10° E. in a swarm which cuts across Howard County from Patuxent River to the Woodstock Dome. There are no deviations reflecting influence of rock composition or structure. This is quite different from the older pegmatites, which are locally controlled. In Montgomery

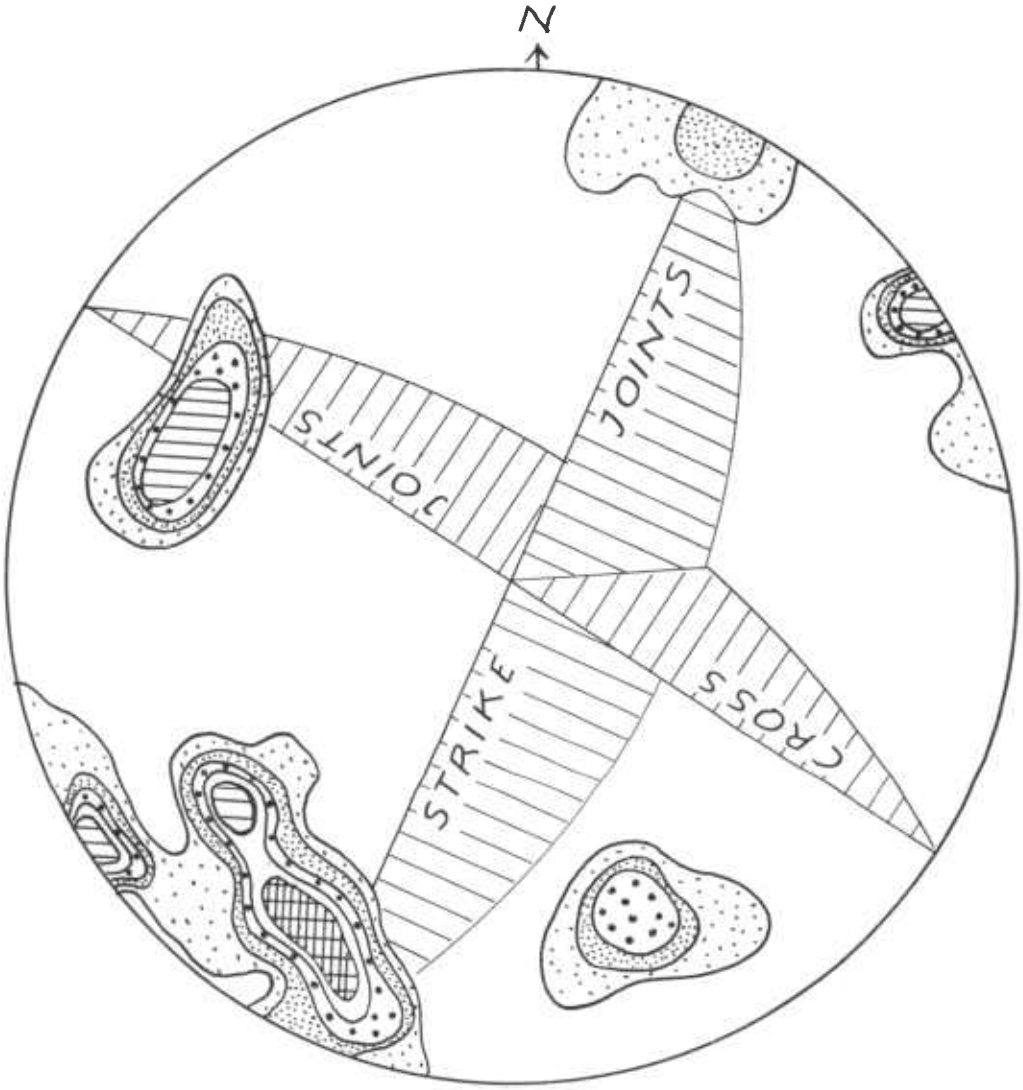


FIGURE 70. Statistical orientation diagrams of joints at Annapolis Rock

A. Southwest half of quarry. 125 joints. Contours 1-3-5-10-(12-20) per cent. Two maxima appear in the poles of cross joints and strike joints. In that area cleavage dips southeast, and fold axes generally southwest. A smaller maximum at the periphery suggests one vertical joint system striking northwest, but a conjugate system does not seem to exist here.

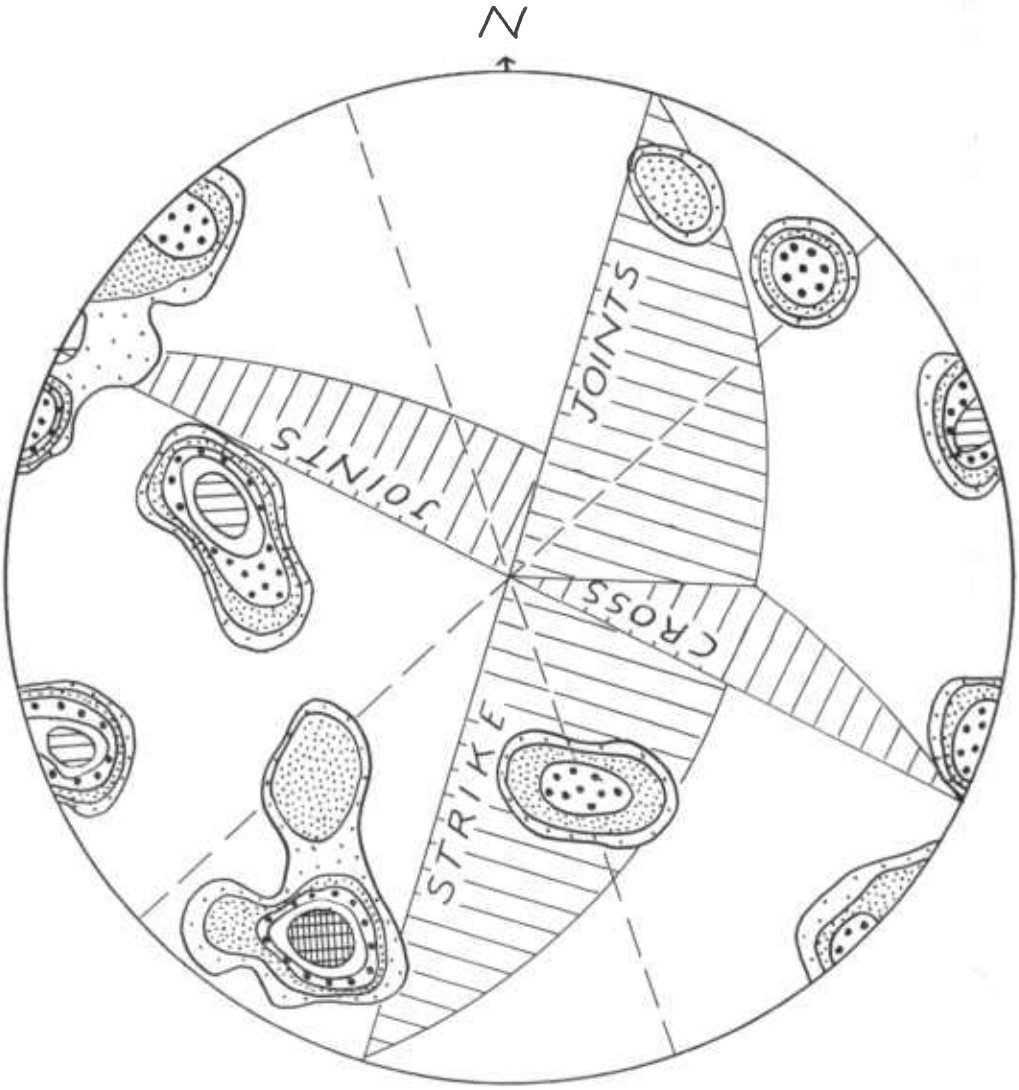


FIG. 70

B. Northeast half of quarry. 330 joints. Contours 1-2-5-8-11-(15-24) per cent. Cross joint and strike joint maxima are dominant in the same position as in A but of slightly different shape. In addition two peripheral maxima show that there are a fair number of conjugate shears at an angle of 66°. They are not symmetrical with respect to cross and strike joints and are possibly later and superimposed directions. They could also be related to the shape of the quartz pod, which is unknown.

County the Triassic diabase dikes also trend almost north-south and cross the western portion of the county disregarding formational boundaries; they are steep tabular bodies filling late regional fractures. These fractures are the latest events recorded in the tectonic history of the counties.

HISTORICAL SUMMARY

The geologic history of the two counties begins far back in the Precambrian, at least 1000 million years ago, when submarine volcanoes may have been active and produced

the material which makes up the Baltimore Gneiss. This original complex was then deeply buried, folded and invaded by heat and granitic juices from below and made over into the varied and colorful sequence we now find in the Gneiss domes.

The early basement complex, of which we know little, was levelled by erosion and buried under the sediments that now make up the Glenarm Series. This stage is illustrated by Hopson's Figure 31. Deposition is, according to Hopson, probably late Precambrian. Correlation with the known Cambro-Ordovician sequence to the west has been attempted but does not seem entirely satisfactory. The Glenarm Series is unlike the Weverton-Antietam and carbonate sequence, and the Weverton has never been found east of Sugarloaf. Similarities between the Glenarm Series and the Martinsburg are rendered inapplicable for correlations because of the difference in radioactive dating. There is also a facies change from west to east as shown by Hopson in addition to an increase in metamorphic intensity. The volcanic rocks in Frederick County are also a problem because they are very similar to and not far from the Catoctin Greenstones but occur within the uppermost Glenarm Series; this alone would tend to put the Glenarm sequence below the basal Cambrian Weverton.

During the basin-filling phase subaquatic slides occurred, and large masses of unsorted debris slid from a rising submarine ridge or scarp and furnished material of the Sykesville and Laurel Formations. Sedimentation continued, and the facies changed from Flysch to shallow-water Molasse-type materials.

When orogeny began, with folding and intrusion of materials from below, the beds were compressed into folds. The Baltimore Gneiss was mobilized, squeezed upward, and carried the Setters and Glenarm rocks on its back. At some places the mobilized gneiss complex broke through its cover as Gunpowder Granite; mostly, however, it stayed below its blanket of Setters. The unconformity above the gneiss was erased as a new foliation formed, mostly parallel to the bedding of the Setters. At some places faulting occurred, and the Setters and Cockeysville layers were stripped off the gneiss.

The intense metamorphism in the gneissic centers grades outward into the overlying formations but generally weakens away from the Gneiss domes. Recrystallization took place along bedding planes, forming a bedding schistosity which is gradually lost in the western phyllites as distance from the Gneiss domes increases. The rocks became coarser in the east, metamorphic minerals like staurolite and kyanite appeared, micas grew larger, and quartz recrystallized in fold crests as rods or in cleavage as lenticular and spindle-shaped masses. The Sykesville Formation, which shows no bedding, became foliated and lineated; the freshly injected Kensington Quartz Diorite became foliated and lineated as it appeared on the scene; and the Ellicott City Granodiorite may also have been drawn into this process in its later stages. The earlier Baltimore Gabbro lost some of its primary structures and assumed a new foliation and intense lineation.

At Bear Island, the gabbroic sills were intensely folded and rolled up into spindle-shaped amphibolite bodies with only occasional massive centers. Small bodies were overwhelmed by the newly forming schistosity, and only the large massive amphibolites retained their primary structures. Beautiful folds appeared everywhere in the Piedmont, and wherever laminated rocks occur they can now be admired and studied. When folding became intense and bedding (S_1) and bedding schistosity (S_2) were tightly compressed a new axial-plane cleavage (S_3) formed, and crystallization, especially of micaceous minerals, followed it.

As the Baltimore Gneiss domes rose, dragging the mantling Glenarm Series with them, the Wissahickon and Sykesville Formations assumed a lineation which runs off the domes like water off an umbrella. Since the domes were also pushed eastward and

southeastward the eastern flanks were plowed under and were overridden by the anticlinal crests. Fold axes (B) were bent, deflected, and dragged into abnormal directions; new slickensided surfaces (a) appeared. As this happened in the eastern half of the county the phyllites and western Wissahickon sequence were crowded and folded along generally northeast-trending axes. Parallel to these are the lineations and a steep axial-plane cleavage. There follows a stage of cleavage (S_4) which includes larger rock units and cuts across the axial plane and flow cleavage used for mica orientation. Deformation became more mechanical; there was kinking of earlier cleavages but generally not intense recrystallization. The main phase of mineral orientation and growth was over, and rocks began to fracture. Late in this stage intrusions began to cut across old structures and emplace themselves discordantly; Bear Island, Guilford or Woodstock Granite are examples. All of these are small, and most of them have almost structureless borders. They cut cleavages and bedding, or schist and amphibolite alike.

The sediments were consolidated then, and fractures opened—first parallel with the general "grain" of the area; these were filled with veins and lenses of quartz, material that was close at hand in overabundance. Later fractures were filled with pegmatites or large dikes of granitic rocks. These dikes show banding but no metamorphism, cleavage or folds. Cleavage surfaces are cut by these dikes.

Still later microshears appeared. These are shear planes which occurred probably at the same time that tension fractures were filled with quartz.

Gradations between all these stages render the picture intricate and complicated because competent units fractured at the same time as micaceous layers were mobilized and intruded. Micas grew large and are well oriented in one layer, but only a few small micas may show the orientation in an overlying bed. One entire formation may be devoid of bedding, and others may be laminated. These differences in physical properties make detailed observation necessary and interpretation and generalization difficult.

After Appalachian deformation ceased and the crystallines had been levelled by erosion, the Triassic formations were laid down on a land surface and thereafter were faulted and tilted. The warping of that surface and its deposits produced extensive fractures which were intruded by diabase dikes; the sill at Boyds used the unconformity between the crystallines and Triassic red beds to emplace itself.

The Triassic phase ended, and the whole area was subjected to erosion again, was levelled and prepared for the deposition of the sand and gravels of the Cretaceous. That era began as a transition from land to water and the massive marine Cretaceous deposits which rest discordantly on the crystalline basement. In these deposits we find the sand and gravel pits for the concrete of our highways and the buried sands which have become the most important water aquifers of the eastern seaboard.

The land has risen since the end of deposition in Miocene or Pleistocene time. The rivers have become entrenched, and their former gravels can still be found in western Montgomery County capping the hills above the Potomac River valley. The Chesapeake Bay and lower Potomac River are now navigable waters up to the Fall Line—so called because of rapids and falls along a line from Richmond, Virginia to Philadelphia. Erosion now furnishes the materials which are transported down the rivers to the Atlantic Ocean where sand bars support cities like Ocean City and cause controversies about their uses.

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MINERAL RESOURCES OF MONTGOMERY AND HOWARD COUNTIES

BY

EMERY T. CLEAVES

INTRODUCTION

The important industrial products found in the area include serpentine, mica schist, marble, gabbro, and sand and gravel. In addition, many small pits and quarries for building stone and sand and gravel have been opened for local use. Deposits which are too small to compete with present-day large-scale operations but are of historical interest include chromite, feldspar, quartz, iron ore, granite and gold. Localities of quarries and mines are shown on Figure 71.

The total value of the Maryland Mineral Industries exceeds 60 million dollars of which Montgomery and Howard Counties make up only a small fraction.

MINERAL DEPOSITS BEING ACTIVELY WORKED

Montgomery County

Serpentine is taken from the Rockville Crushed Stone Company quarries near Rockville and used for concrete aggregate and roadstone. The Stoneyhurst quarries near Bethesda quarry mica schist from the Sykesville Formation. The rock is used for building stone, flagging, and occasionally for riprap.

Sand and gravel has been quarried intermittently, but only for local use since 1955. Potential areas for sand and gravel development include deposits along the Potomac River and Coastal Plain sediments along the County's southeast border. The latter deposits probably will not remain available because of mushrooming housing developments.

The unique Triassic diabase sill near Boyds is an excellent and inexhaustible source of road metal.

Howard County

In the past 10 years intermittent operations have been carried out on a number of mineral deposits. Gravel pits have been operated near Waterloo and Jessup by the Cosca Sand and Gravel Company. The Arundel Corporation has removed sand and gravel from pits near Laurel for paving and road material. The Cockeysville Marble has been quarried near Clarksville by T. D. Nichols. Small quantities of mica and beryl have been taken from the Maryland Mine near Simpsonville by the Green and Taylor Nurseries. The Day Maryland Quartz Company has obtained crude quartz from a pit near Marriottsville for use as a filler medium at its mill in Sykesville. In 1961 the Arundel Corporation was quarrying gabbro near Savage for concrete aggregate.

MINERAL DEPOSITS OF HISTORICAL INTEREST

Chromite

Chromite has been found at two localities in Montgomery County, Etchison and Lyde Griffith property (Fig. 71). Although neither area was mined extensively, the chromite

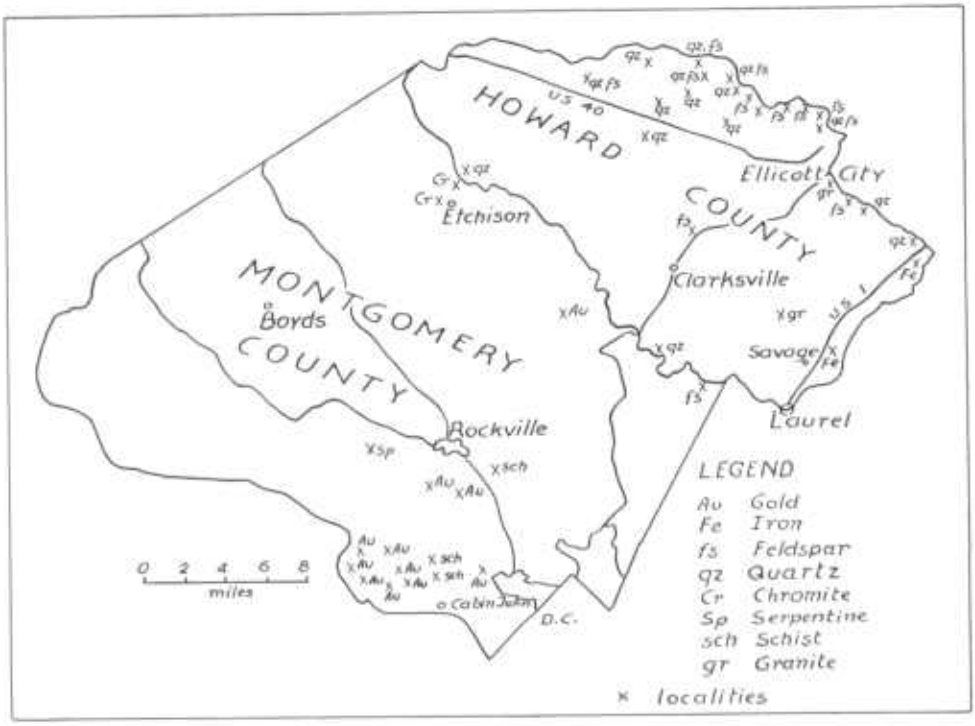

FIGURE 71. Map of Howard and Montgomery Counties showing mine and mineral localities

is of interest as chromite mines in Baltimore, Harford, and Cecil Counties were a major world source in the 1830's and 1840's (Ostrander *et al.*, 1946, p. 15) and supplied a substantial part of the domestic chromite until the 1880's (Pearre and Heyl, 1960, p. 754).

Chromite at Etchison was mined intermittently prior to the Civil War. Apparently three pits were opened (Singewald, 1928, p. 191), but about 1950 the workings were bulldozed over (Pearre and Heyl, 1960, p. 755). Spot samples from what appeared to be an old ore pile were analyzed by Shannon (1926, p. 19) with the following results:

SiO_2	16.42	CaO	1.08
Al_2O_3	17.88	MgO	20.98
Cr_2O_3	20.56	H_2O	6.64
total Fe	17.62		

A small pit was opened on the Lyde Griffith property from which several tons of ore was extracted. The ore was analyzed as follows (Ducatel, 1838, p. 34):

Cr_2O_3	45.5	total Fe	45.0
Al_2O_3	7.5	SiO_2	2.5

GRANITE

The Ellicott City and Guilford Granite in Howard County were once extensively quarried for building stone. Quarries below Ellicott City were probably opened late in the eighteenth century (Mathews, 1898, p. 148) and were worked intermittently until the early twentieth century. The granite, with its characteristic phenocrysts of feldspar, was extensively used for curbing in Baltimore. Quarries near Guilford on the Little Patuxent

River were opened in the Guilford Granite about 1834 (Mathews, 1898, p. 156) and worked until 1860. Operations resumed in 1887 and continued intermittently until the 1920's.

IRON

Both iron carbonate and magnetite were mined in Howard County prior to 1911 (Singewald, 1911). The carbonate ores occurred in clays of the Potomac Group as scattered lumps and nodules of various sizes. Magnetite ore was found in the green chloritic schist of the Wissahickon Formation, and two small mines operated for a short time.

In the nineteenth century two small iron works were built in Howard County. In 1826 a furnace was erected at Elkridge, about half a mile east of Elkridge Station. The works were rebuilt in 1854 and then abandoned in 1872. Two furnaces were built on the north bank of the Little Patuxent River at Savage in 1835. They were operated for a short time, reopened in 1864, and operated until 1874 when the works were abandoned.

FELDSPAR

Numerous feldspar quarries and pits were operated in Howard County; the majority were grouped along the Patapsco River at Marriottsville, Hollifield, and Ilchester (Singewald, 1928, p. 119–129). The feldspar occurs in pegmatitic masses, dikes, and bands genetically related to the granite in the area. The feldspar was crushed and used in stucco finish or ground and used for cleanser, stucco, and composition flooring. Other uses included ceramics and fertilizer.

One quarry operated briefly 4½ miles northwest of Laurel in Montgomery County.

QUARTZ

Vitreous quartz, under the trade name of flint, was quarried on a small scale in Howard County. Seventeen quarries of various sizes once operated (Singewald, 1928, p. 149–153), of which one, the Day Mine, still operates intermittently. The quartz was used for pottery manufacture, abrasives, filler, roofing, stucco, and poultry grit. The quartz occurs as dikes, veins, and lenticular masses of limited longitudinal extent.

GOLD

Gold, in veins of quartz, was first reported in Montgomery County in 1849 on the farm of Samuel Ellicott near Sandy Spring (Justice, 1849, p. 84). Although gold was reported in Maryland as early as 1830 (Am. Jour. Sci., 1830, p. 202) it was not commercially exploited until 1867 when mines in the area of Great Falls were opened. The mines operated intermittently until 1951.

The gold-bearing veins occur in schists and gneisses that have been deeply weathered, at places to depths of 100 feet. Mining operations have remained above the water table and for the most part have been in the weathered rock or saprolite.[1]

The gold occurs in thinly fissile, sericitic schist carrying quartz in small particles and in small and large lenses (Weed, 1905, p. 131). In the lenses the quartz is white, vitreous, and crystalline. The veins generally strike N.-S. and dip almost vertically. In their larger features the veins are regular, but in detail they are irregular and wavy, sometimes reach-

[1] Saprolite is a general name for thoroughly decomposed, earthy, but untransported rock (Becker, 1895, p. 289) in which the feldspars alter to clayey products and the rocks cohere sufficiently to retain most of their textural and structural characteristics (Mertie, 1959, p. 239).

ing thicknesses of 5 feet or wedging out and leaving only a mineralized zone without quartz (Emmons, 1890, p. 406).

Free gold occurs in quartz associated with limonite left by decomposition of pyrite crystals, or in clay selvages (Emmons, p. 407; Weed, 1905, p. 128). The gold also occurs in pyrite. In general the richest ore is coarse, honeycombed quartz, known as hickory bark, and quartzitic schist. Galena and some silver are irregularly distributed along some veins, and tetradymite, a telluride of bismuth, was found in one locality (Emmons, 1890, p. 407).

The "gold rush" began in 1864 (Shosteck, 1953). A regiment of California and Oregon volunteers camped on a plateau behind the Old Angler's Club at Cropley. While washing his skillet in a stream, one of the soldiers, an ex-placer miner, found flakes of gold dust. The resulting search for ore revealed numerous exposures of quartz veins, but the regiment soon moved out. Apparently only one soldier, John Clear, returned. He organized the Maryland Mining Company in 1867, sank a shaft, and struck ore. The vein soon petered out, and the works were abandoned. The mine was briefly reopened prior to 1890 by Watson, a Chicago engineer. In the early 1900's the mine was again opened and further developed.

GOLD LOCALITIES

Rock Run (Shosteck, 1953): In 1874 Parsons discovered gold in the sands of Rock Run. Because of lack of money, prospecting was superficial and soon discontinued.

Montgomery Mine (Ostrander, 1938; Ulke, 1939; Shosteck, 1953): Gold was discovered about 1871, and the mine opened in 1875. In the few years that it operated, about $8000 worth of gold was recovered.

Allerton-Ream Property (Ulke, 1939; Emmons, 1890; Shosteck, 1953): The area was opened in the 1870's and prior to 1890 worked by open cuts. No quartz veins were found, but hickory bark and siliceous bands of pyrite-bearing schist were mined. In 1890 the Ford Mine was started about a quarter of a mile west of the open cuts. A shaft was sunk, and two drifts were driven along gold-bearing quartz veins. Some thousands of dollars of gold was recovered from ore which consisted of hickory bark at the surface and pyritic ore at depths below about 15 feet.

Harrison Mine and Property (Emmons, 1890; Weed, 1905; Ulke, 1939): Gold was discovered in 1888 by a Mr. Kirk. Of the eight quartz veins prospected on the property, only vein number 7, the Harrison vein, was mined. A shaft 160 feet deep was sunk, and drifts were run at the 90- and 160-foot levels. Bullion produced through 1889 totalled about $12,000. Work was abandoned prior to 1904.

Irma Mine (Emmons, 1890): A shaft about 50 feet deep was sunk, but this mine was apparently of short life.

Bogley and Eagle Mines (Emmons, 1890; Ulke, 1939): There is no information about either of these mines. They apparently were of minor importance.

Huddleston Gold Mine (Emmons, 1890; Ulke, 1939): A 50-foot shaft was sunk along a quartz vein, and a tunnel was driven at 36 feet. The mine closed in 1889.

Anderson Property and Potomac Mine (Weed, 1905): Seven veins were prospected on the Anderson propery, of which only the Potomac vein was mined.

Sawyer Mine: This mine is mentioned by Nitze and Wilkens (1895).

Maryland Mine (Weed, 1905; Ulke, 1939): The Maryland Mine had the longest life and was the most valuable of all the mines in the Great Falls area. The original shaft was opened by Clear in 1867. It was abandoned, and a new one was driven in 1903. The new

TABLE 52

GOLD PRODUCTION

CALENDAR YEAR	VALUE	CALENDAR YEAR	VALUE
1867–1879	$2,500*	1901	$139
1880	250	1902	2,719
1881	500	1903	632
1882	1,000	1904	2,379
1883	500	1905	14,821
1884	500	1906–1913	—
1885	2,000	1914	225
1886	1,000	1915–1916	—
1887	500	1917	66
1888	3,500	1918–1932	—
1889	3,500	1933	473
1890	16,884†	1934–1935	—
1891	11,212†	1936	23,380
1892	1,000	1937	36,400
1893	114	1938	29,925
1894	978	1939	2,485
1895	499	1940–1949	—
1896	1,038	1950	700
1897	364	1951	35
1898	890	1952–1961	—
1899	1,173		
1900	392	Total value.....	$164,673

* Reference: Director of the Mint, Ann. Rept., U. S. Bureau of the Mint, 1920–1962. The years 1867 to 1917 are estimated production figures, see Ann. Rept., 1920, p. 60.

† Values from: Rept. of Director of Mint upon production of Precious minerals in U. S., 1890, 1891

shaft penetrated to 150 feet with drifts at the 100- and 150-foot levels. By 1939 the shaft was deepened to 200 feet, and the third drift was driven. From about 1936 to 1938 approximately $50,000 in gold was produced.

Bethesda Mine: (Nitze and Wilkens, 1895): This mine had a shaft 102 feet deep, at which depth the ore shoot pinched out. A total of $20,000 to 30,000 in gold was reported produced prior to 1895.

GOLD PRODUCTION

All told the gold mines of Montgomery County produced approximately $164,000 worth of gold. Yearly production figures are shown in Table 52.

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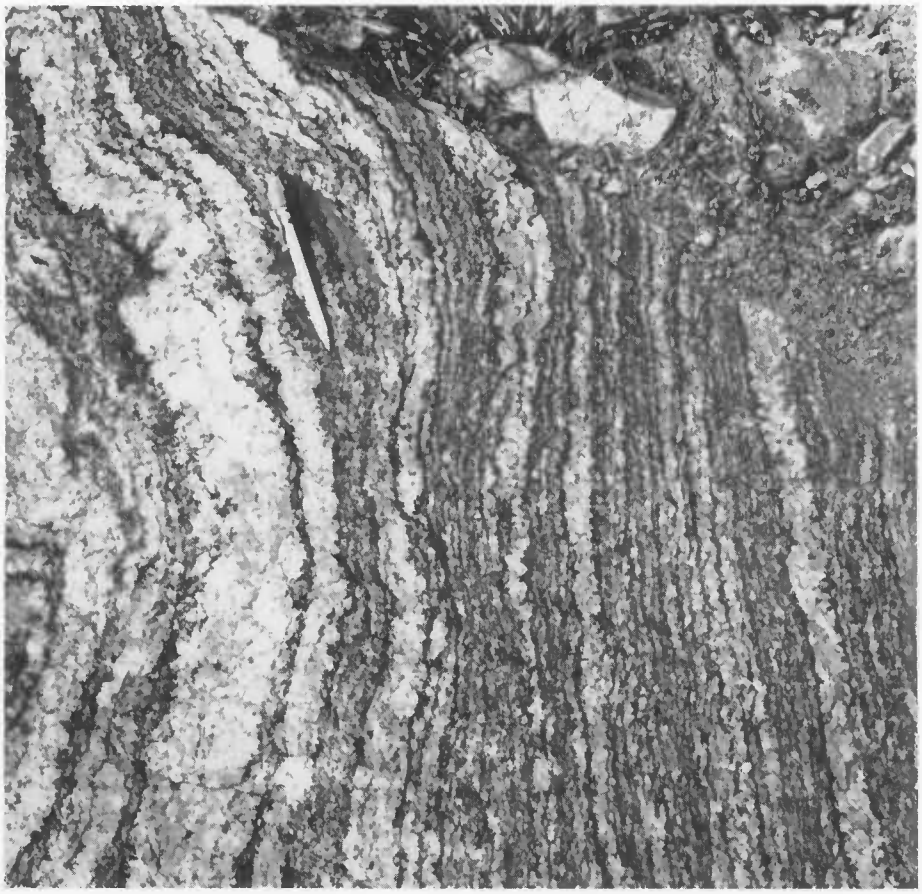
DESCRIPTION OF PLATES

PLATE 1, *Figure 1* (upper). Veined gneiss (right) grading into migmatite. Coarse porphyroblastic granite forms segregations and concordant replacement bodies in biotite gneiss. Biotite concentrated along their borders. Baltimore Gneiss at the northern edge of Phoenix dome, half a mile west of Verona.

Figure 2 (lower). Migmatitic Baltimore Gneiss. Porphyroblastic alaskite replacing hornblende-biotite gneiss. Phoenix dome, half a mile west of Verona.



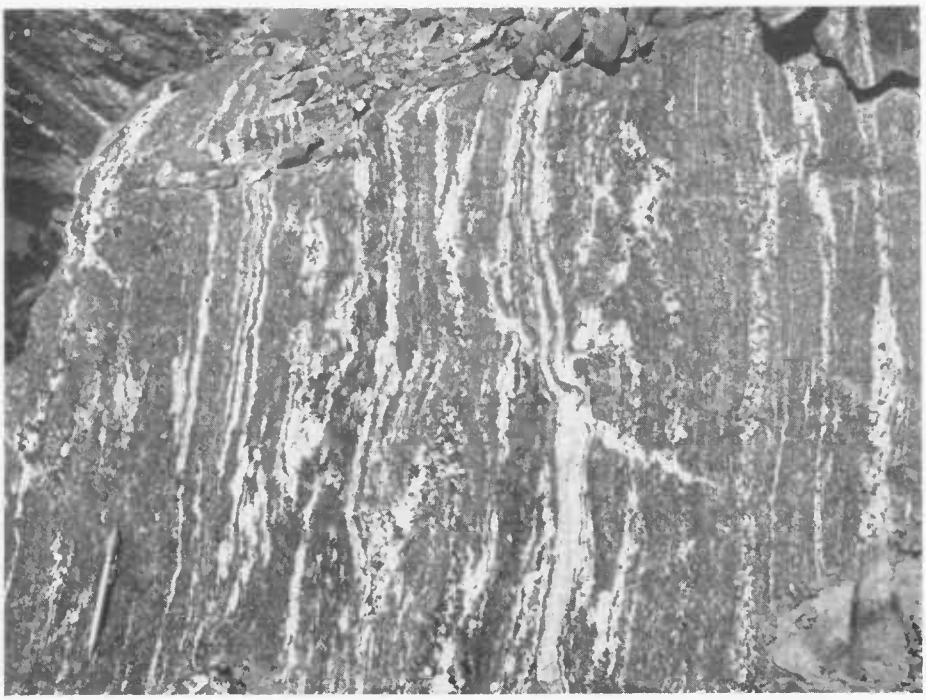
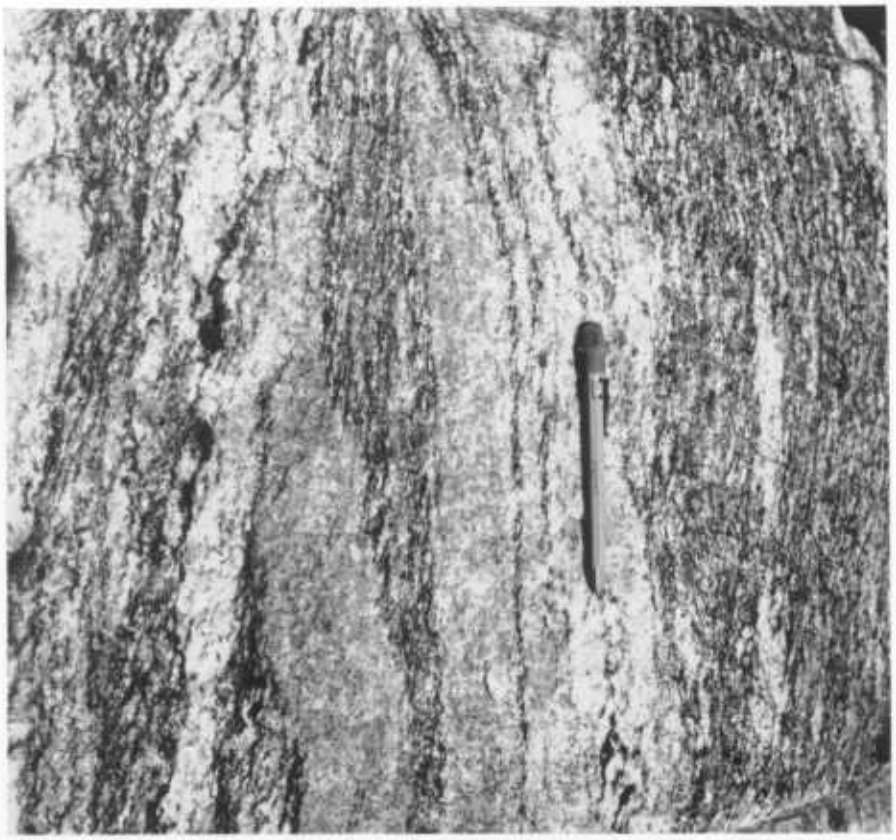


PLATE 2, *Figure 1* (upper). Migmatite, with alaskitic granite replacing biotite gneiss. Faint relict gneissic structure can be traced through the slightly discordant granite bodies. Baltimore Gneiss at northern edge of Phoenix dome, half a mile west of Verona.

Figure 2 (lower). Migmatitic Baltimore Gneiss; granite replacing biotite gneiss. Undisturbed relict gneissic structure runs through the granite layer. Expelled biotite is densely clustered within the granite body, and along its margin. Phoenix dome, half a mile west of Verona.



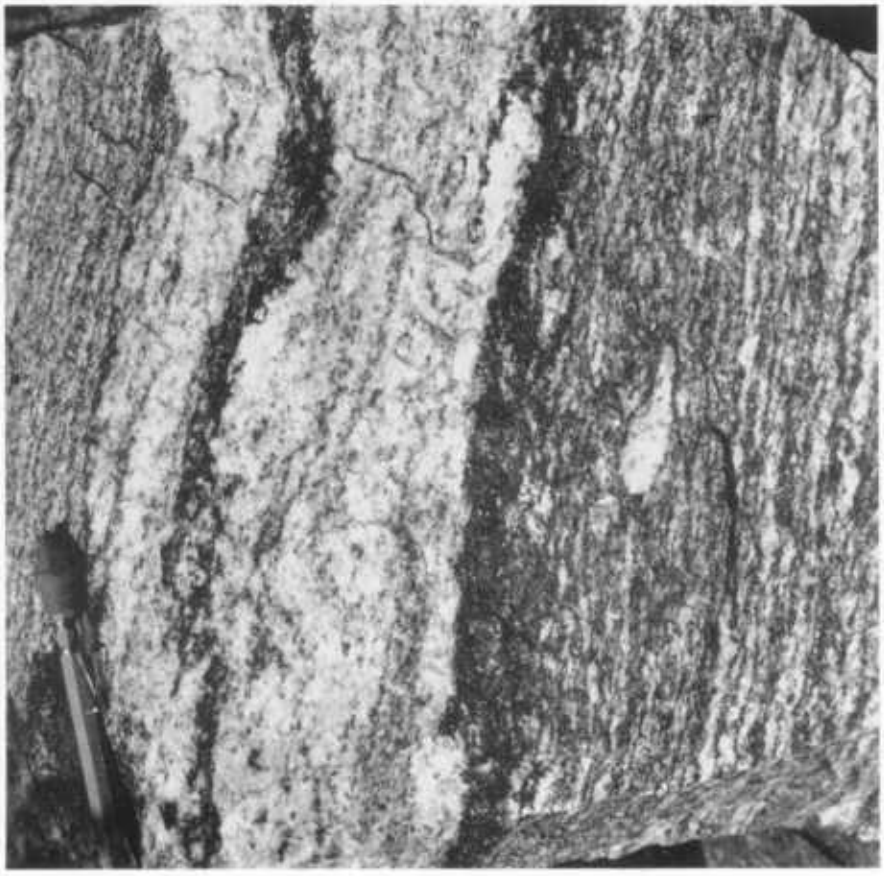


PLATE 3, *Figure 1* (upper). Contorted layering in migmatitic Baltimore Gneiss. Coarse porphyroblastic microcline-rich granite concordantly replaces thin-banded biotite gneiss. The microcline porphyroblasts were partly crushed during folding, and recrystallized. Specimen from Phoenix dome, west of Verona.

Figure 2 (lower). Ptygmatically folded replacement or exudation vein in hornblende-biotite gneiss. Note the concentration of mafic minerals along the sides of the vein. Baltimore Gneiss at Phoenix dome, half a mile west of Verona.

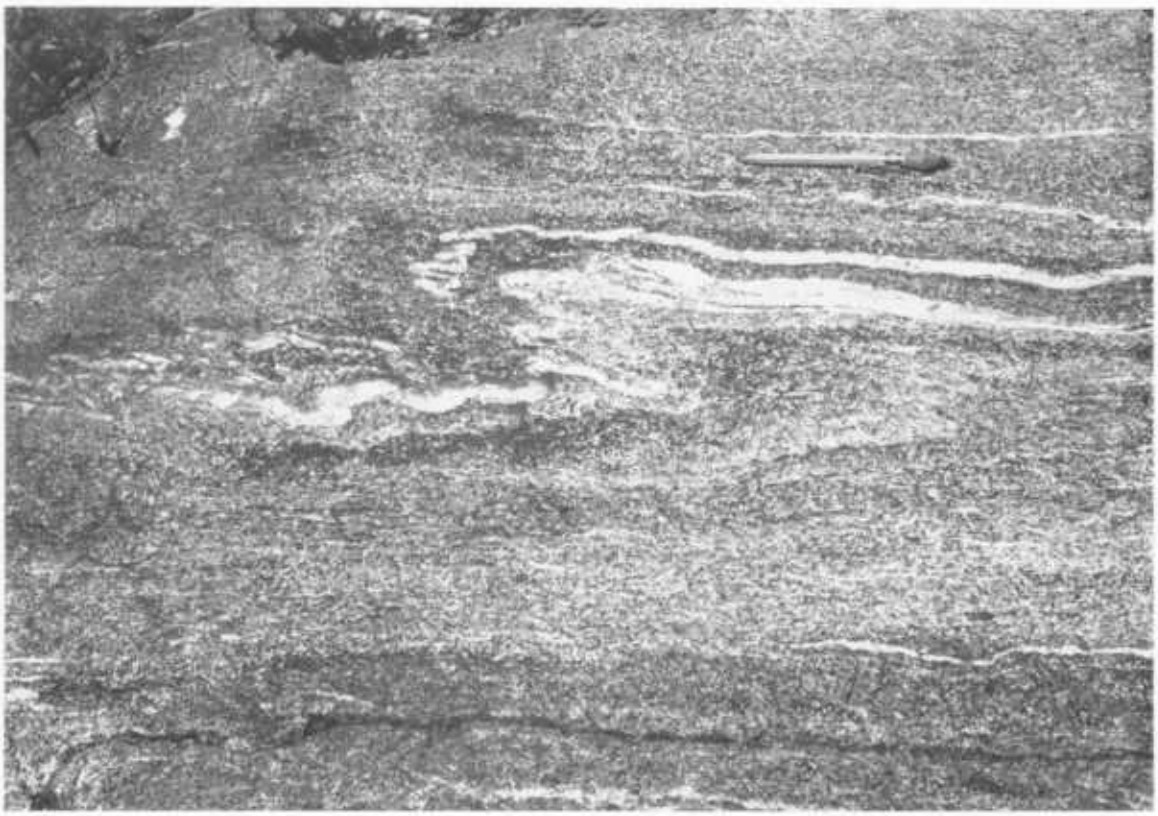
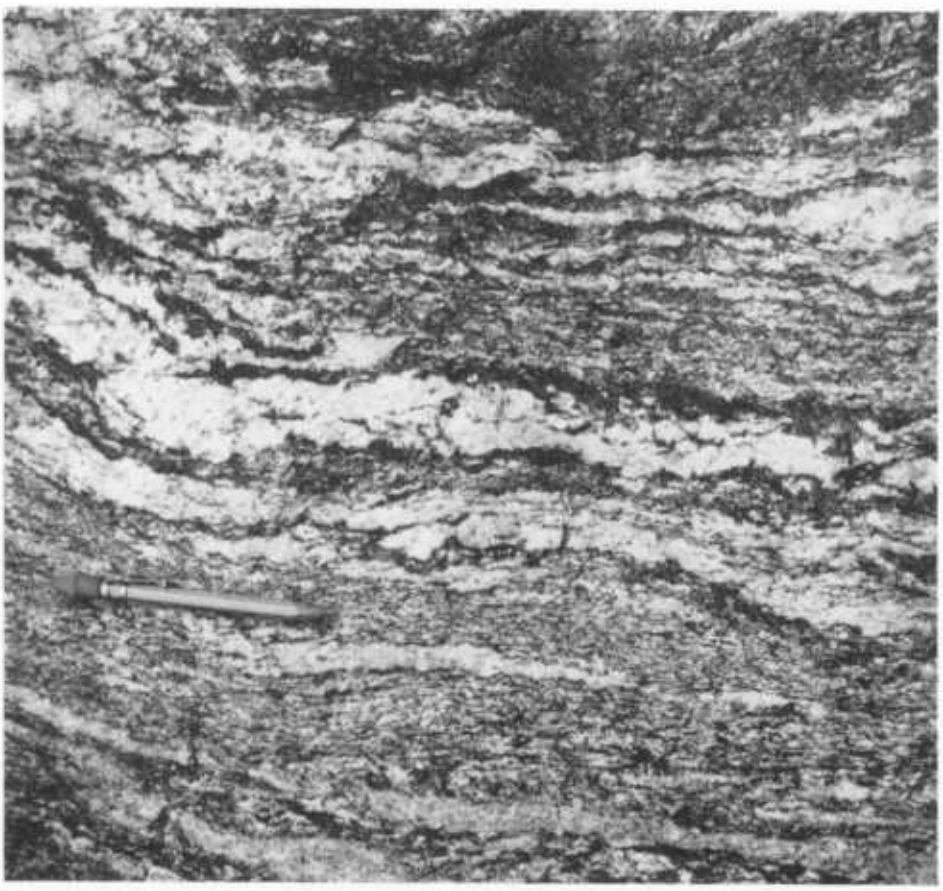


PLATE 4, *Figure 1* (upper). Migmatite in Baltimore Gneiss, showing the diffuse, gradational contacts of the light-colored granite replacement bodies, and the clots of expelled biotite concentrated at their margins. Phoenix dome, half a mile west of Verona.

Figure 2 (lower). Specimen of granitic gneiss, from Baltimore Gneiss at western end of Towson dome. Charles St. Extended south of Joppa Road, Towson.



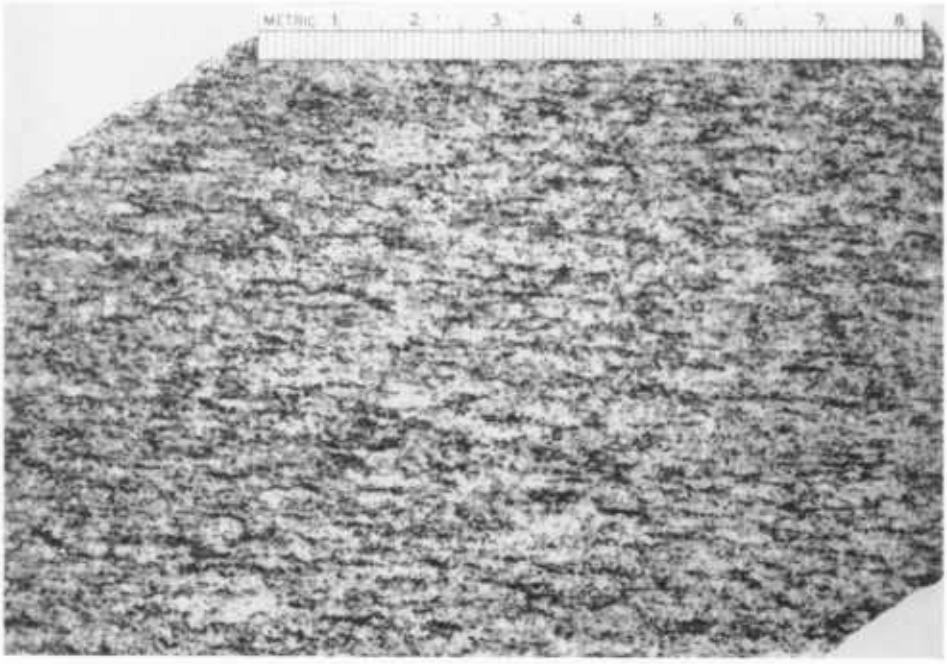

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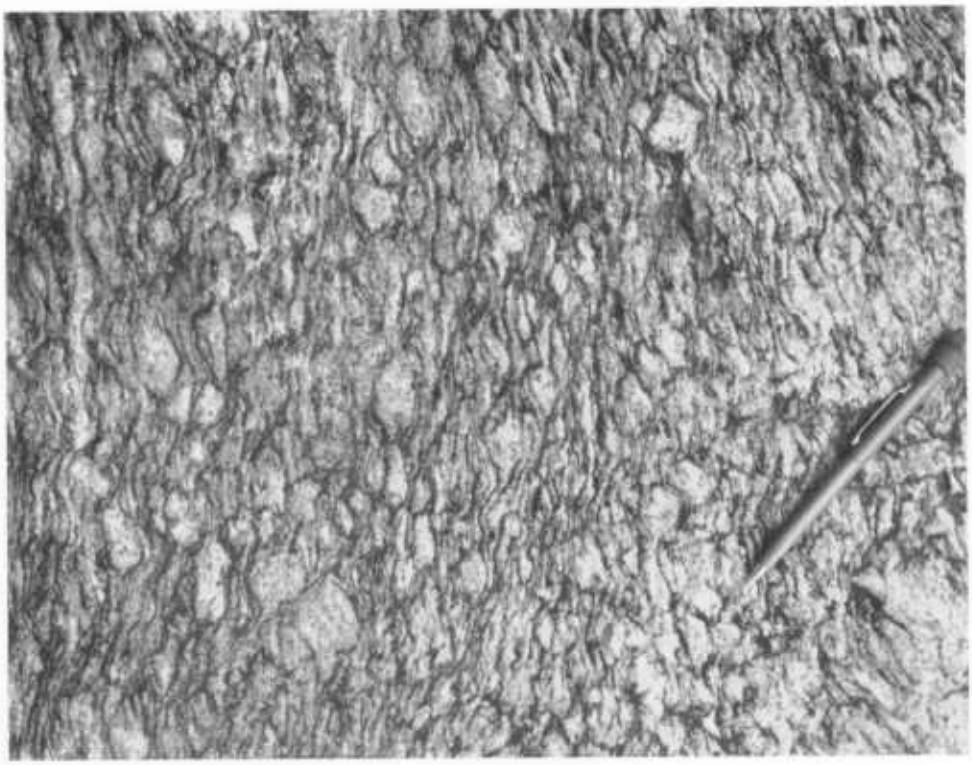
PLATE 5, *Figure 1* (upper). Augen gneiss. Near the pencil the densely clustered microcline augen merge, forming a patch of coarse-grained granite. Northern edge of Towson dome, at Cub Hill Road.

Figure 2 (lower). Augen gneiss. Large deformed microclines in fine-grained biotite-quartz-oligoclase gneiss. From migmatitic Baltimore Gneiss at the southern edge of Texas dome, Pot Spring Road.



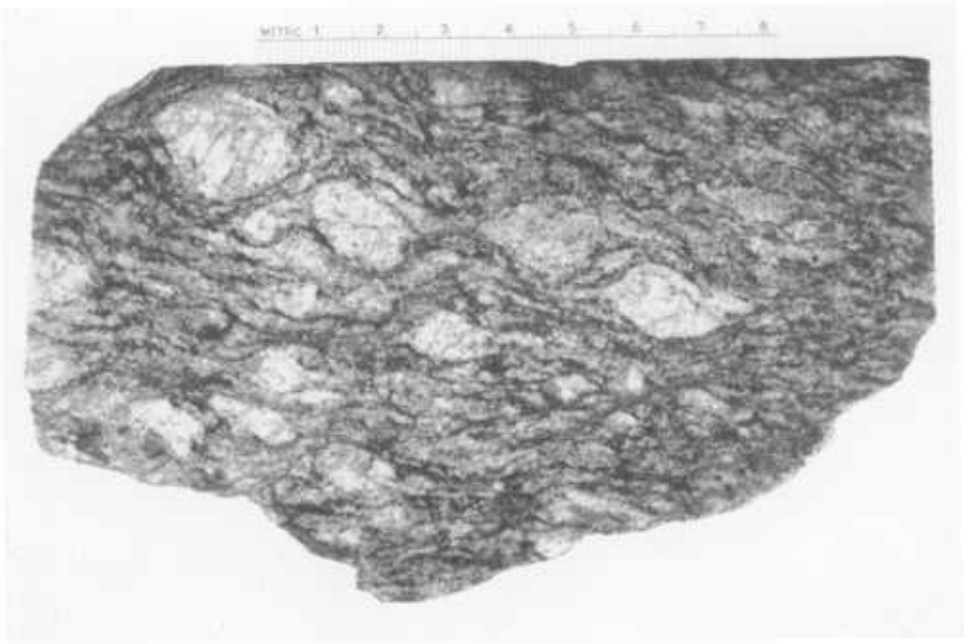

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PLATE 6, *Figure 1* (upper). Granitized layer in Wissahickon mica schist. Note gradational contacts and undisturbed patches of mica preserved within the feldspathic layer. Specimen from southeastern limb of Towson dome, Harford Road at Gunpowder Falls.

Figure 2 (lower). Discordant intrusive contact between fine-grained Gunpowder Granite (dark) and Baltimore Gneiss. Flow structure in the granite parallels the contact, and truncates the foliation in the gneiss. Specimen from Gunpowder Falls, Towson dome.

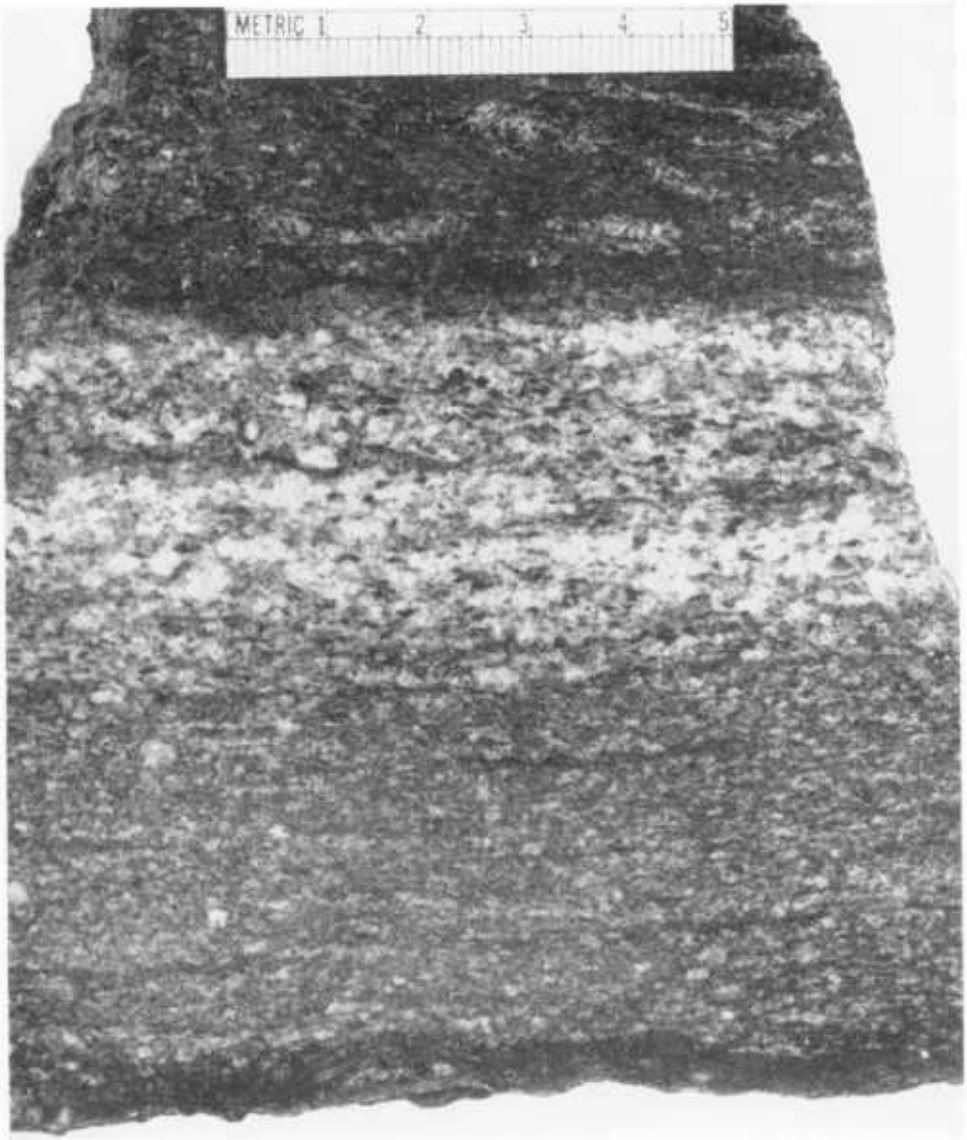
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Plate 7 follows Plate 45

PLATE 8, *Figure 1* (upper). Thin-bedded quartzite from the Setters Formation, middle member. The light-colored layers are rich in microcline, and the darkest layers rich in muscovite. The medium-gray layers are chiefly quartz. Northern rim of Phoenix dome, half a mile west of Verona, Baltimore County. The scale is in centimeters.

Figure 2 (lower). Relict bedding in the Wissahickon Formation, eastern sequence. Light-colored beds of fine-grained psammitic granulite are interstratified with darker, coarse-grained staurolite-garnet schist. Well-developed bedding schistosity is cut by a weak axial-plane slip cleavage at the sharp folds in the foreground. One and one-half miles southwest of Fulton, Howard County.

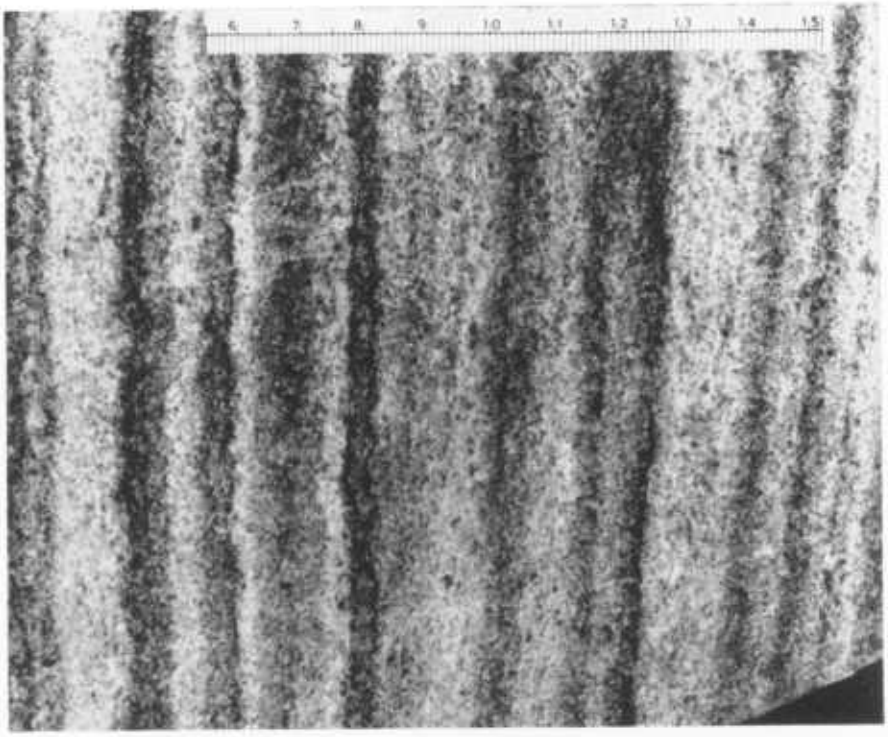
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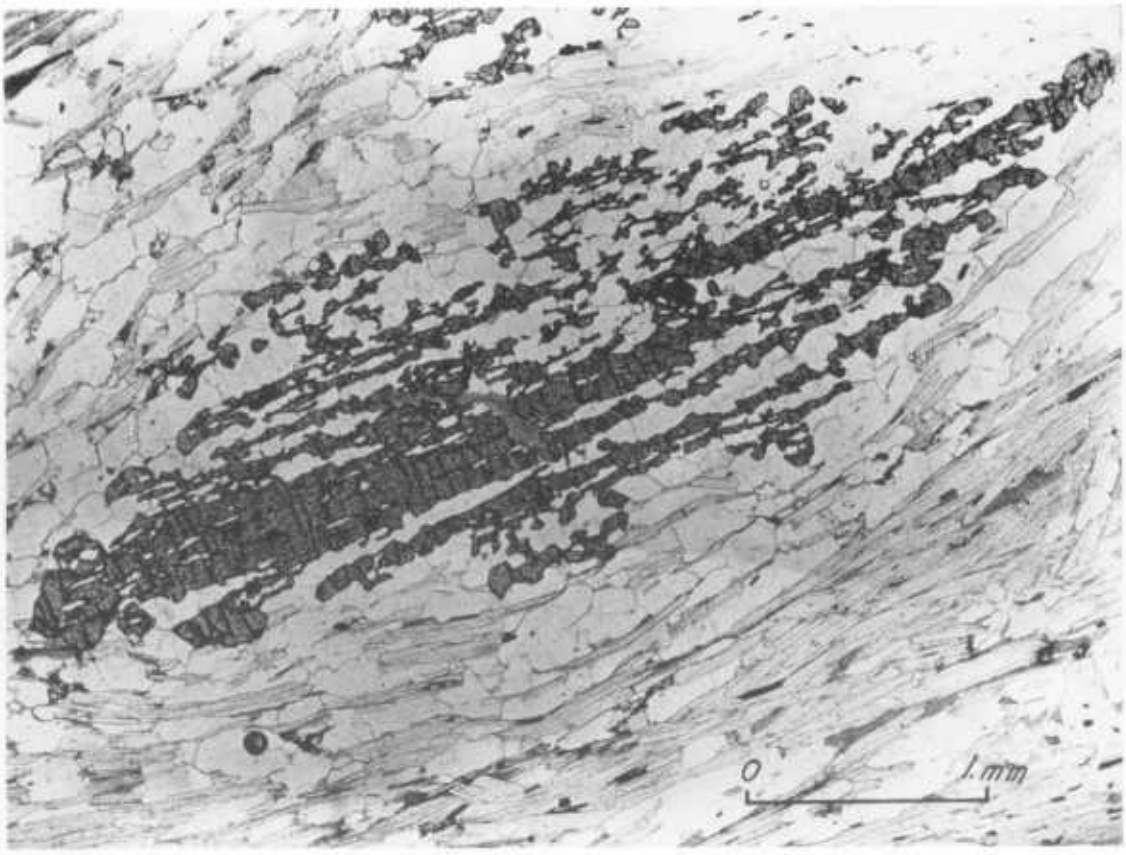
PLATE 9, *Figure 1* (upper). Skeletal garnet porphyroblast, in staurolite-garnet schist. Lines of quartz inside the porphyroblast preserve the trend of the bedding schistosity. Wissahickon Formation, eastern sequence; 1 mile southwest of Butler, Howard County. Plane light.

Figure 2 (lower). Sieved garnet porphyroblasts, in staurolite-garnet schist. The garnets have nearly straight internal trend lines, moved out of alignment with the external schistosity (vertical). The garnets were twisted after they had ceased to grow. Wissahickon Formation, eastern sequence; Tucker Lane, 200 yards south of Patuxent River, Montgomery County.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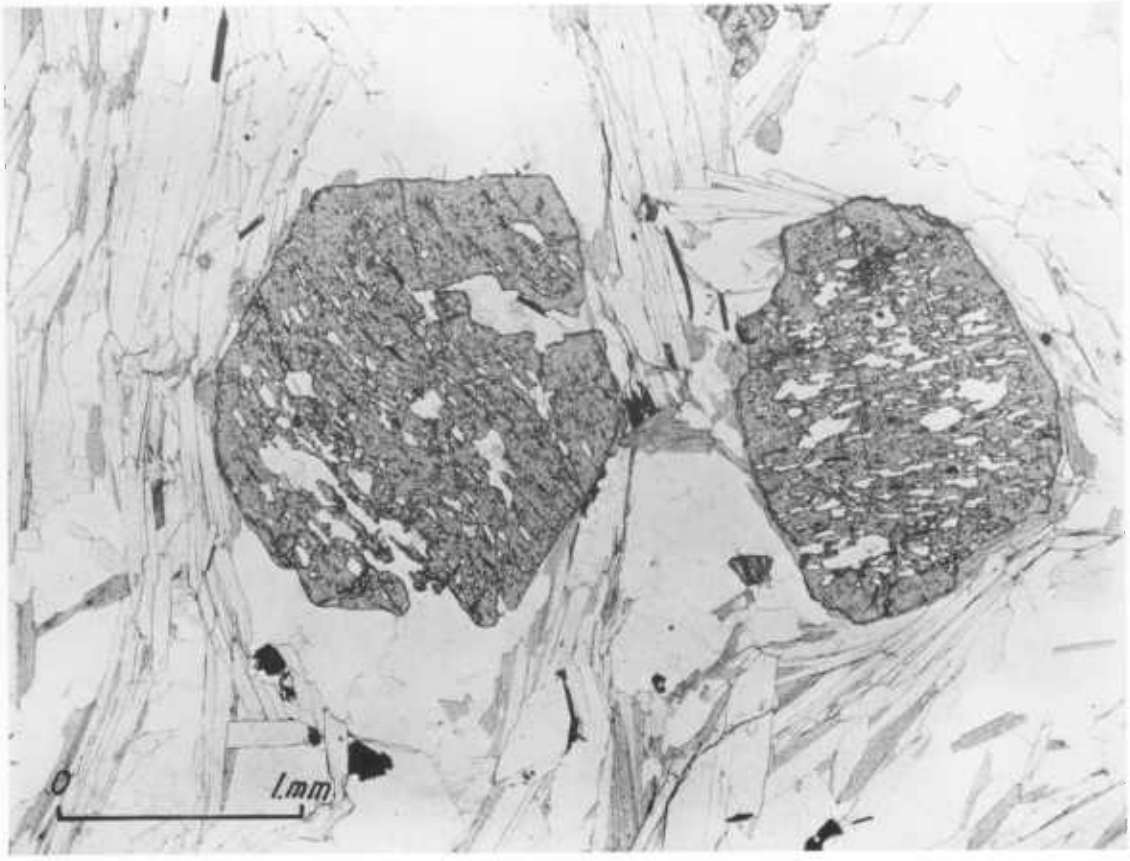
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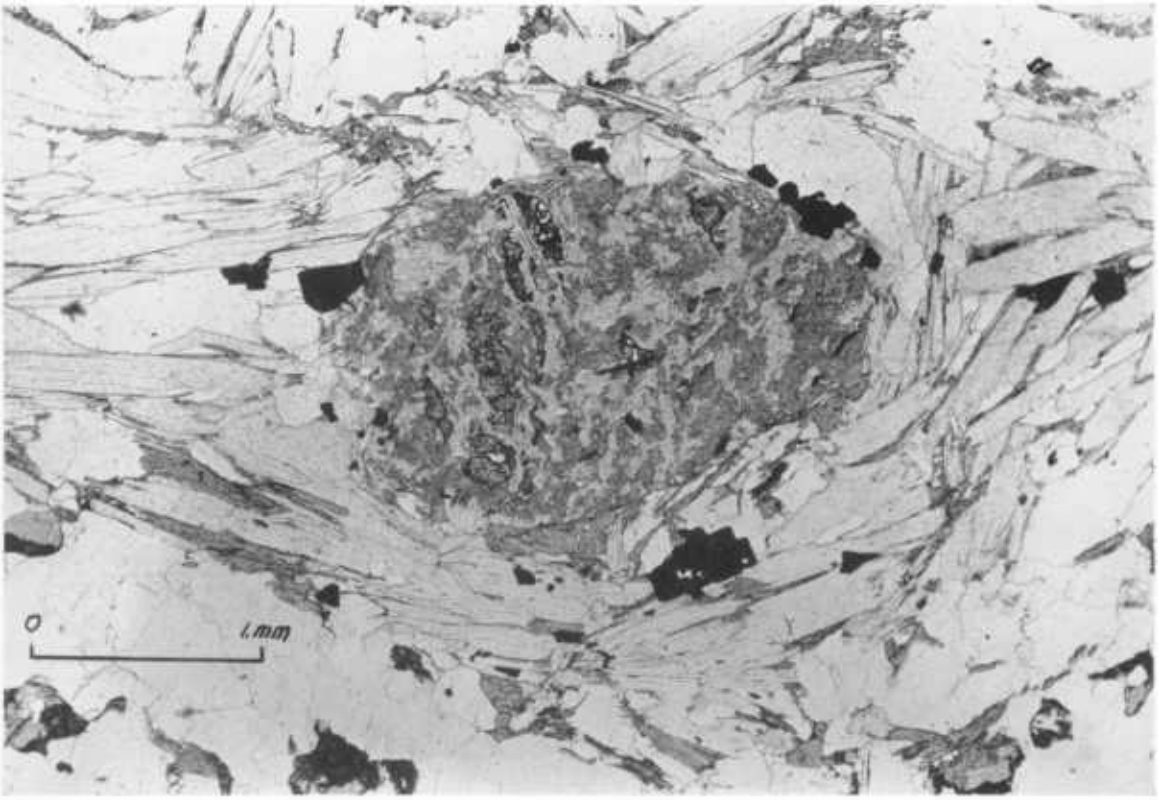
PLATE 10, *Figure 1* (upper). Relict garnet prophyroblast, in retrograded kyanite-staurolite-garnet schist. The garnet is almost completely altered to chlorite, sericite, and magnetite, although small remnants remain left of center. Biotite in the surrounding rock is strongly chloritized. Wissahickon Formation, eastern sequence; 1 mile southwest of Butler, Howard County. Plane light.

Figure 2 (lower). Relict staurolite porphyroblast, in retrograded Wissahickon schist. The crystal is partly altered to sericitic shimmer aggregate (light-colored), which encloses a remnant of fresh staurolite (nearly black) in the core. Same specimen as Figure 1. Nicols crossed.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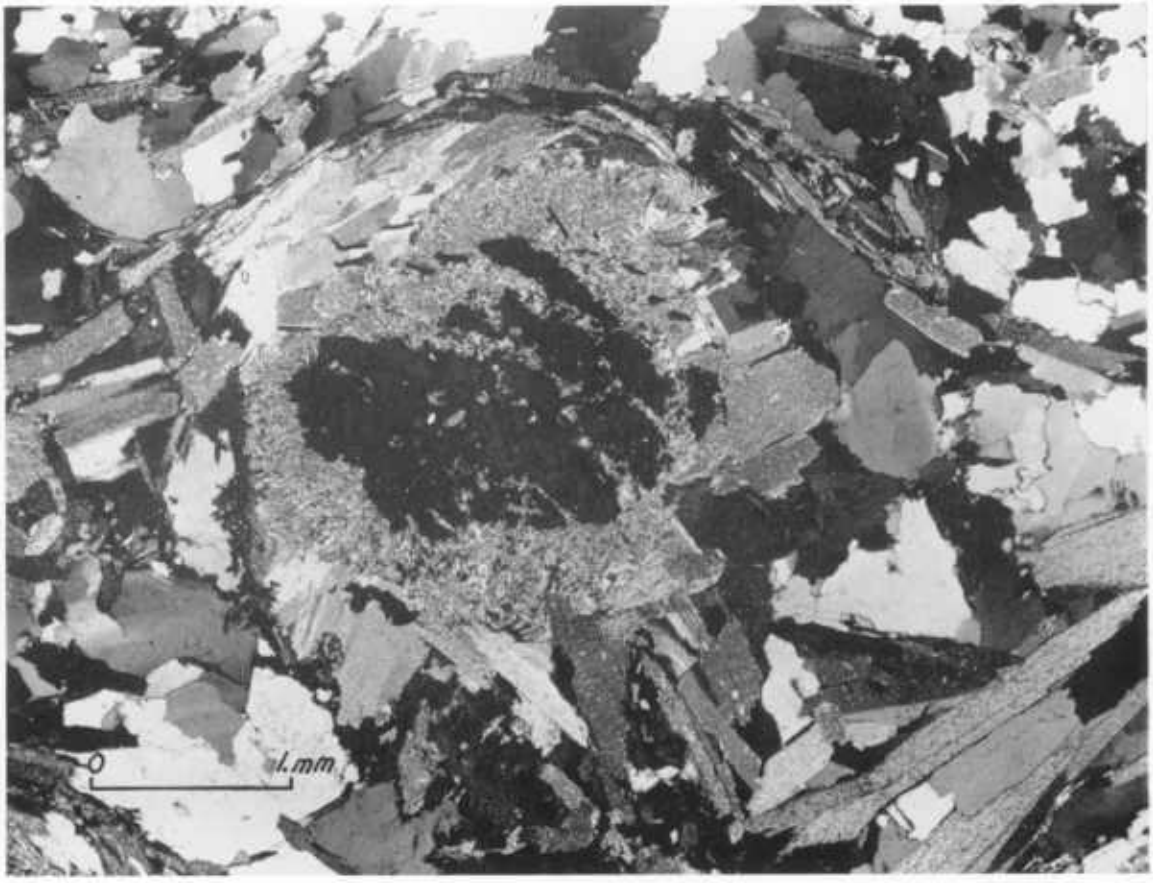
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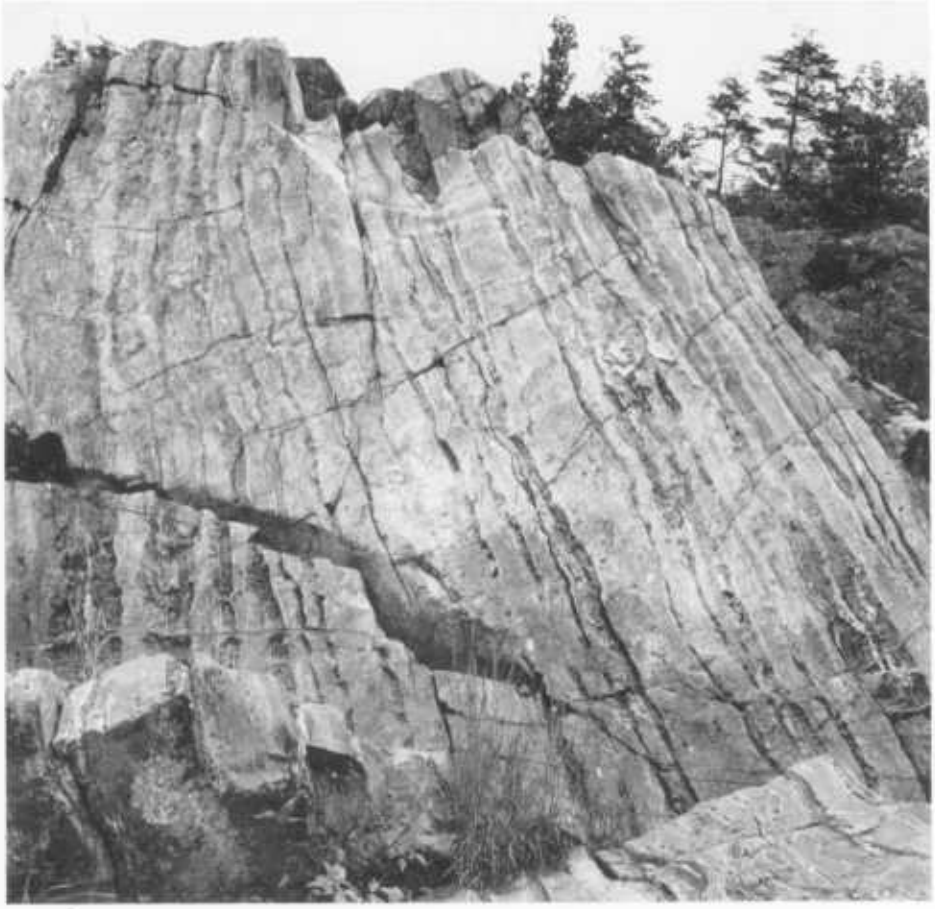
PLATE 11, *Figure 1* (upper). Bedding in the Wissahickon Formation. Thin alternating beds of metagraywacke and sillimanite-mica schist (dark layers). Some of the metagraywacke beds are graded; heir tops are to the right. Bear Island, Potomac River.

Figure 2 (lower). Graded metagraywacke bed (spanned by the pencil). Grades from light-colored metagraywacke upward into dark pelitic schist. The base of the bed rests sharply on the underlying pelite. Wissahickon Formation, at Bear Island.



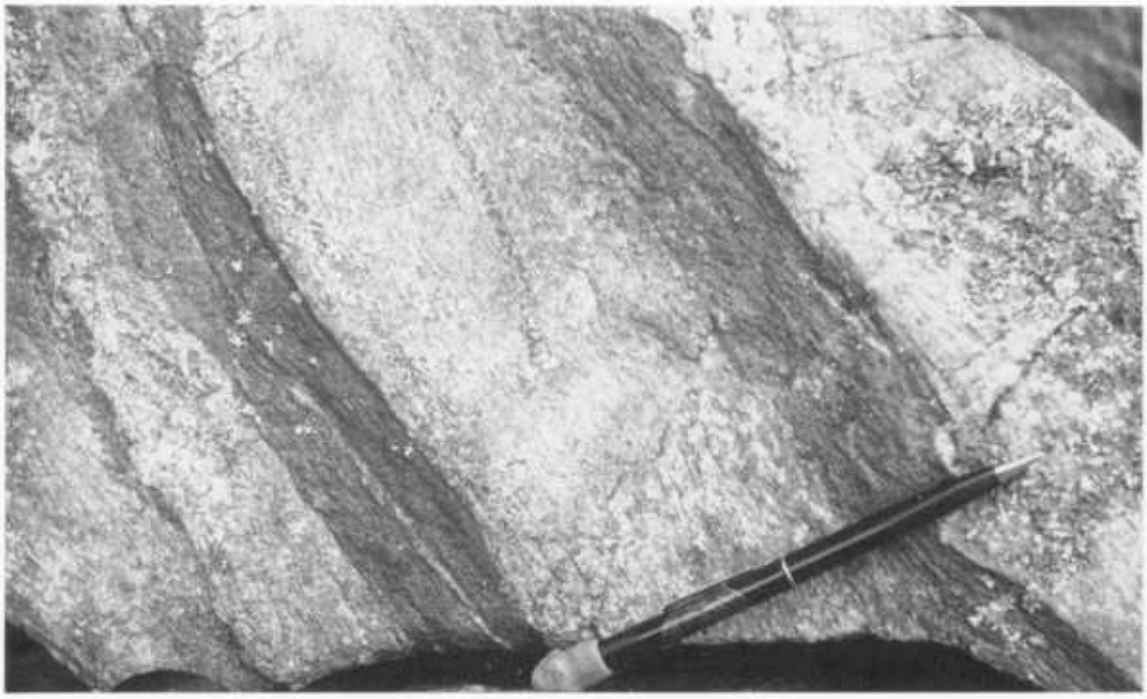


PLATE 12, *Figure 1* (upper). Graded metagraywacke bed, between thinner beds of pelitic schist The bed grades upward from medium-grained metagraywacke to metasiltstone; relict clastic grains are well preserved. Note the sharp basal contact for this and the overlying psammitic beds. Wissahickon Formation at Rocky Island, below Great Falls on the Potomac River.

Figure 2 (lower). The coarse-grained base of a 30-inch metagraywacke beds rests sharply on the silty top of an earlier graded bed. Relict sand grains up to 3 mm in diameter are visible in the upper bed, which grades up to metasiltstone (not visible). The upper part of the bed beneath displays fine bedding laminations and has ripple cross laminations (not visible in photograph) at the very top. Wissahickon Formation (staurolite zone) at Rocky Island, below Great Falls on the Potomac River.

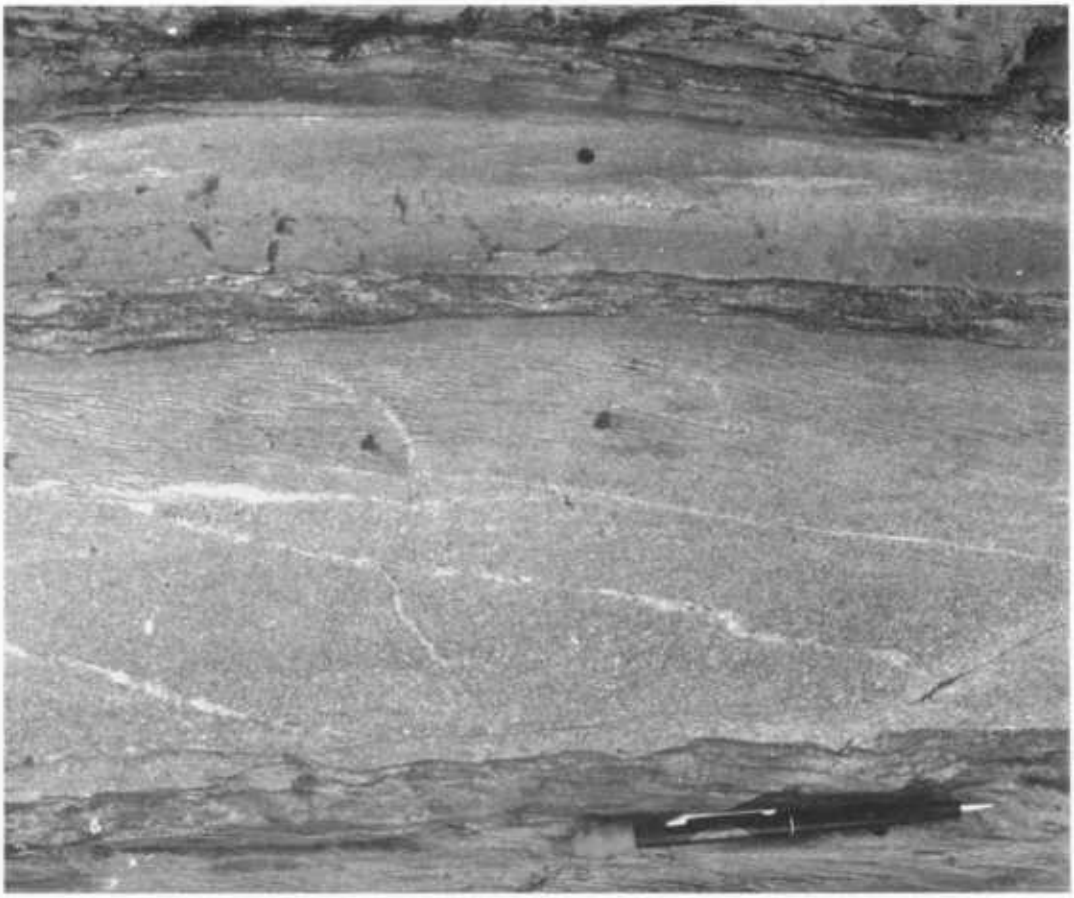


PLATE 13, *Figure 1* (upper). Graded metagraywacke bed between two pelitic (sillimanite schist) layers. The lower part of the graded bed is massive, the upper part has fine parallel laminations, and faint ripple cross laminations occur at the very top. Note the sharp lower contact and indistinct upper contact. The original material graded from very fine sand at the base to silt at the top. This is one of the beds shown in Figure 1 of Plate 11. Wissahickon Formation, at Bear Island.

Figure 2 (lower). Two graded metasandstone beds in sillimanite-andalusite schist. Each bed becomes increasingly micaceous toward the top and grades into the overlying pelite. The upper bed has faint cross-lamination near the top and bulges at the base that may be sole markings.

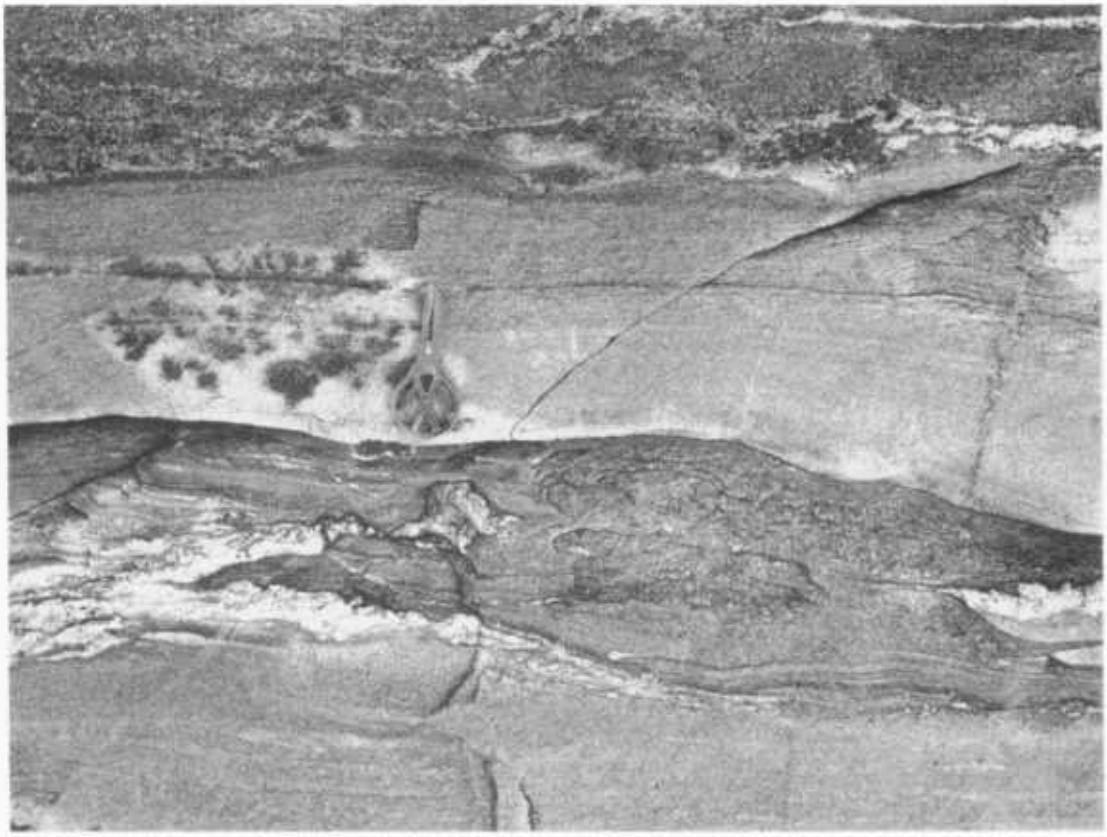
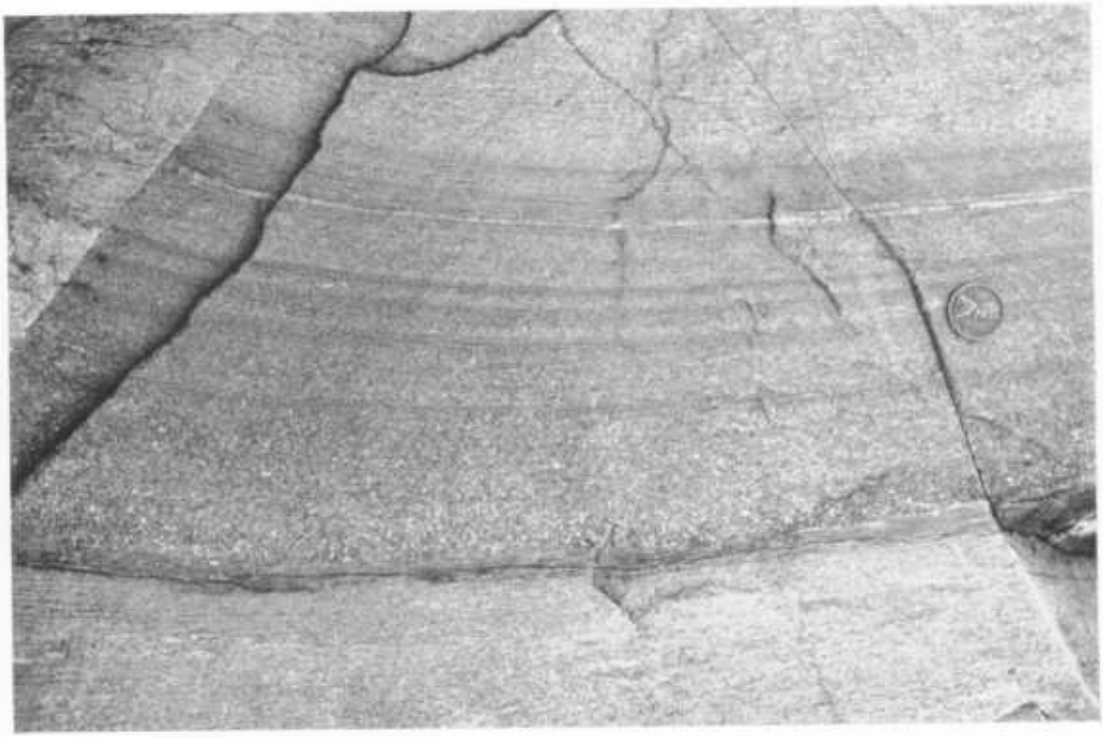


PLATE 14, *Figure 1* (upper). Contact between two laminite beds (metagraywacke). Very thin, parallel sedimentary laminations occur throughout each bed. The upper bed is graded at the base, and there are very thin, recurrent graded horizons just above (opposite the penny). Wissahickon schist (staurolite zone) below Great Falls, Potomac River.

Figure 2 (lower). Two laminite beds (metagraywacke) in pelitic schist, showing relict convolute bedding. The thin laminations are strongly contorted in the middle and upper part of the top bed, but the contortions die out at the contact. The edges of the bed are undeformed. Wissahickon schist (sillimanite zone) at Bear Island, Potomac River.



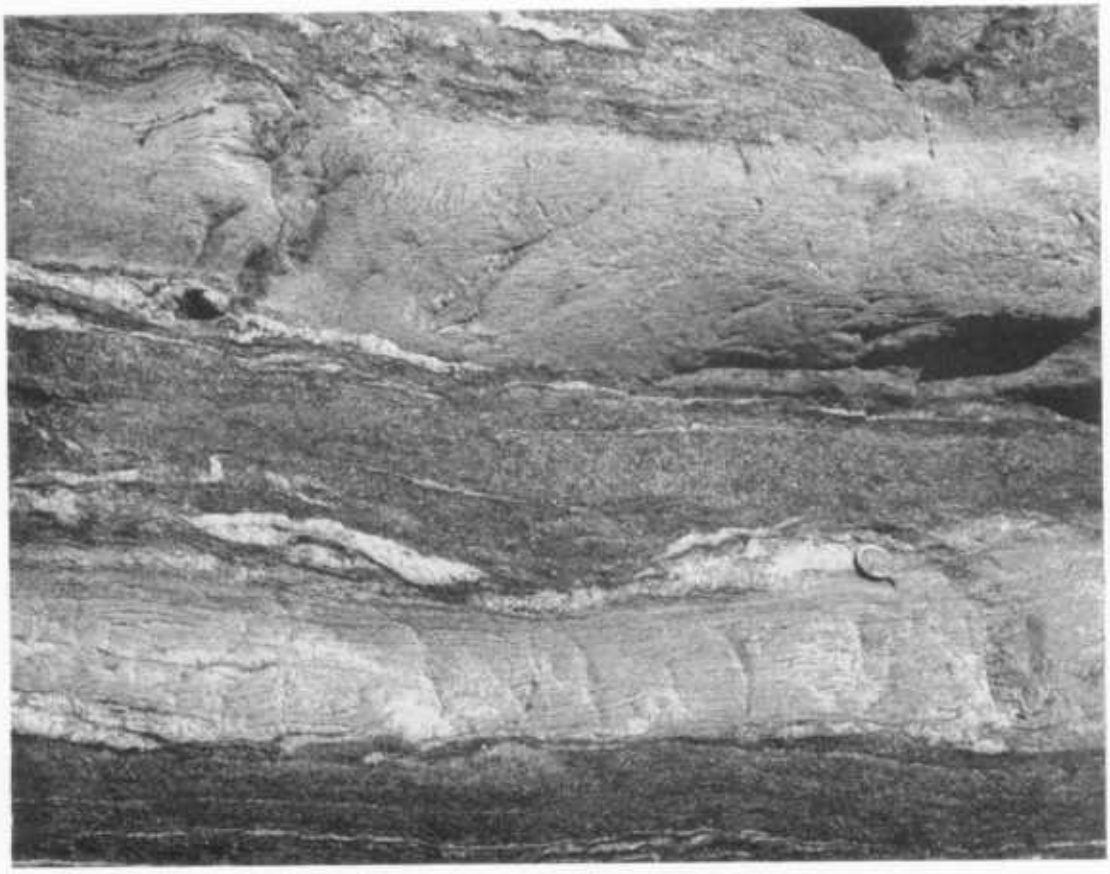
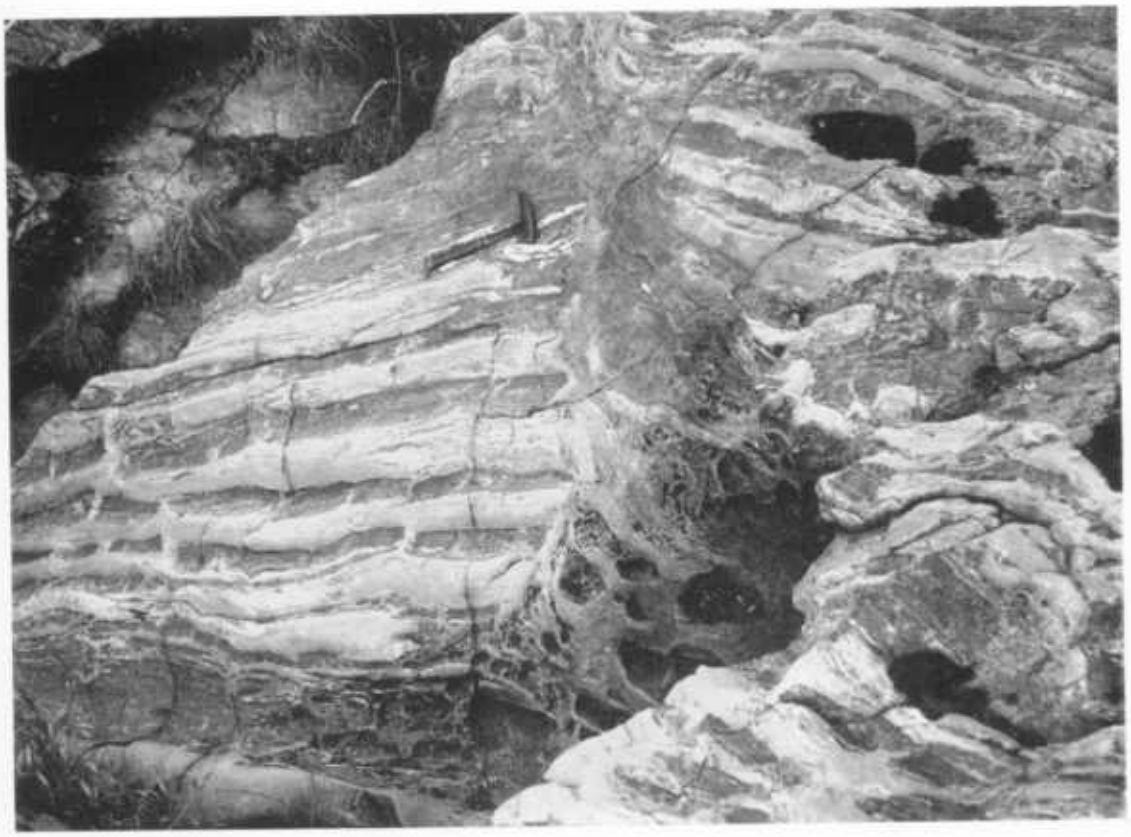


PLATE 15, *Figure 1* (upper). Disrupted beds of pelitic schist in metagraywacke. The dark pelitic beds have been pulled apart and engulfed by the light-colored psammitic material. One segment of a thin pelitic bed is doubled back on itself (left of the pencil point). Deformation resulted from liquefaction and flowage of the sand, soon after the beds were deposited. Wissahickon Formation (sillimanite zone); Bear Island, Potomac River.

Figure 2 (lower). Relict slump structure in Wissahickon schist. The dark pelitic beds (sillimanite schist) are crumpled and disrupted, and the psammitic material (metagraywacke) appears to have flowed around and between them. This strongly disturbed zone, 1–2 feet thick, can be traced for 40 feet and is enclosed on both sides by straight, undeformed beds (the lower undeformed bed is at the bottom of the photograph, and part of the upper bed is barely visible at the upper left). Bear Island, Potomac River.

PLATE 16, *Figure 1* (upper). Relict clastic dike in the Wissahickon schist. The strata were faulted (note offset beds) and injected by quicksand. In the lower part of the dike the sand carried fragments of disrupted pelitic beds, but higher up the sand and mud mixed together. The beds are right-side up, and it appears that the dike was injected upward. Note the tiny "dikes" that link some of the adjacent metasandstone beds (light-colored) at the left side of the photograph. Bear Island, Potomac River.

Figure 2 (lower). Graded metagraywacke beds, showing relict soft sand intrusion structures. At 1 a bulbous mass of sand was squeezed into the overlying shale and across a higher sand layer. At 2 a sand dike has intruded the shale bed beneath. The syncline is a later tectonic fold. Wissahickon Formation (sillimanite zone) at Bear Island, Potomac River.



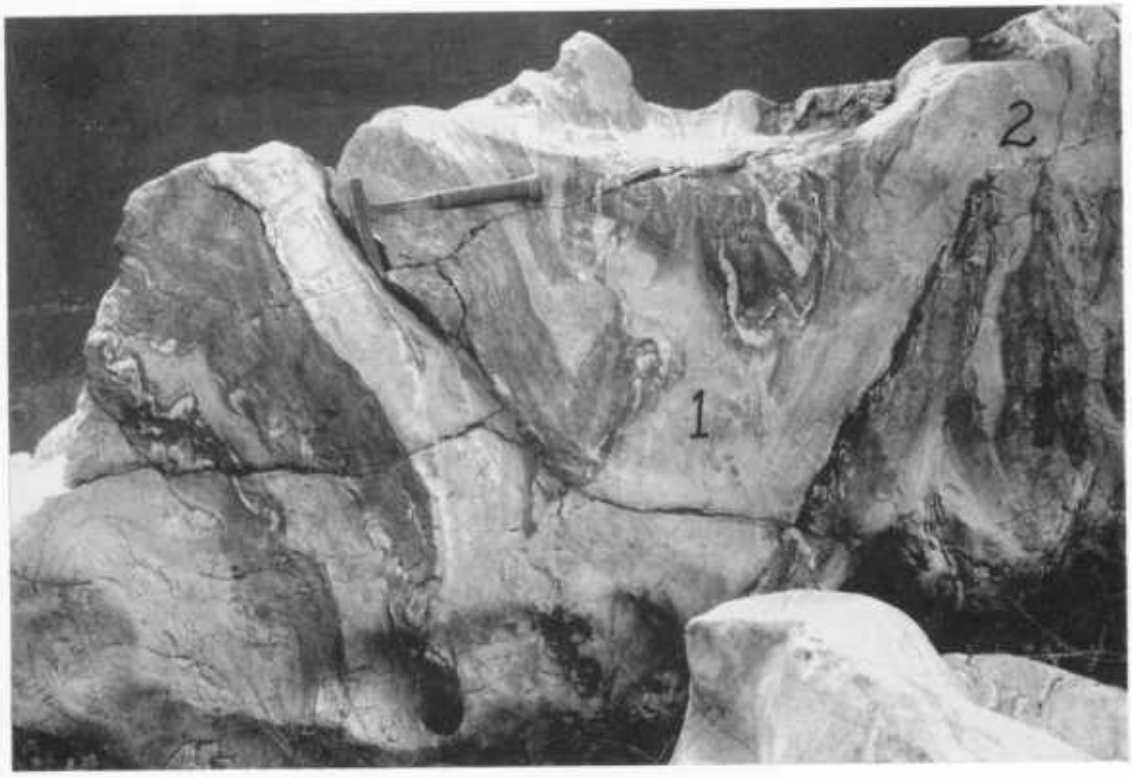

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PLATE 17, *Figure 1* (upper). Laminated metasiltstone, distorted by slumping while the sediments were still soft. Six feet farther to the right the bed becomes completely dismembered (Fig. 2, below). Wissahickon schist, at Bear Island.

Figure 2 (lower). Laminated metasiltstone fragment in pelitic schist. The bed shown in Figure 1 (above) has been pulled apart into a string of disconnected fragments like this one. Disruption occurred while the sediments were still soft, as shown by the irregular "slurried" shape of the fragment, by its mixing with the enclosing pelite, and by the intense contortion of its internal laminations.

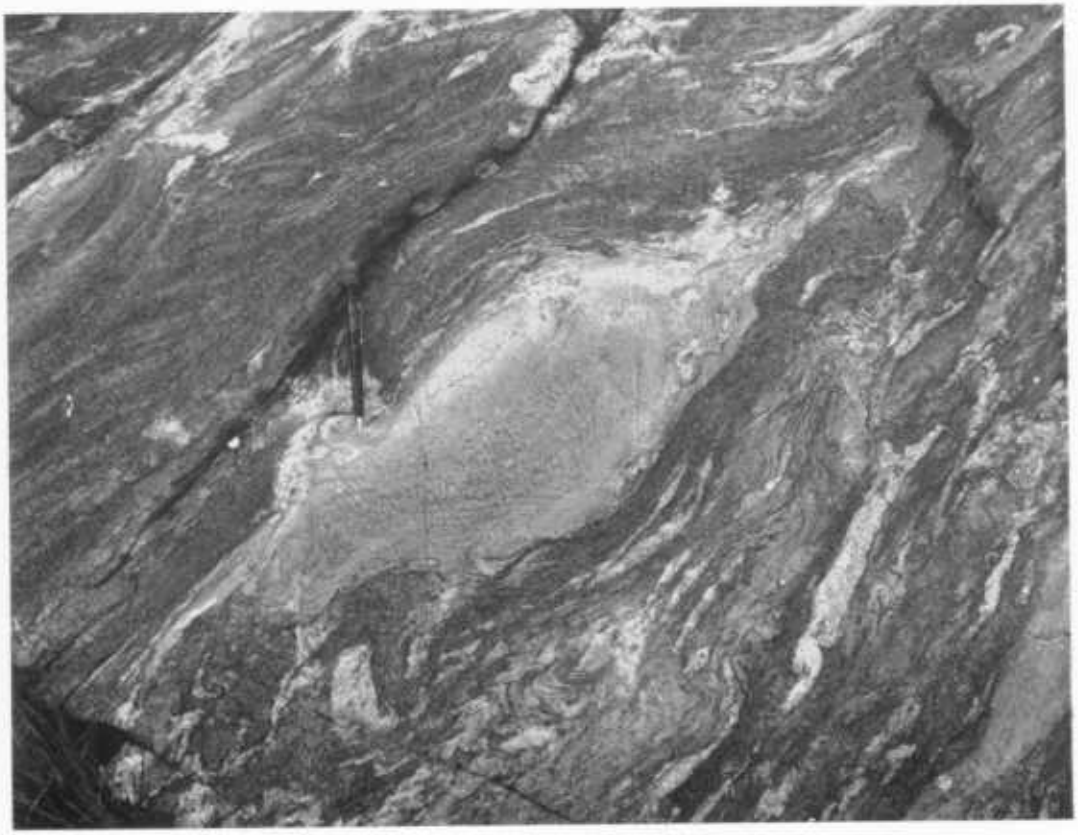
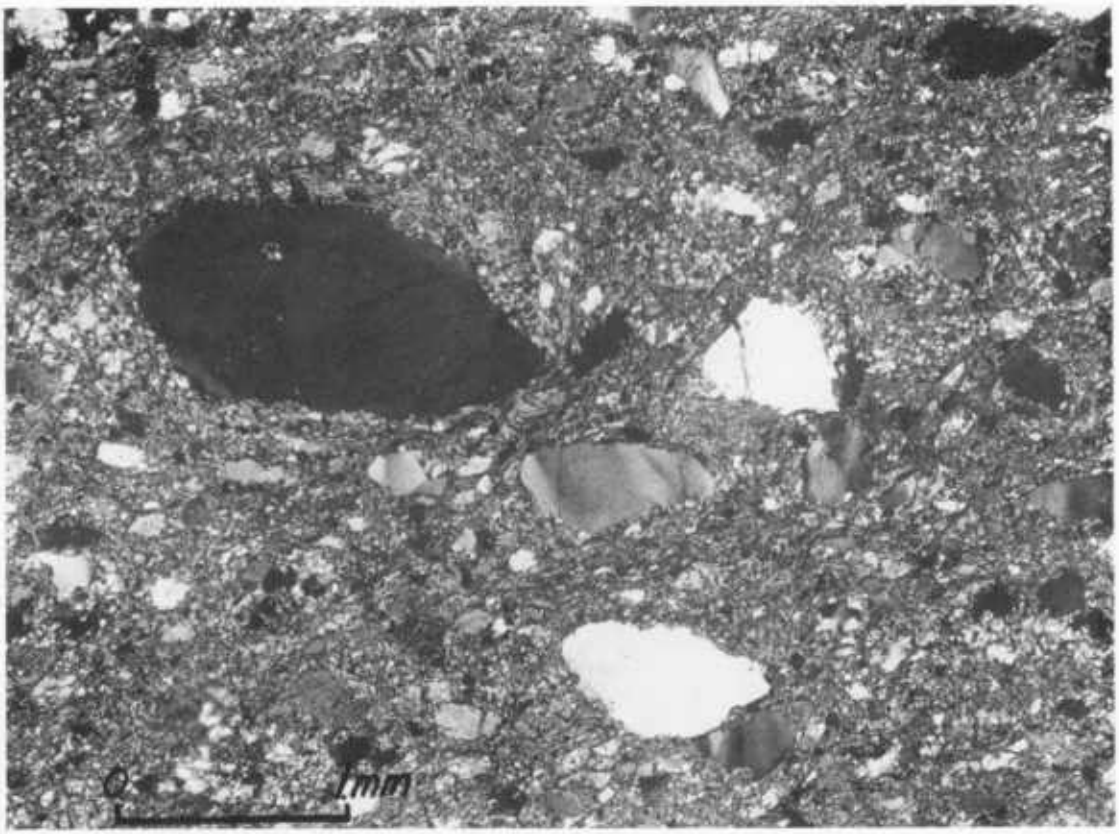


PLATE 18, *Figure 1* (upper). Laminite bed, disrupted by slumping. Wissahickon schist, at Bear Island.

Figure 2 (lower). Remnants of a disrupted sand bed, engulfed in pelitic schist. Wissahickon schist, at Bear Island.

PLATE 19, *Figure 1* (upper). Photomicrograph of metagraywacke, showing poorly sorted subangular relict quartz grains in a fine micaceous-chloritic matrix. Wissahickon Formation, western sequence (chlorite zone). Crossed nicols.

Figure 2 (lower). Photomicrograph of metagraywacke, showing relict clastic plagioclase. Wissahickon Formation, western sequence (chlorite zone). Crossed nicols.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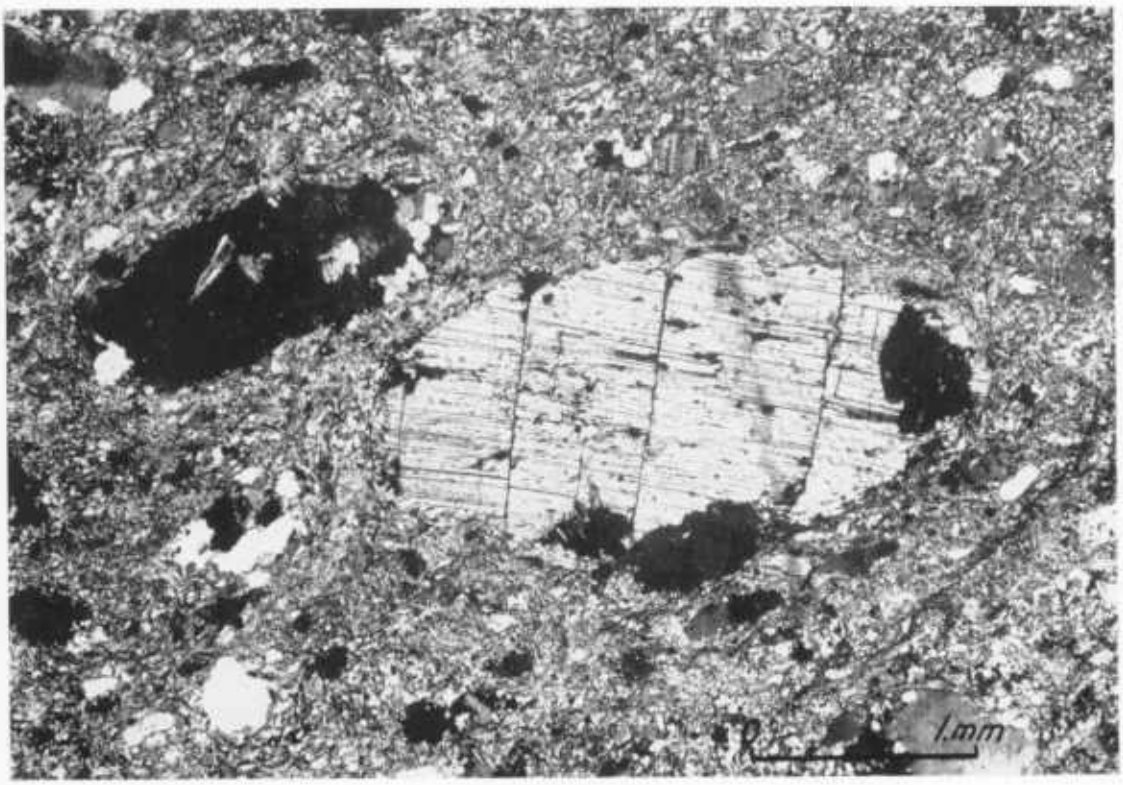
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PLATE 20, *Figure 1* (upper). Relict andalusite porphyroblast in pelitic schist. The andalusite is pseudomorphed by sericite. Wissahickon schist at Bear Island, Potomac River.

Figure 2 (lower). Relict sillimanite (?) porphyroblasts in pelitic schist. The original crystals are pseudomorphed by sericite. Wissahickon schist at Bear Island, Potomac River.

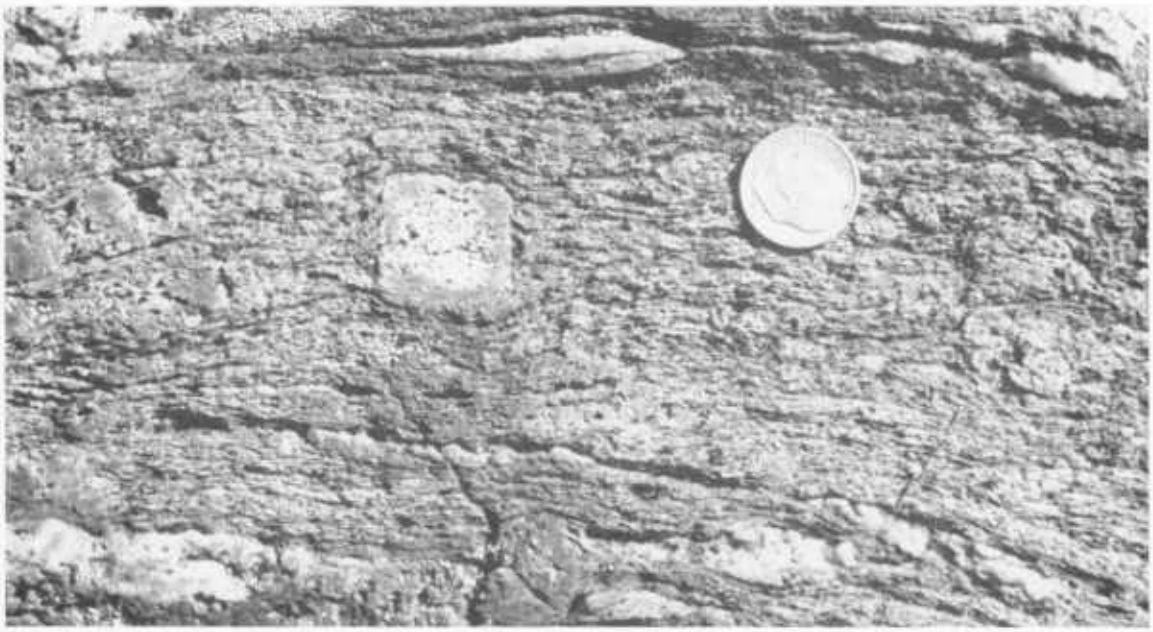
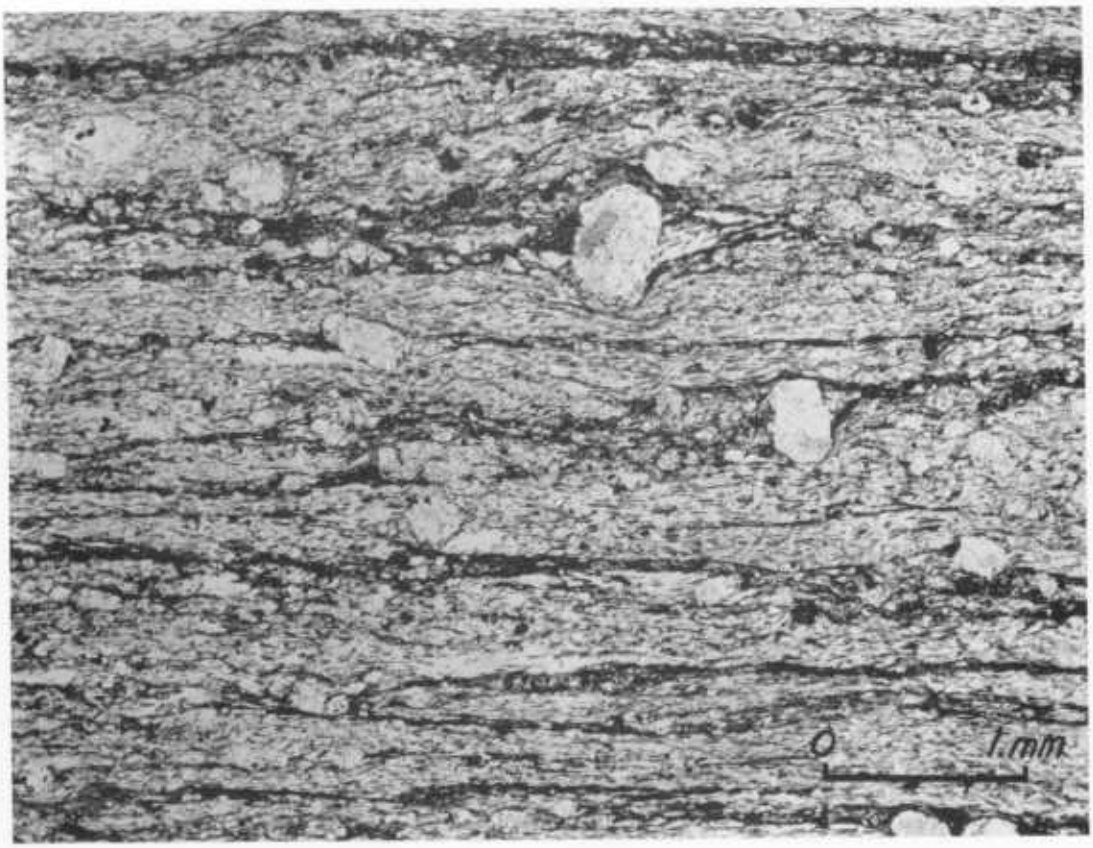


PLATE 21, *Figure 1* (upper). Photomicrograph of a laminated metasiltstone, containing sparse relict sand grains. Wissahickon Formation, Western sequence (chlorite zone); near Blackrock Mill, Montgomery County. Plane light.

Figure 2 (lower). Photomicrograph, showing a sieved plagioclase porphyroblast (black). On each side of the plagioclase are relict clastic quartz grains, recrystallized to sutured mosaics. Metamorphosed sandy mudstone, Sykesville Formation. Near Glenwood, Howard County, crossed nicols.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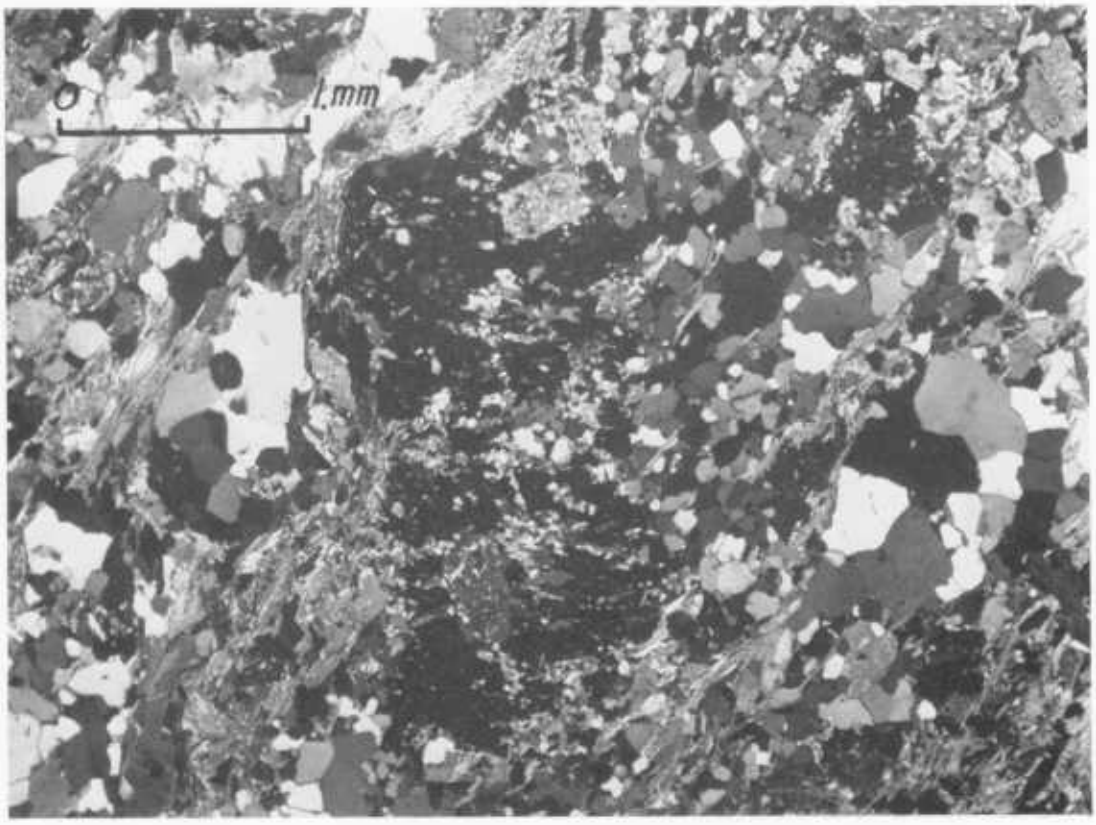
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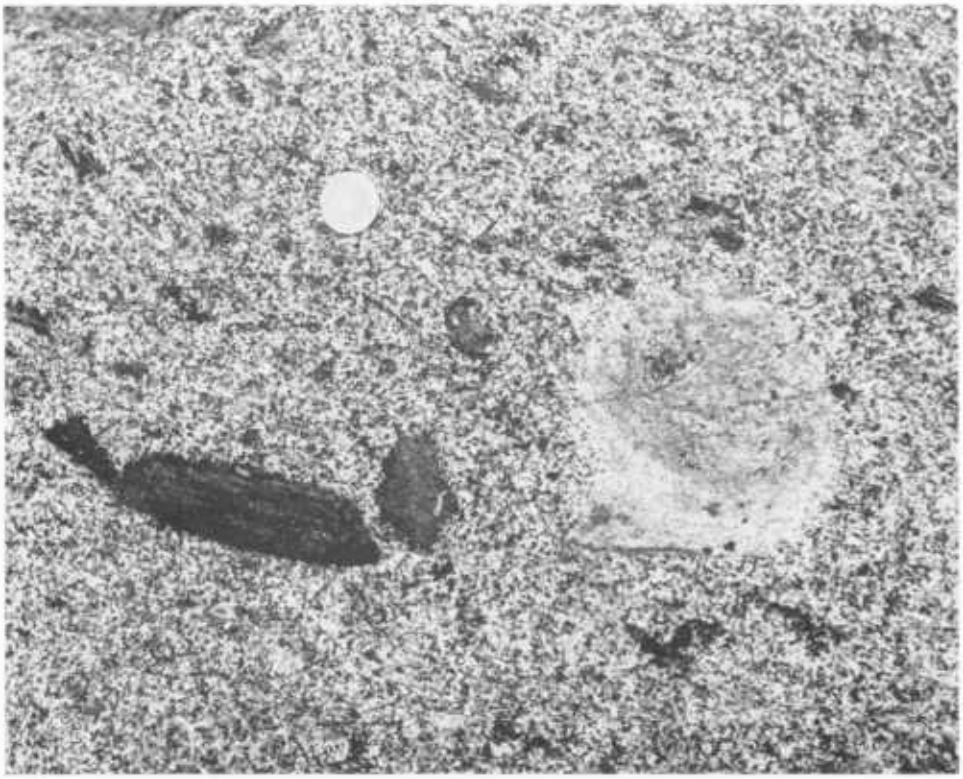
PLATE 22, *Figure 1* (upper). Granitic-appearing rock from the Sykesville Formation. Fragments of mica schist (dark) and metagraywacke (light), which resemble xenoliths, are enclosed in rock with a pseudo-granitic texture. Patapsco River, near Sykesville.

Figure 2 (lower). Calc-silicate inclusion in the Sykesville Formation. The inclusion consists of a contorted lens of quartz-epidote rock (1), enclosed by metasiltstone (3) and pelitic schist (4). A narrow quartzo-feldspathic reaction rim (2) is developed between 1 and 3. The Sykesville host rock (5) resembles granite. Photograph of a specimen collected at Sykesville by C. R. Keyes, and figured in U. S. Geol. Survey 15th Ann. Rept., Pl. XLVIII, p. 736, 1895. The scale is 1 cm.



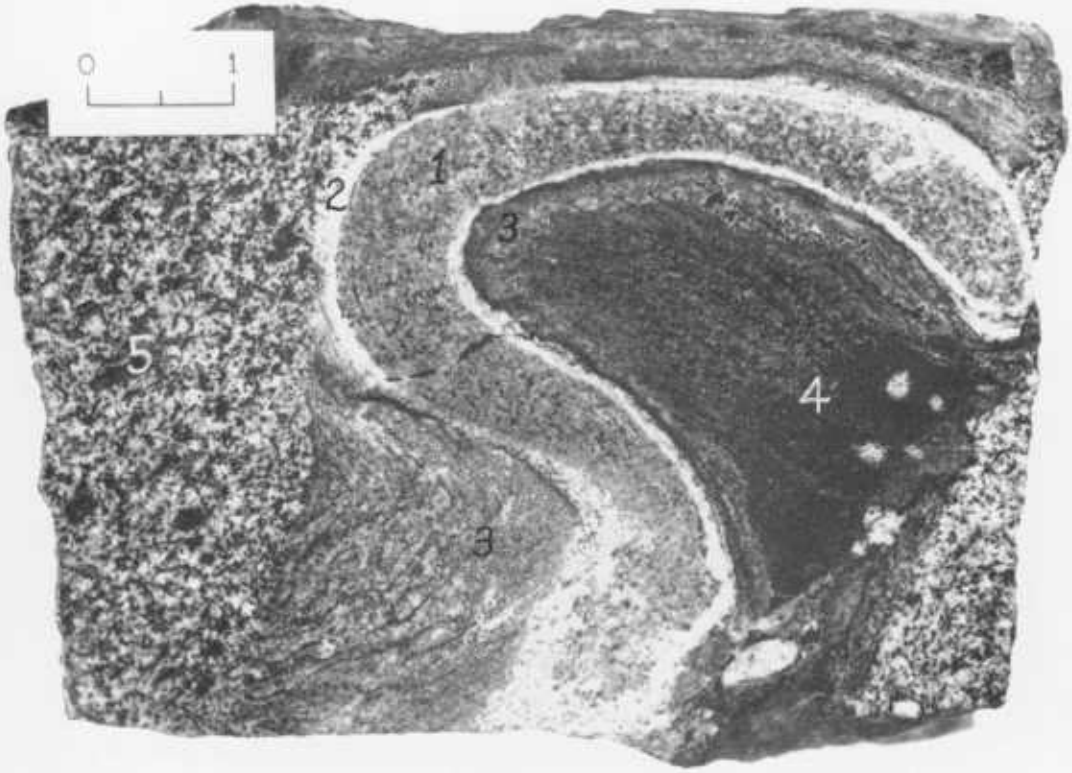

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PLATE 23, *Figure 1* (upper). Relict quartz pebble in metasedimentary rock of the Sykesville Formation. A penny gives the scale. Outcrop near Sykesville Road, 1 mile south of Sykesville, Howard County.

Figure 2 (lower). Small fragments of mica schist in rock of the Sykesville Formation. The fragments, which resemble xenoliths or autoliths in granite, were shale chips in mud-rich graywacke. Same outcrop as Figure 1, above.

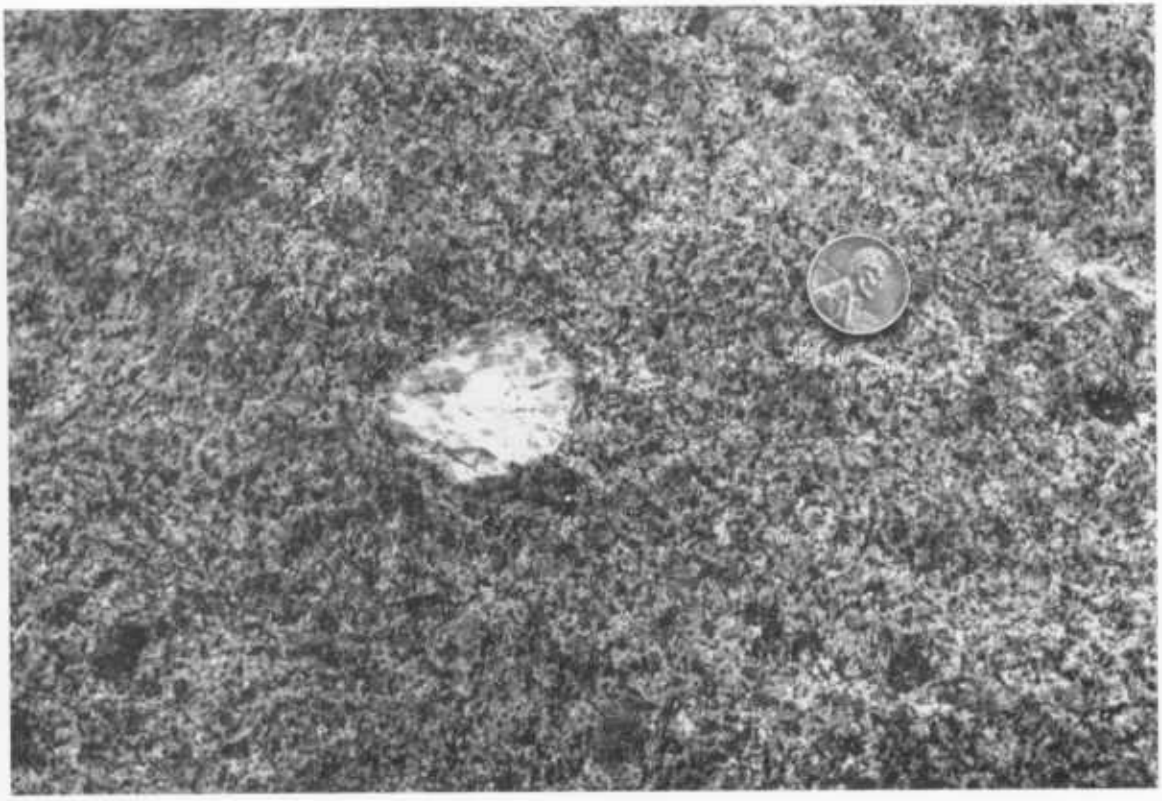


PLATE 24, *Figure 1* (upper). Metamorphosed sandy mudstone of the Sykesville Formation, containing blocks of quartzite and abundant small fragments of dark pelitic schist. Quartz pods and veinlets are white. Laminations in the lower quartzite block lie at right angles to the foliation of the host rock. Abandoned quarry, 1 mile west of Brookville, Montgomery County.

Figure 2 (lower). Slab-shaped, randomly oriented psammitic fragments in the Sykesville Formation. Patapsco River, near Sykesville.

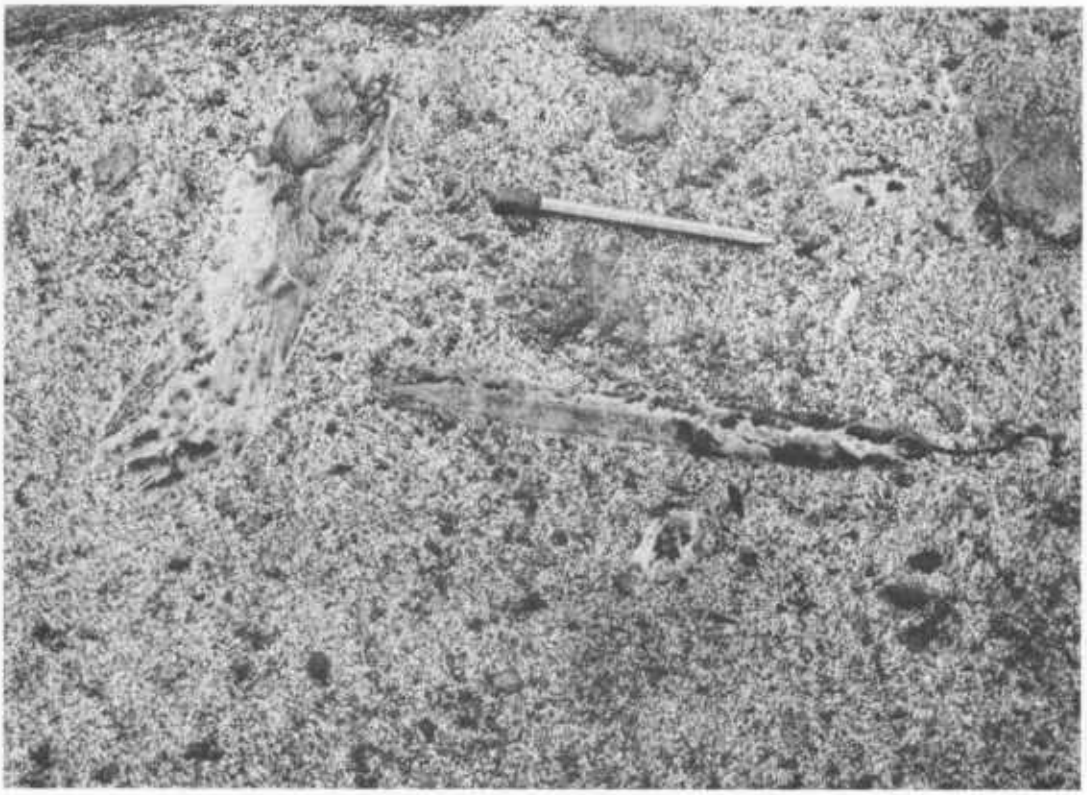
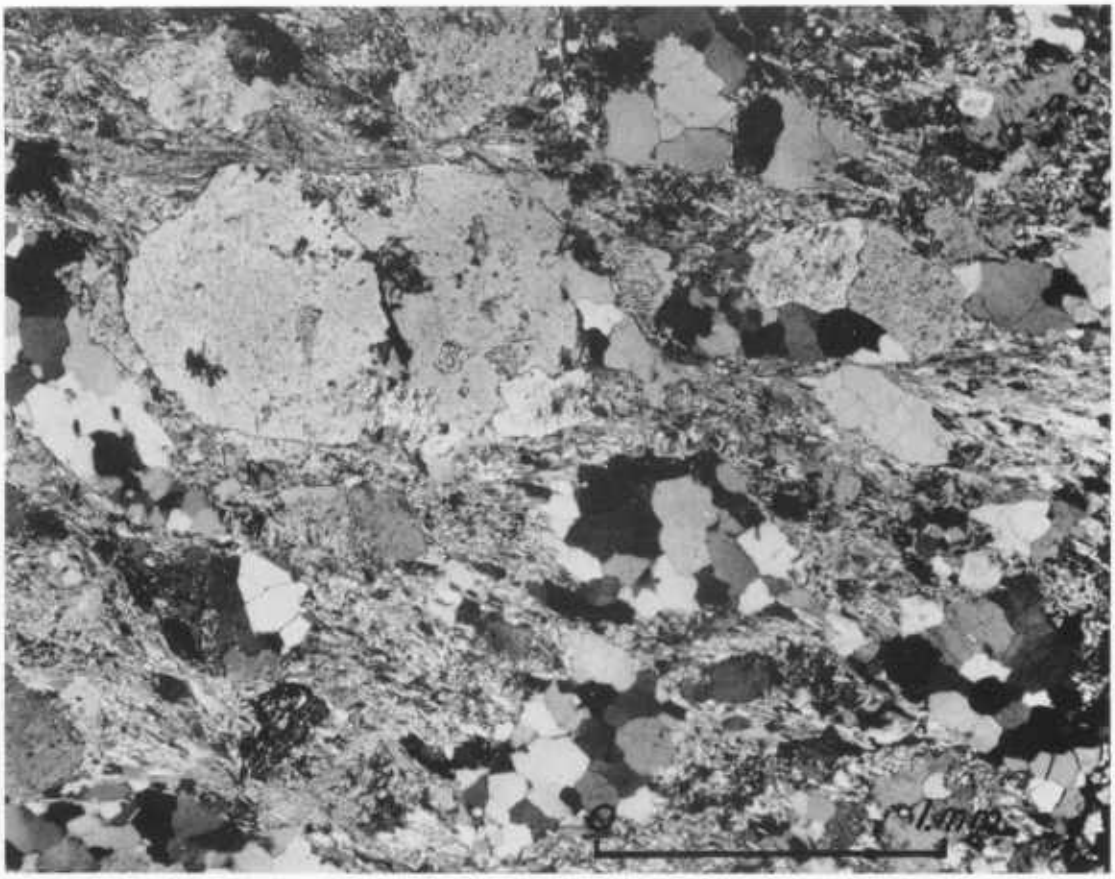


PLATE 25, *Figure 1* (upper). Photomicrograph, showing relict clastic grains in rock of the Sykesville Formation. Two rounded rock particles, each comprised of several grains of feldspar and quartz, are shown in the upper half of the picture. In the lower half are several rounded quartz grains, now recrystallized to mosaics of smaller crystals. The matrix is finely crystalline muscovite, biotite, chlorite, quartz, albite, and epidote. Cabin John Creek near Bradley Boulevard, Montgomery County. Crossed nicols.

Figure 2 (lower). Photomicrograph, showing relict clastic texture in rock of the Sykesville Formation. Slightly deformed quartz grains still show rounded detrital outlines. The detrital grains formed a partly unsupported framework, in a matrix of argillaceous composition (now chiefly micas, chlorite, and fine quartz). The dark mineral at the upper right is garnet. Howard County, near Sykesville. Plane light.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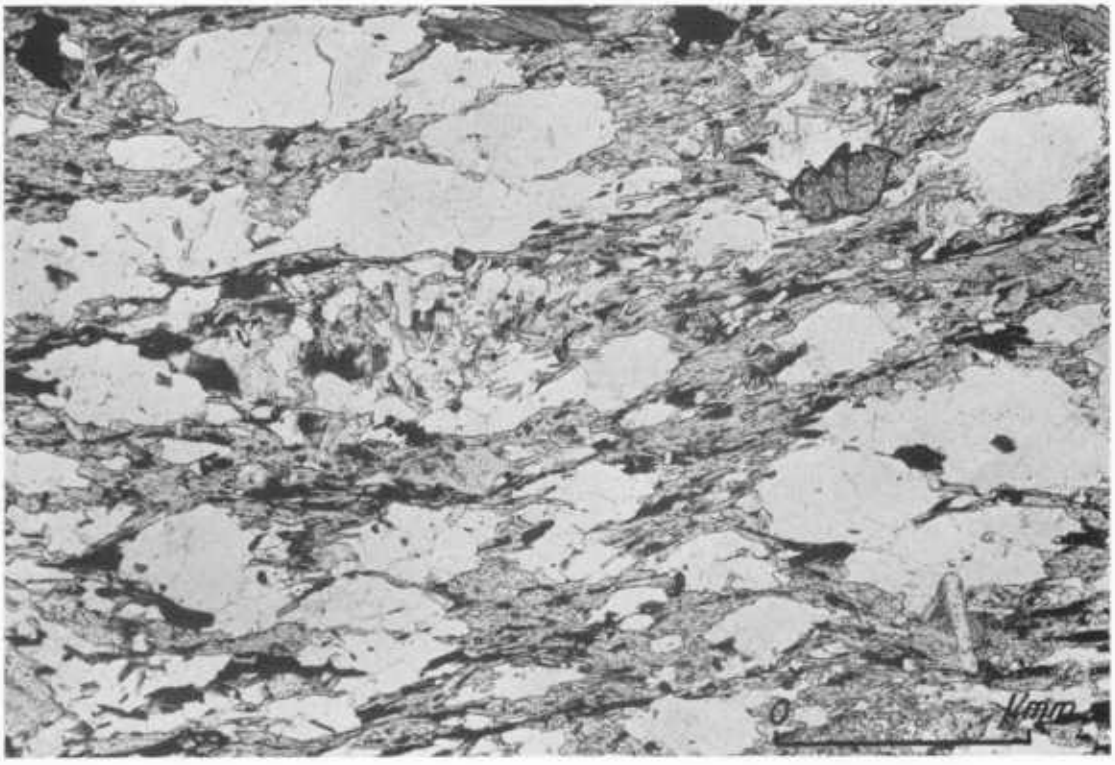
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PLATE 26, *Figure 1* (upper). Photomicrograph of Ijamsville Phyllite, showing chloritoid porphyroblasts. The matrix consists chiefly of muscovite, with small amounts of paragonite, chlorite, and magnetite. Plane light. Specimen from 1½ miles southwest of Damascus, Montgomery County.

Figure 2 (lower). Ripple marks, cast on the base of a metasandstone bed in the Harpers Phyllite. Little Monocacy River southeast of Dickerson, Montgomery County.

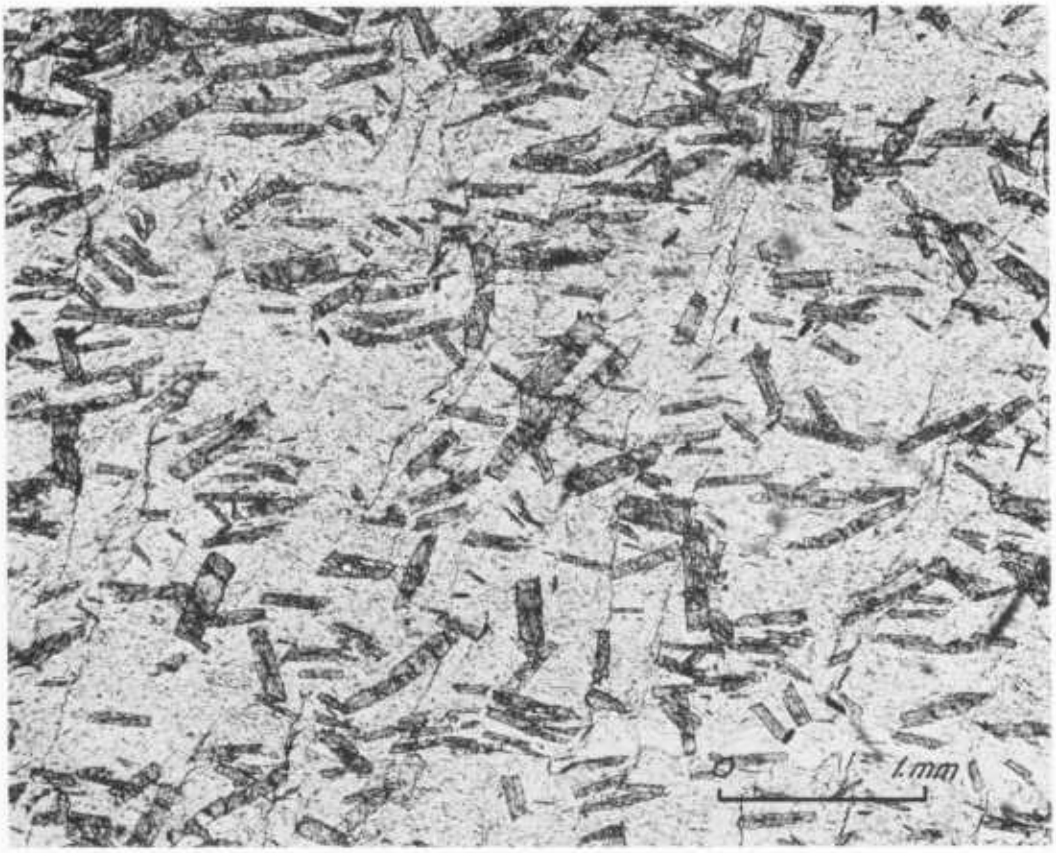
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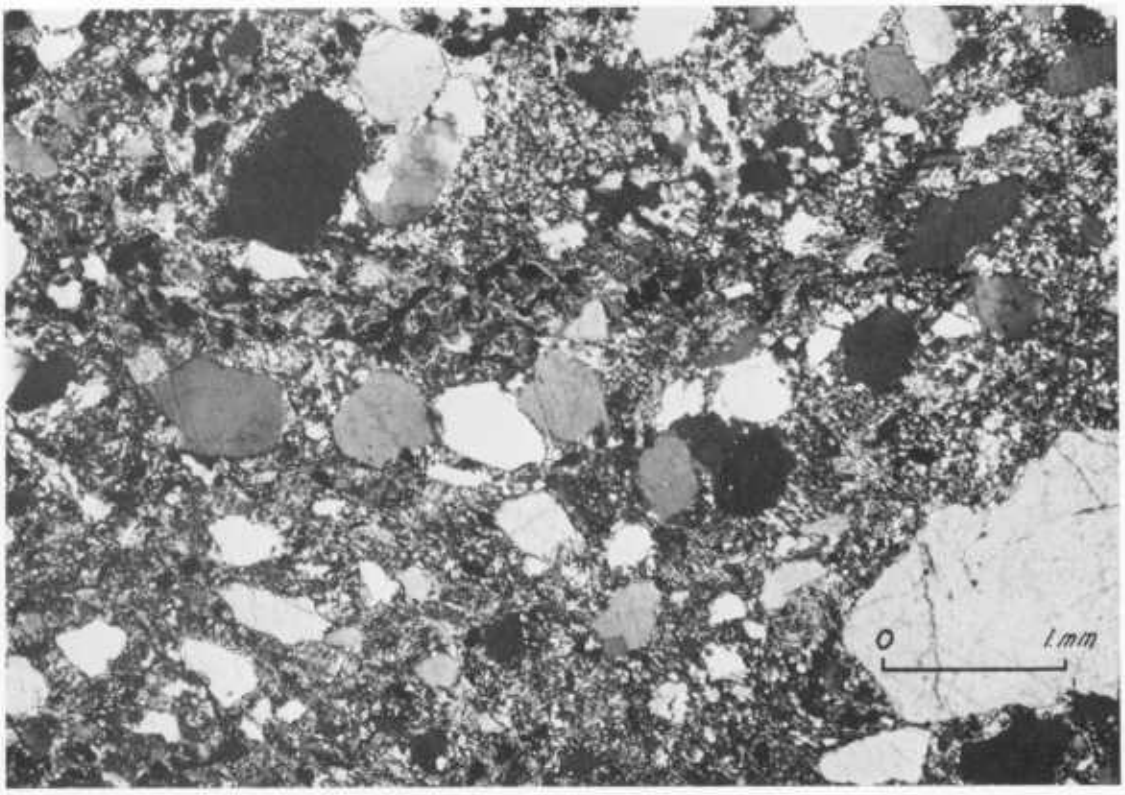
PLATE 27, *Figure 1* (upper). Photomicrograph of metagraywacke, showing relict clastic texture. Angular to subrounded quartz grains, in a matrix of fine-grained chlorite, sericite, stilpnomelane, quartz, albite, and magnetite. Crossed nicols. Harpers Phyllite, at Little Monocacy River, southeast of Dickerson.

Figure 2 (lower). Photomicrograph of orthoquartzite, from a bed about 40 feet thick in the Harpers Phyllite. Well-rounded quartz grains, in a recrystallized quartz matrix. One mile southeast of Dickerson, Montgomery County.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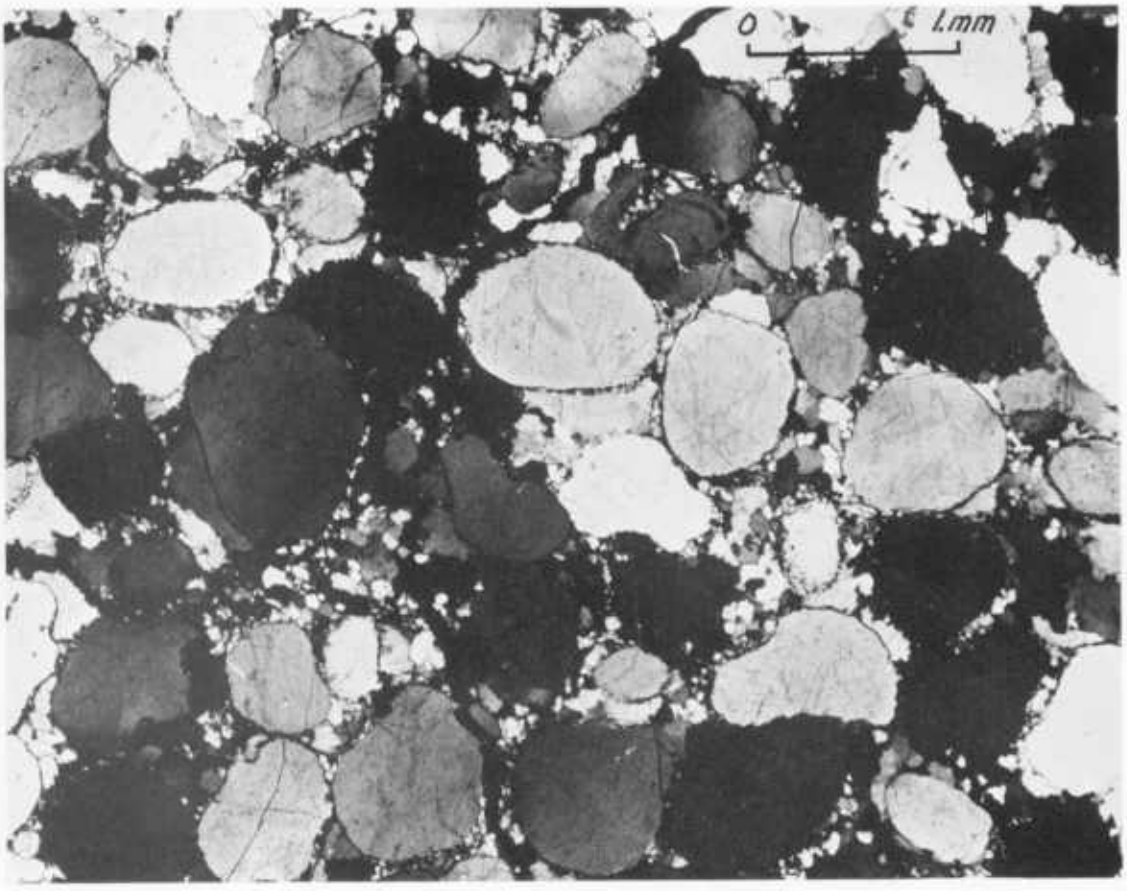
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PLATE 28, *Figure 1* (upper). Specimen of uralite gabbro, showing rhythmic layering. The dark layer of mafic gabbro in the middle grades upward into feldspathic gabbro. The scale is 1 inch. Baltimore Gabbro Complex, near the Baltimore Beltway at Dogwood Road.

Figure 2 (lower). Rhythmically layered ultramafic rocks of the Baltimore Gabbro Complex. The light layers are metapyroxenite, the dark layers metaperidotite. Quarry near Patapsco River bridge, south of Hollofield, Baltimore County.

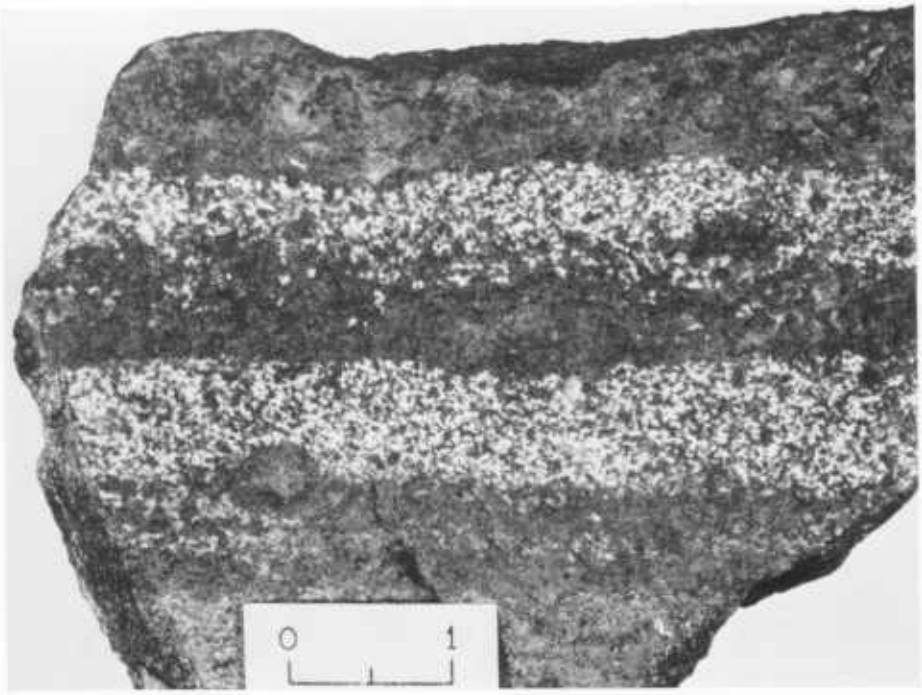
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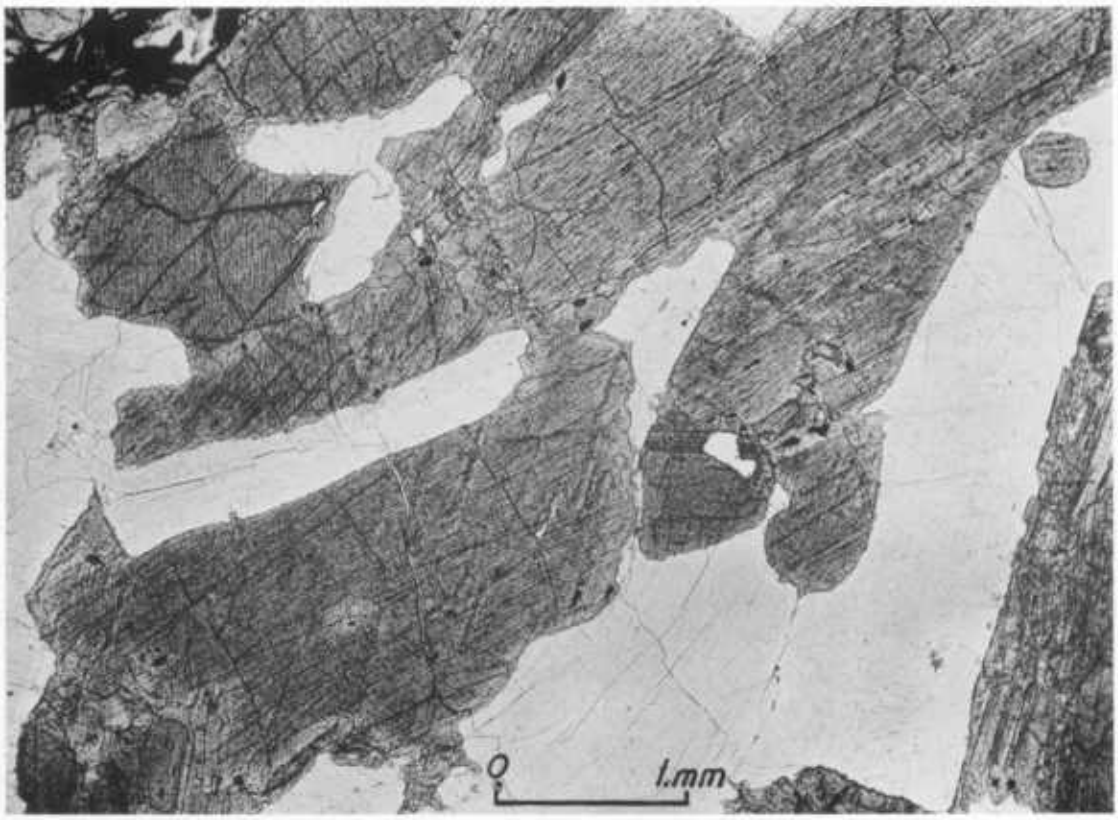
PLATE 29, *Figure 1* (upper). Feldspathic net veins in metagabbro. The net veins cut across a weak protoclastic foliation, which runs parallel to the pencil. Baltimore Gabbro Complex at Gwynns Falls, half a mile east of Franklintown, Baltimore City.

Figure 2 (lower). Amphibolite dike cutting metagabbro. Note the feldspathic margins, and feldspathic schlieren inside the dike. The pencil lies parallel to weak foliation in the metagabbro. Baltimore Gabbro Complex, Gwynns Falls east of Franklintown. The rocks shown in Figures 1 and 2 are located about 40 feet apart.

PLATE 30, *Figure 1* (upper). Photomicrograph of relict gabbro, showing igneous texture. The dark mineral is chiefly clinopyroxene, and the light mineral plagioclase (An_{80}). Baltimore Gabbro Complex, 1½ miles southeast of Ellicott City, Howard County. Plane light.

Figure 2 (lower). Photomicrograph showing part of a relict bronzite oikocryst in metaperidotite. The bronzite (Br) poikilitically encloses euhedral to rounded olivines (Ol). The olivines are partly serpentinized, and serpentine veins (Sp) cut through the bronzite. Some of the less obvious olivine contacts are brought out with ink. Baltimore Gabbro Complex, Johnny Cake road west of Belmont, Baltimore County. Plane light.


O
1.mm

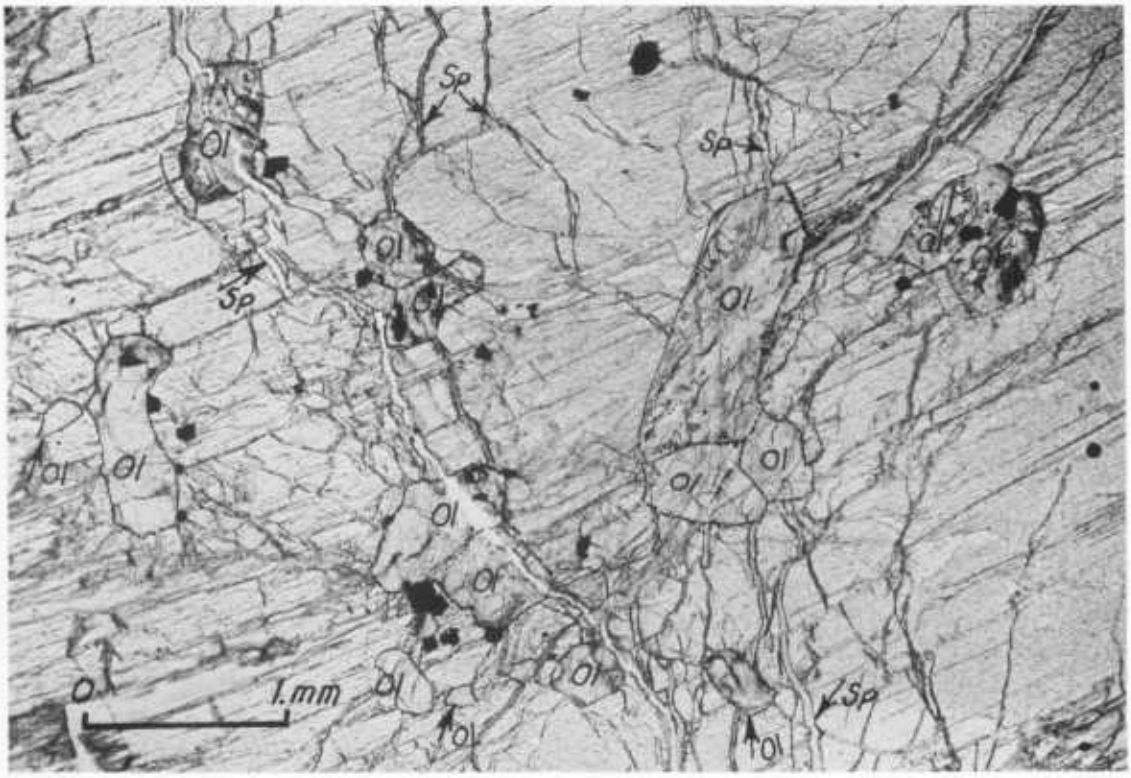
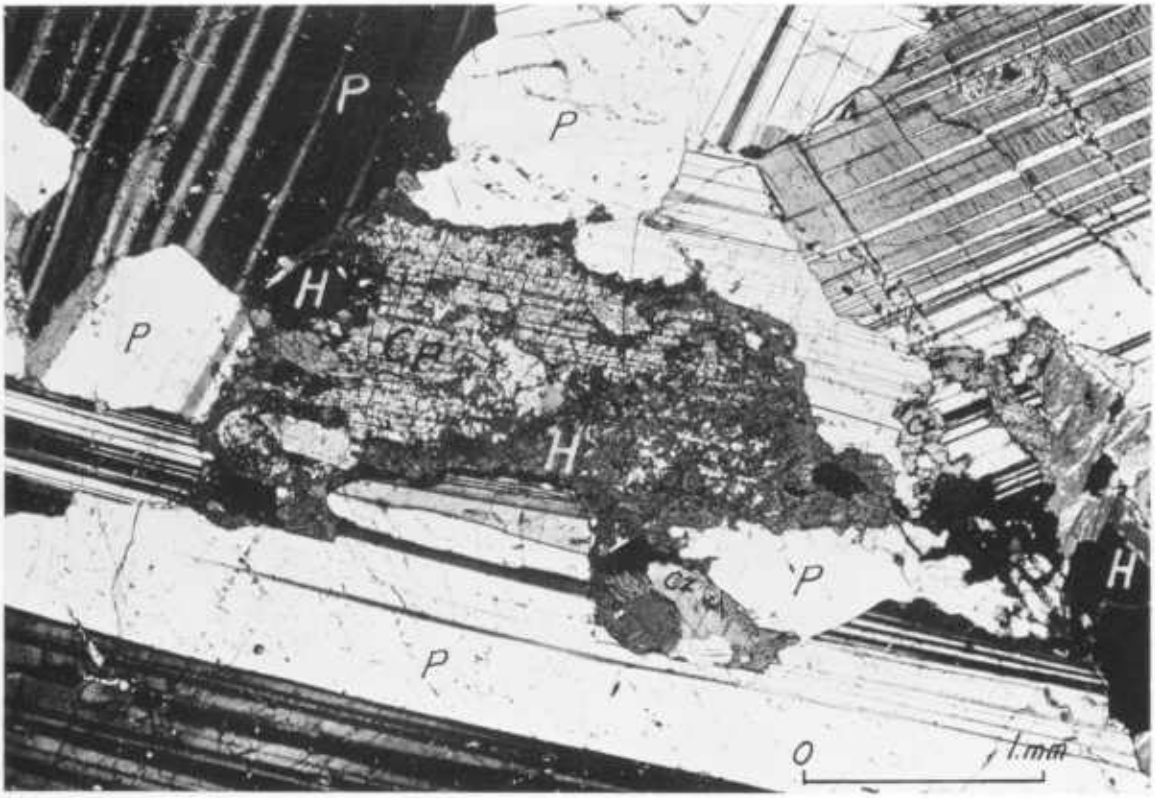

Sp
Sp
Sp
Sp
Ol
Ol
Ol
Ol
Ol
Ol
Ol
Ol
Ol
Ol
Ol
Ol
Ol
Ol
O
1.mm

PLATE 31, *Figure 1* (upper). Photomicrograph of uralitized gabbro, showing relict clinopyroxene (Cp) in the core of a green uralitic hornblende (H). The other minerals are plagioclase (P) and clinozoisite (Cz). Baltimore Gabbro Complex, 1 mile south of Ellicott City, Howard County. Crossed nicols

Figure 2 (lower). Photomicrograph of uralite gabbro. Clusters of green hornblende have replaced most of the original grains of coarse pyroxene, but a few clinopyroxene relics (Cp) are visible. The original igneous texture, with peri-euhedral plagioclase, is still evident. Baltimore Gabbro Complex, 1 mile southeast of Ellicott City, Howard County. Plane light.


P
P
H
P
Cp
H
Cz
P
P
H
0
1.mm

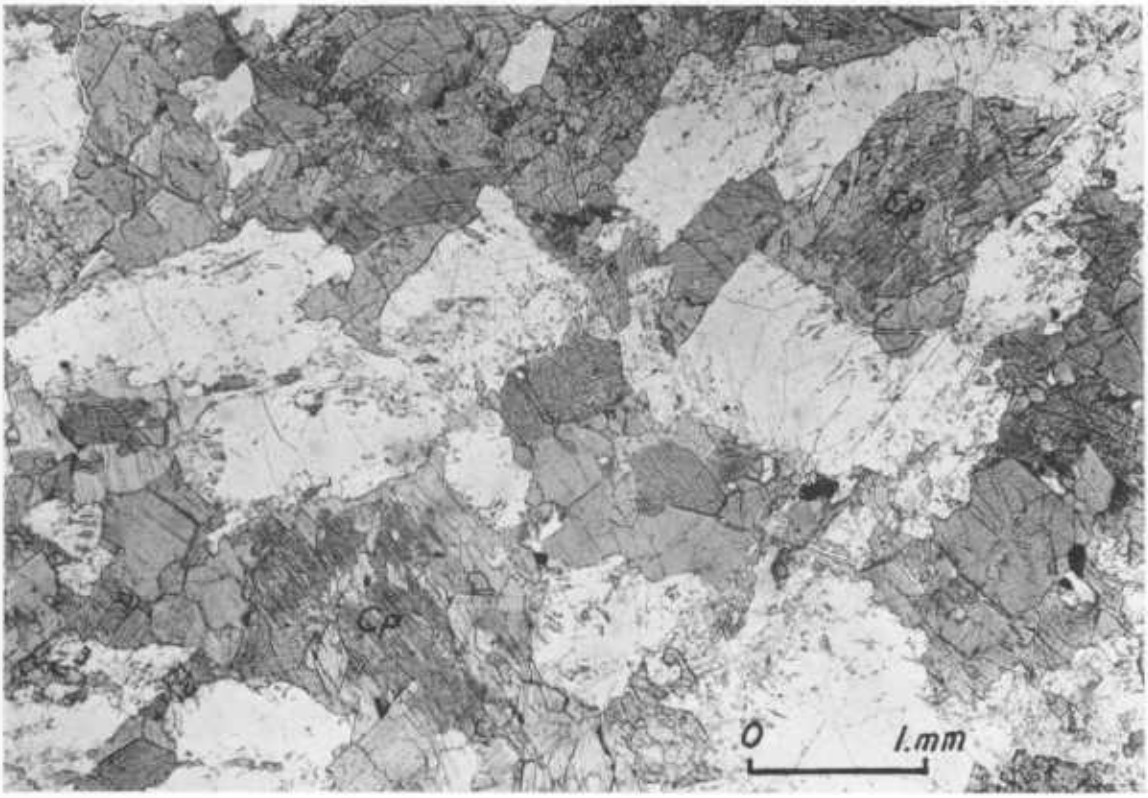
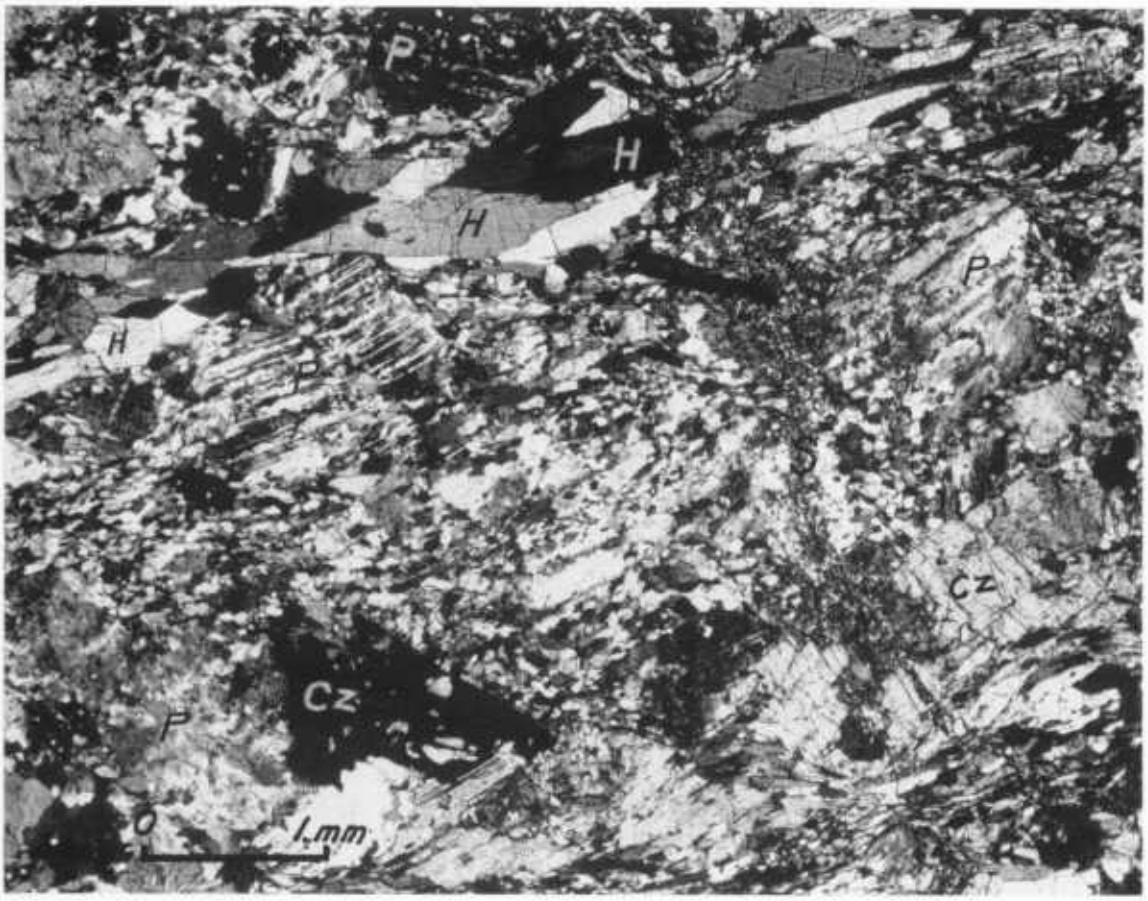

Cp
Cp
0
1.mm

PLATE 32. *Figure 1* (upper). Photomicrograph of gneissic amphibolite. Badly shattered relict plagioclase (Pl) lies in a matrix of finely granulated and recrystallized plagioclase and green hornblende. Large undeformed hornblendes (H) lie along the foliation, and poikioblastic clinozoisite (Cz) partly replaces the granulated matrix. A late cataclastic shear zone (S) cuts across the foliation. Baltimore Gabbro Complex, Patapsco State Park near Orange Grove. Crossed nicols.

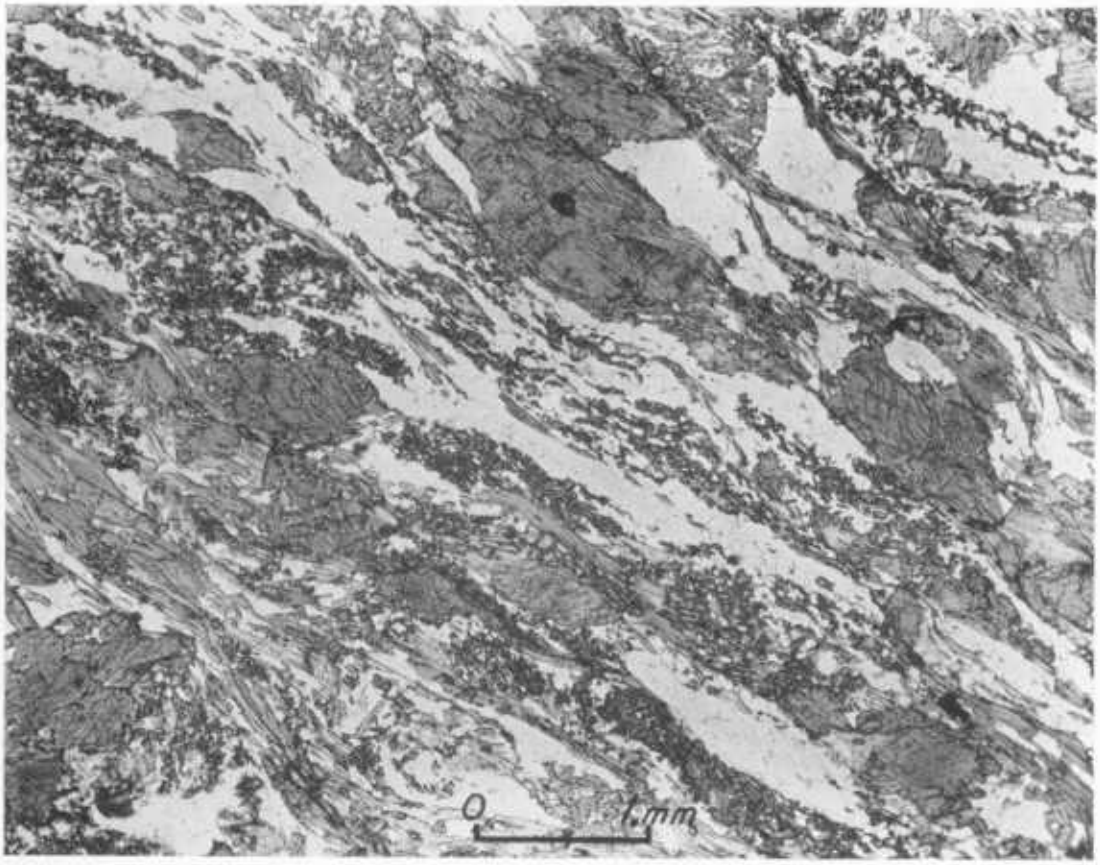
Figure 2 (lower). Photomicrograph of albite granite, showing granoblastic texture. An albite-quartz intergrowth is visible at the upper right. Relay Quartz Diorite at Relay, Baltimore County. Crossed nicols.


P
H
H
H
P
P
Cz
Cz
P
1mm

1mm

PLATE 33, *Figure 1* (upper). Photomicrograph of Norbeck quartz diorite, showing the gneissic texture. The large dark minerals are hornblende, the flakes are biotite, the dark granular aggregates are clinozoisite in plagioclase, and the light areas chiefly quartz. Two miles south of Norbeck, Montgomery Co. Plane light.

Figure 2 (lower). Photomicrograph of Kensington quartz diorite, showing the gneissic texture. Trains of biotite and muscovite alternate with xenoblastic aggregates of feldspar and quartz. The large mineral above the center is relict plagioclase. Rock Creek Park, Washington, D. C. Crossed nicols.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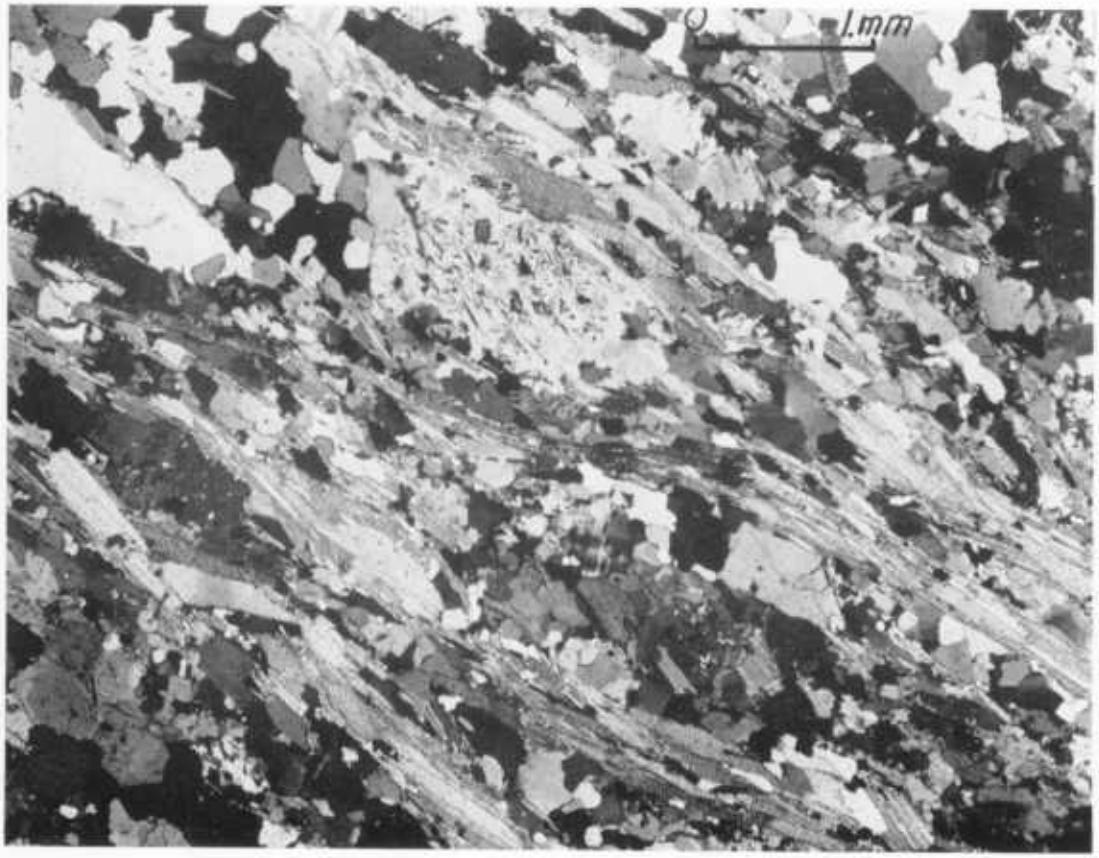
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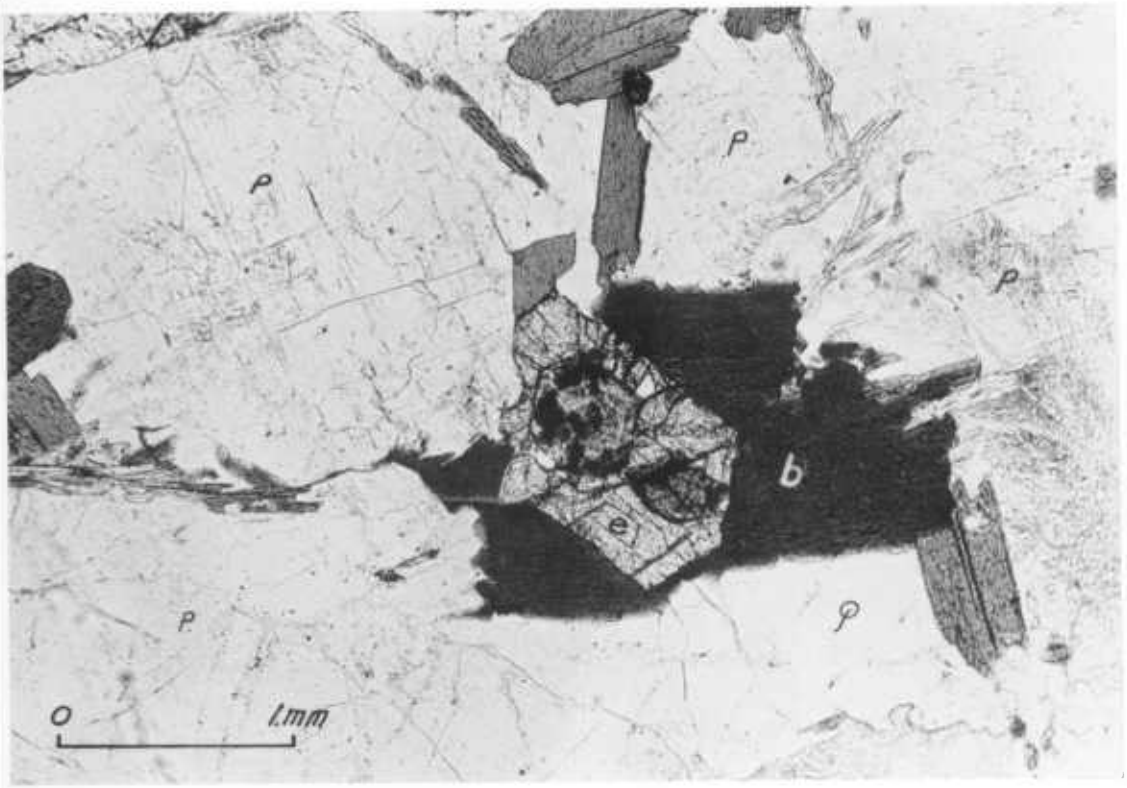
1.mm

PLATE 34, *Figure 1* (upper). Photomicrograph of Ellicott City granodiorite, showing primary epidote. The euhedral epidote (e) contains euhedral zoned allanite. Biotite (b) encloses the epidote. The other minerals are plagioclase (p) and quartz (q). Plane light.

Figure 2 (lower). Inclusion of migmatitic Baltimore gneiss in the Ellicott City Granodiorite. The gneiss forms a sinuous mass over 20 feet long, only a short part of which is shown. Smaller schlieren of similar rock are common in the surrounding granodiorite. Ellicott City, about 400 feet east of the Patapsco River bridge.


p
p
p
b
e
p
p
0
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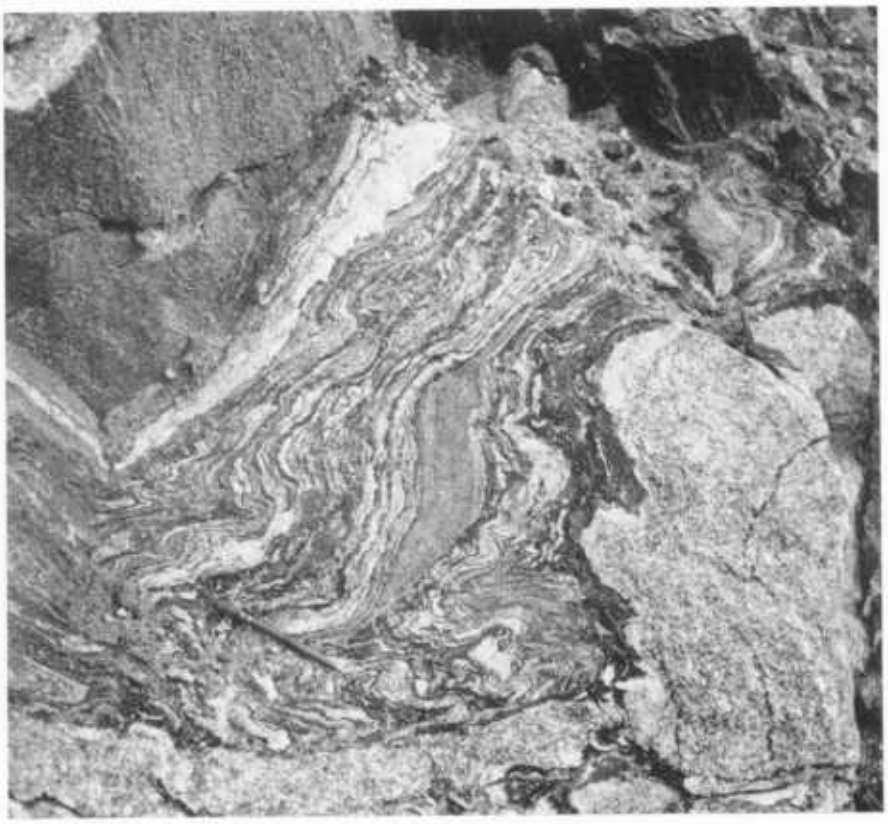
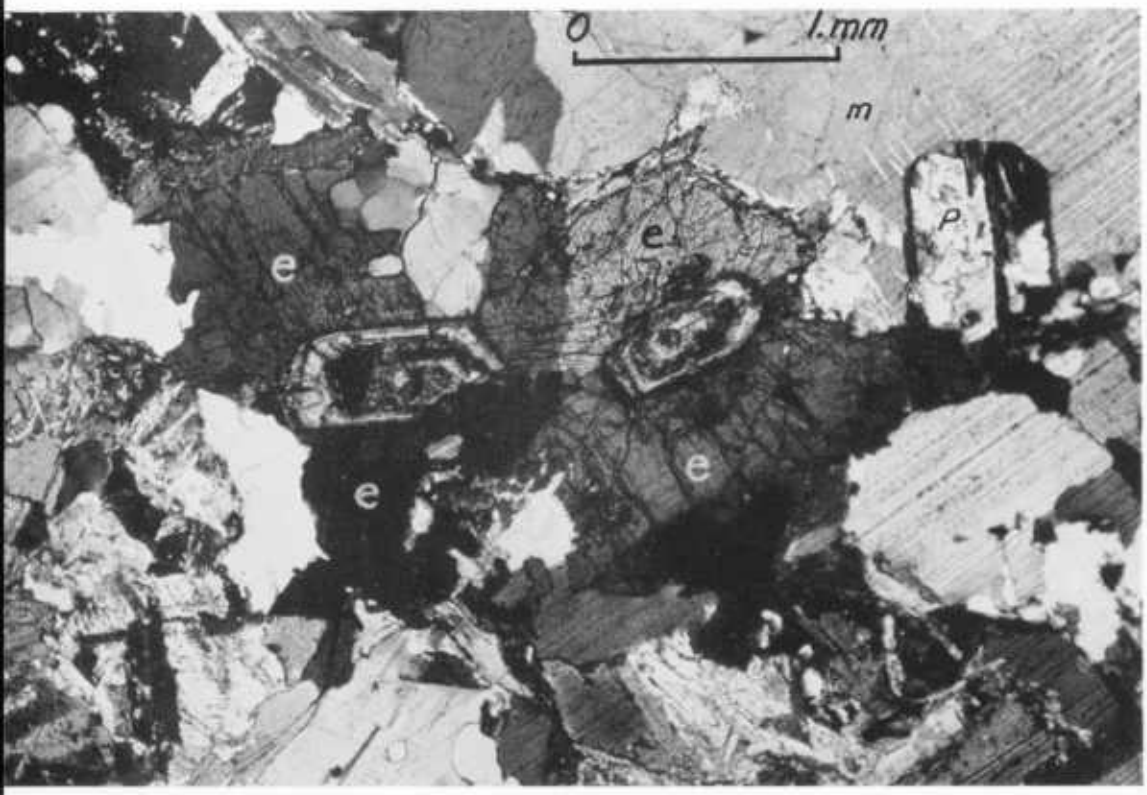


PLATE 35, *Figure 1* (upper). Photomicrograph of Woodstock quartz monzonite, showing two euhedral zoned allanites, enclosed by epidote (e). Microcline (m) at the upper right is locally perthitic where it encloses euhedral plagioclase (p). (The striations are on the glass slide.) Crossed nicols.

Figure 2 (lower). Photomicrograph of massive Woodstock quartz monzonite, showing hypidiomorphic texture. Anhedral quartz (q) and microcline (m) fill the spaces between zoned plagioclase (p). This texture is characteristic of the late kinematic rocks. Crossed nicols.


0
1.mm
m
p
e
e
e
e
e

m
p
p
q
q
p
p
m
m
q
p
m
0
1.mm

PLATE 36, *Figure 1* (upper). Well-bedded Wissahickon schist below Anglers Inn, Potomac River. Light and dark bands are cut by an axial plane cleavage (S_3). Quartz veinlets follow cleavage and bedding in which case they are folded. Height of outcrop about 4 feet.

Figure 2 (lower). Wissahickon schist below Anglers Inn, Potomac River. Cleavages are bedding schistosity (S_2) and axial plane cleavage (S_3) with much displacement on S_3.

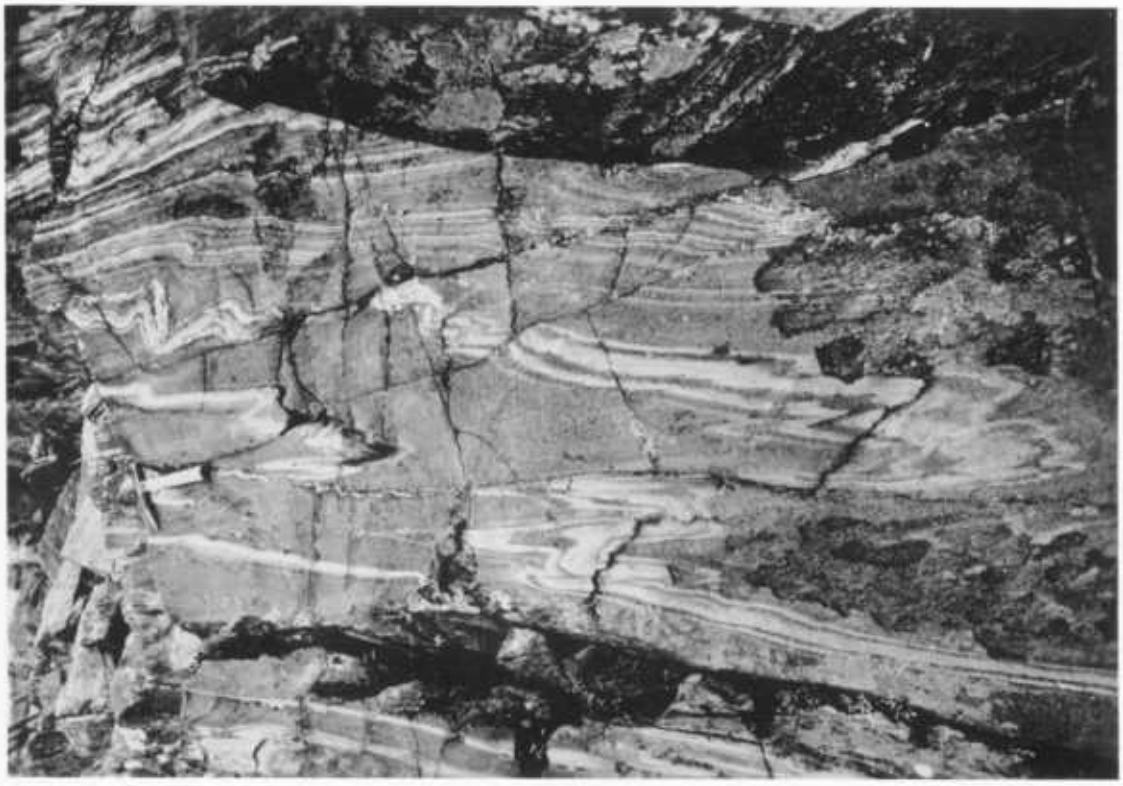
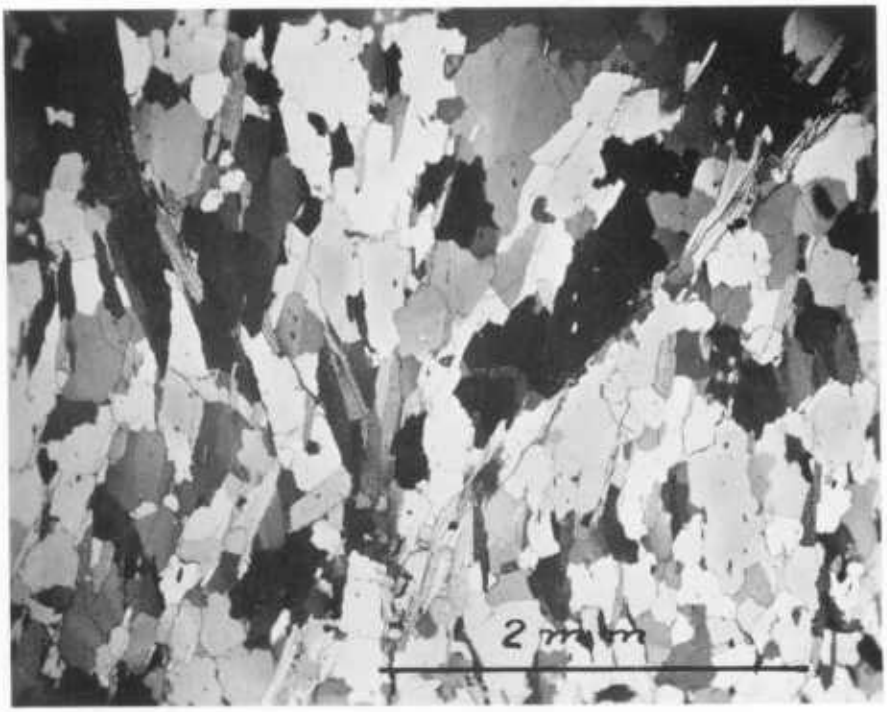


PLATE 37, *Figure 1* (upper). Discordant Bear Island granite dike (light) cuts across schist but weathers and erodes more readily. Opposite Anglers Inn, Potomac River, Virginia side.

Figure 2 (lower). Discordant granite cuts strongly lineated schist along a sharp contact. The granite is practically structureless. Bear Island, Potomac River.

PLATE 38, *Figure 1* (upper). Photomicrograph of Setters quartzite near Alberton, Maryland. Compare with drawing, Figure 52. Large mica grains follow bedding in left half of photograph and smaller ones in right limb of syncline. In fold crest large and small micas parallel the axial plane. Above and below bedding plane smaller muscovite flakes also parallel axial plane cleavage (S_3).

Figure 2 (lower). Cleavage (S_3) cuts bedding in Harpers Phyllite. No bedding schistosity visible. Quartz grains in light layer are lenticular due to shearing on axial plane cleavage which is here a fracture cleavage.


2 mm

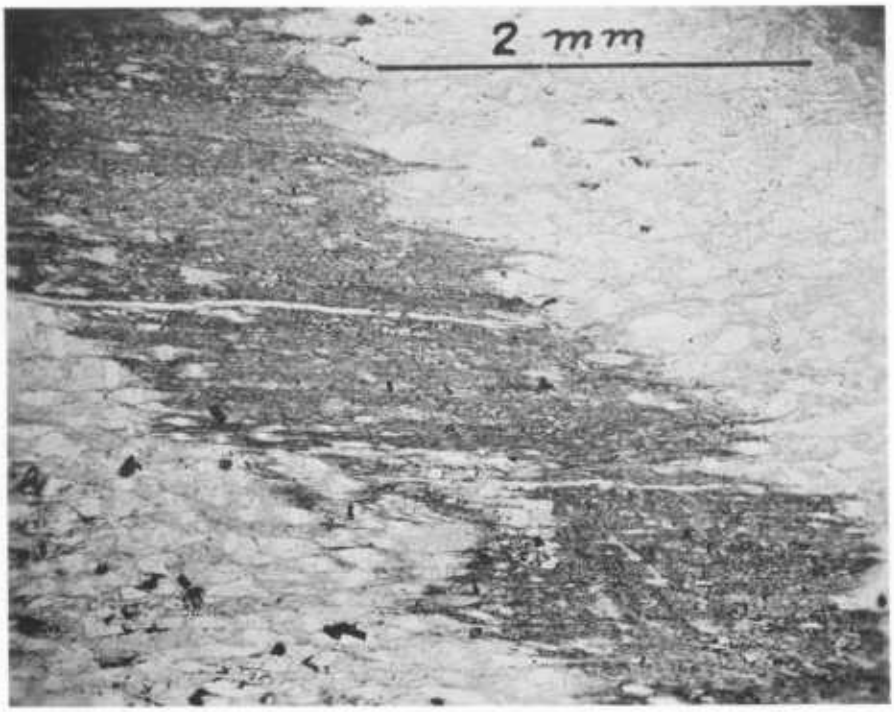
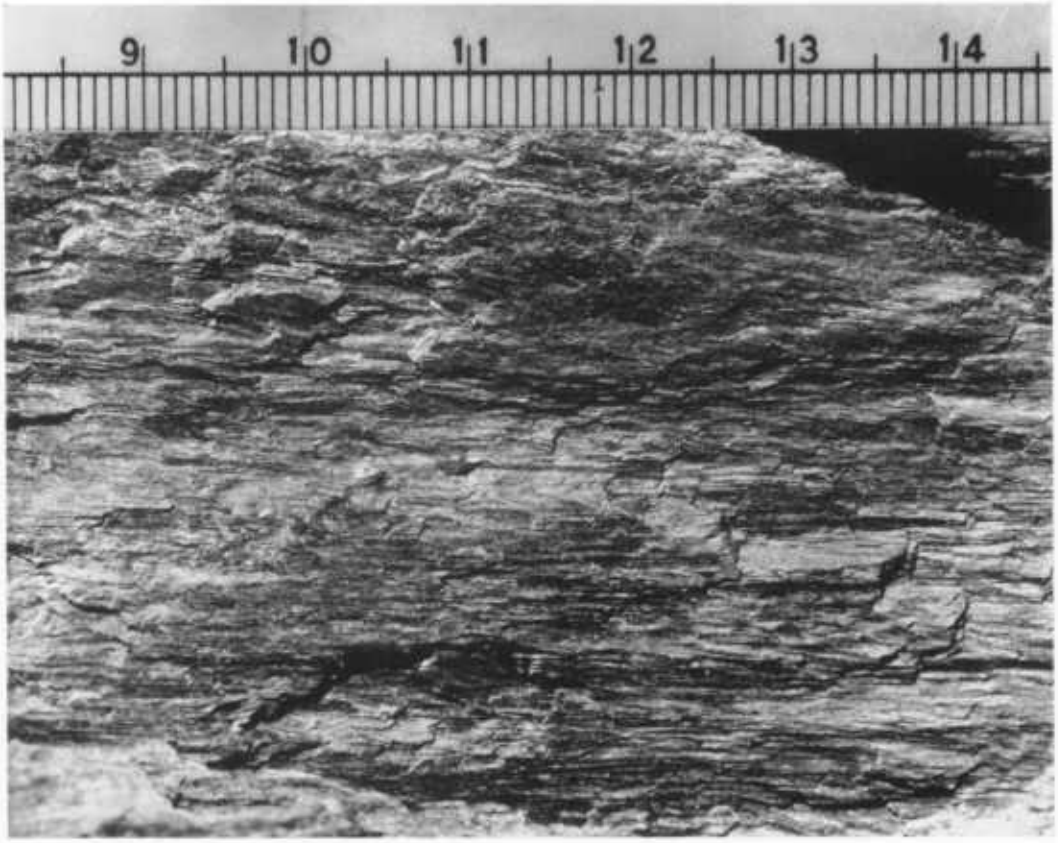

2 mm

PLATE 39. Folds in Wissahickon schist, western sequence. Intensely folded bedding with mica growing parallel as bedding schistosity (S_2). In fold crests mica cuts across bedding where movement (shearing) is intense. Small mica follows bedding around the fold, larger ones tend to grow in axial plane cleavage direction (S_3).

3 mm

PLATE 40, *Figure 1* (upper). Lineated cleavage surface in phyllite. Wissahickon, western sequence, Seneca Creek. Two directions intersect at low angle in left half of photograph. Fractures cut lineation at angle of 65°–75°.

Figure 2 (lower). Lineated bedding surface in quartzite. Rough fractures cut across lineation at about 70°.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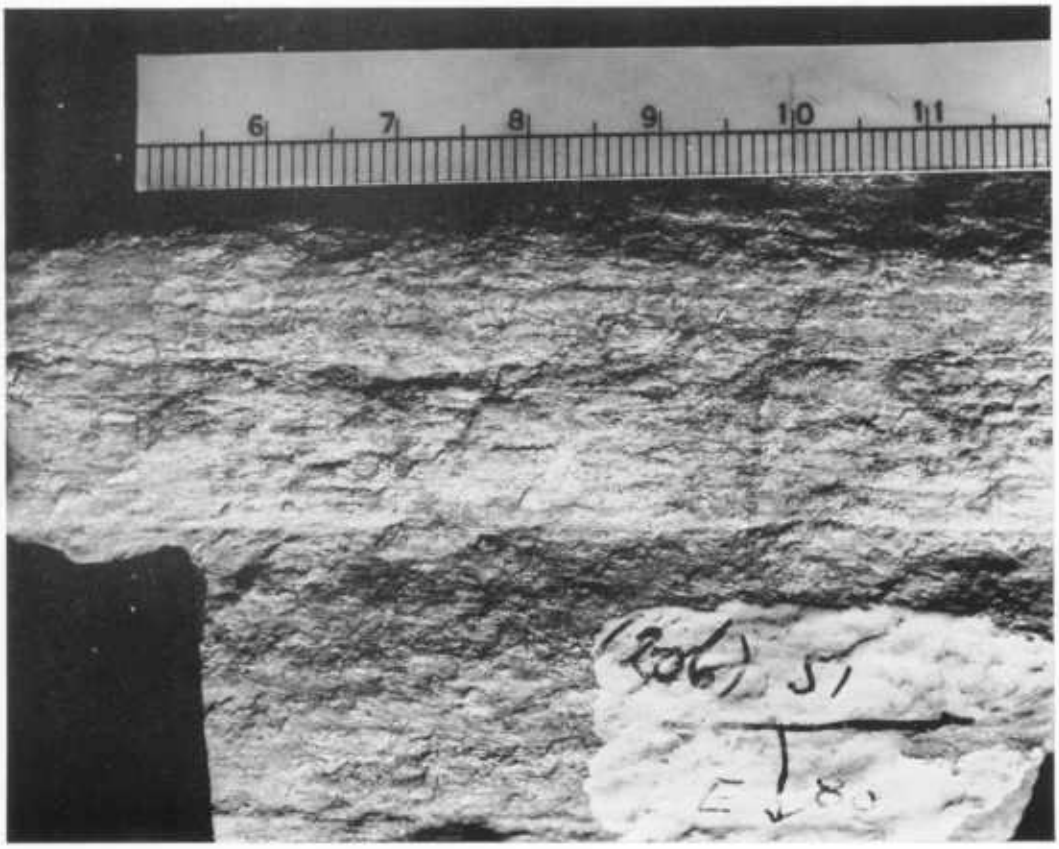
6
7
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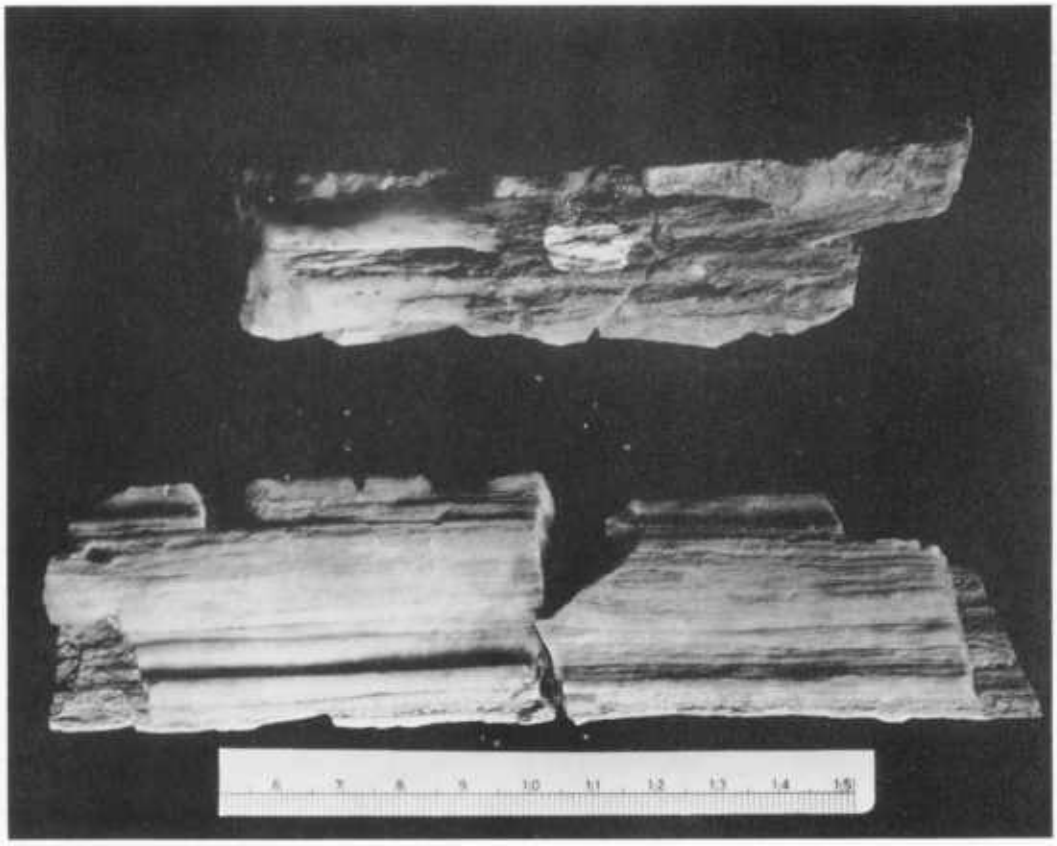
PLATE 41, *Figure 1* (upper). Quartz rods from Wissahickon schist, western sequence, Seneca Creek. Lower rod is cast of a small fold crest. Rods are parallel fold axes.

Figure 2 (lower). Detail of rod in Figure 1. Tension fractures in quartz are irregular but suggest a conjugate system including an angle of about 50 to 60 degrees. Rods are molds of small folds.



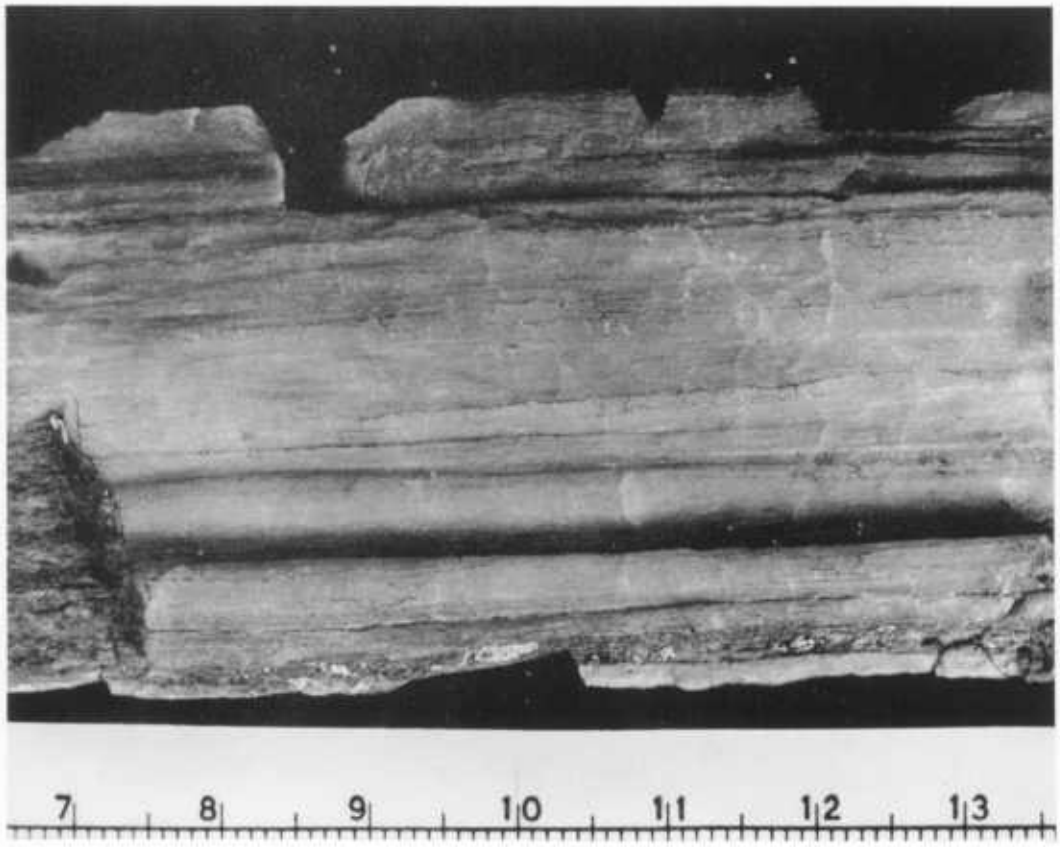

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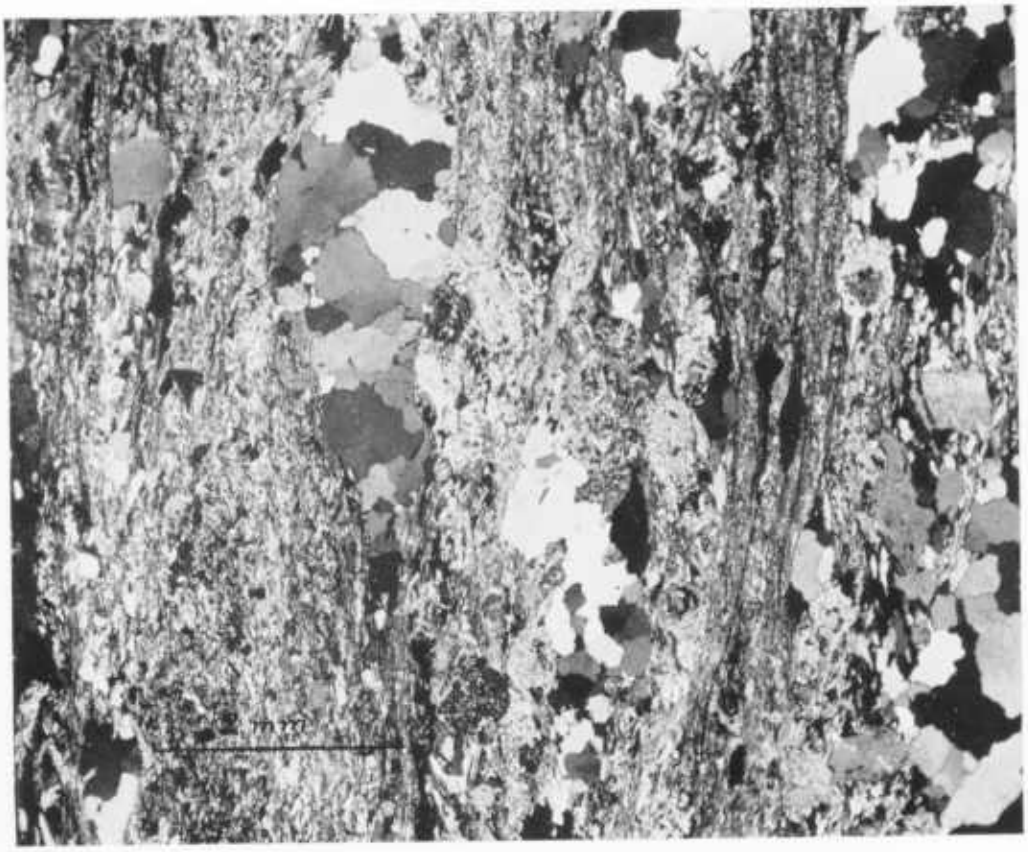
PLATE 42, *Figure 1* (upper). Microphotograph, Sykesville Formation parallel to lineation, normal foliation. Quartz clusters are former pebbles as shown in Figure 62. Foliation is intense and mostly followed by chlorite. Small muscovite concentrate in less foliated areas. For mica orientation see Figure 60.

Figure 2 (lower). Photomicrograph, Sykesville Formation. Normal lineation 2 inches from Figure 1 of Plate 42. Foliation much less distinct. Micas, especially muscovite, are very small and swirl around quartz pebbles. Pebbles much less stretched than in Figure 1. Sketch of slide normal, Figure 62; Diagram of mica orientation, Figure 59.


2 mm

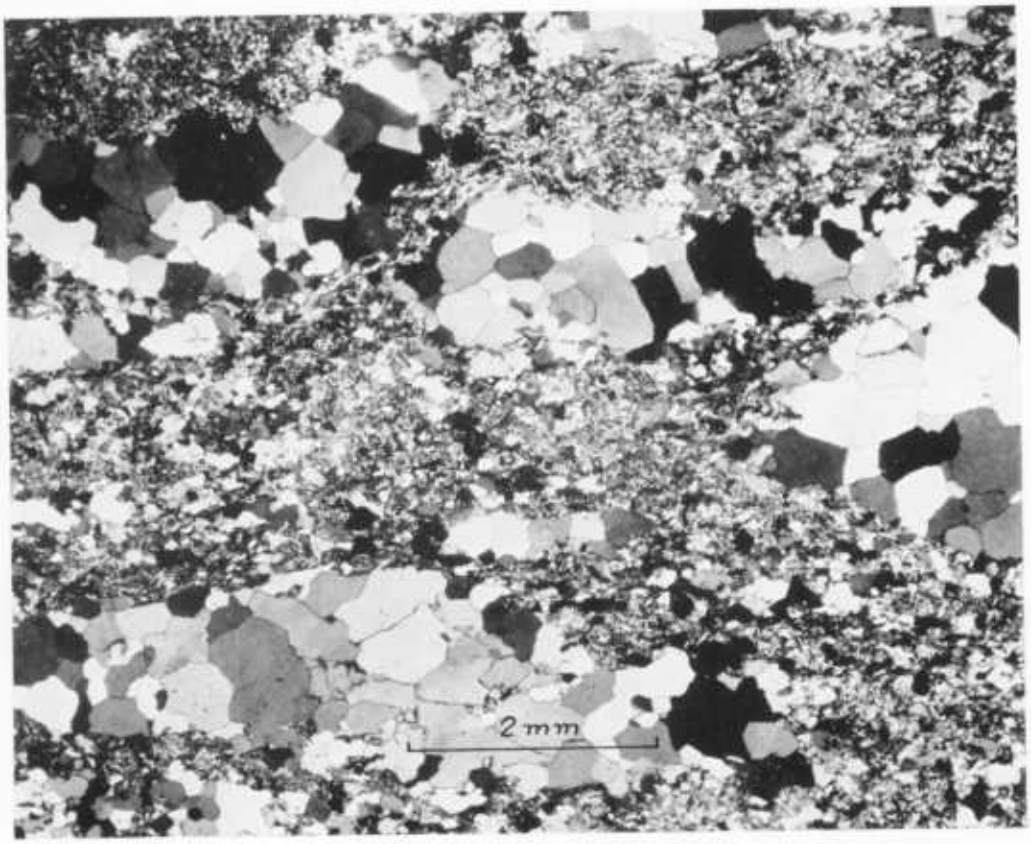
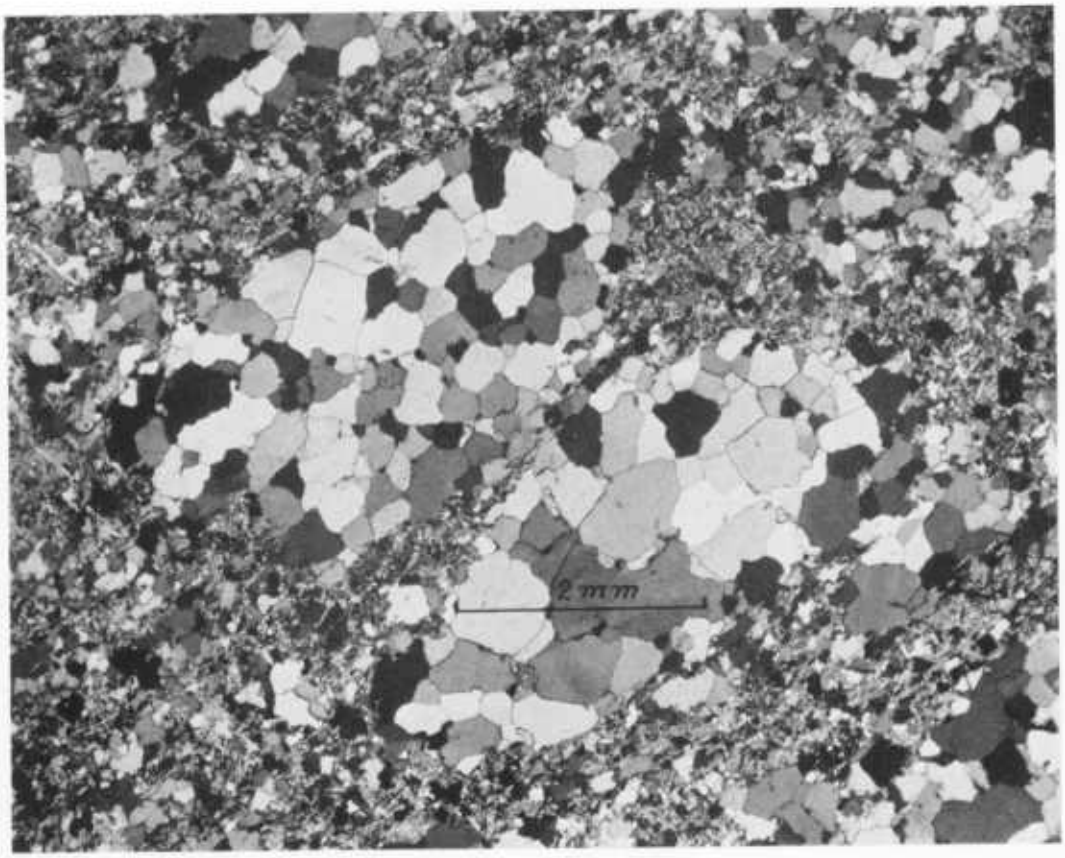

2 mm

PLATE 43, *Figure 1* (upper). Quartz pebbles in Wissahickon Formation, western sequence for comparison with pebbles in Figure 2 of Plate 42.

Figure 2 (lower). Hand specimen surface of amphibolite, Bear Island. Normal lineation. A vague foliation is visible diagonally across photograph.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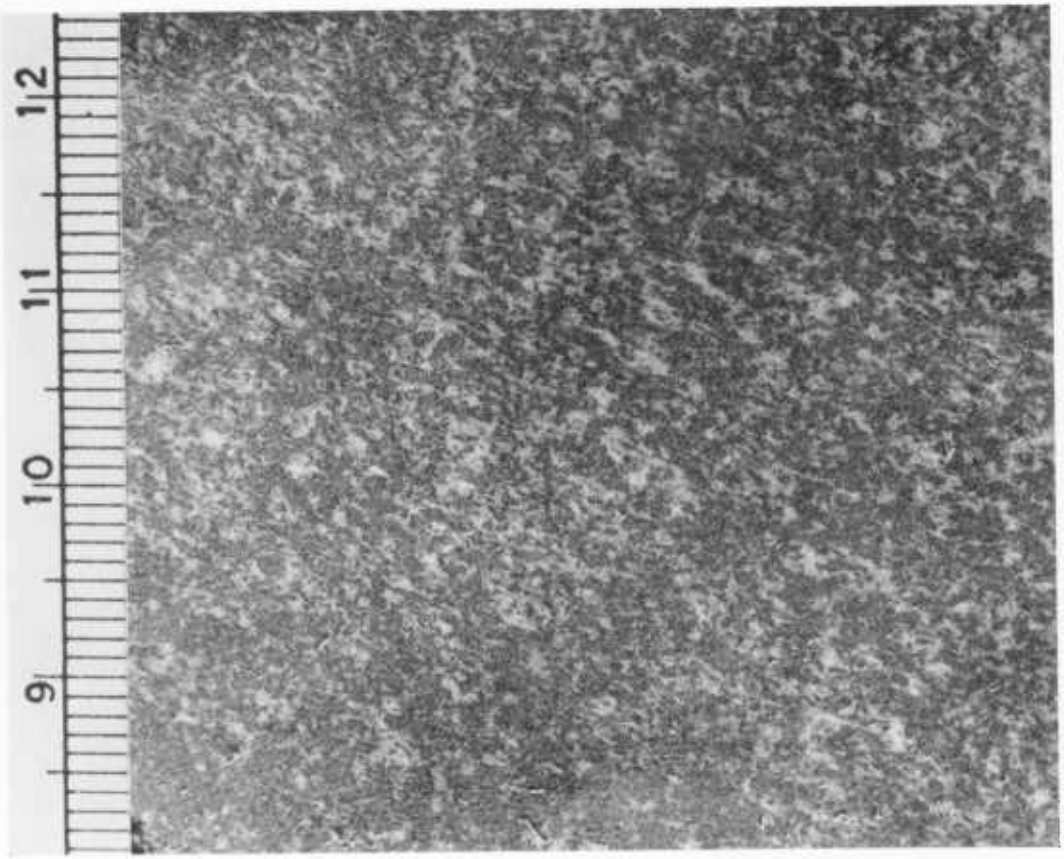
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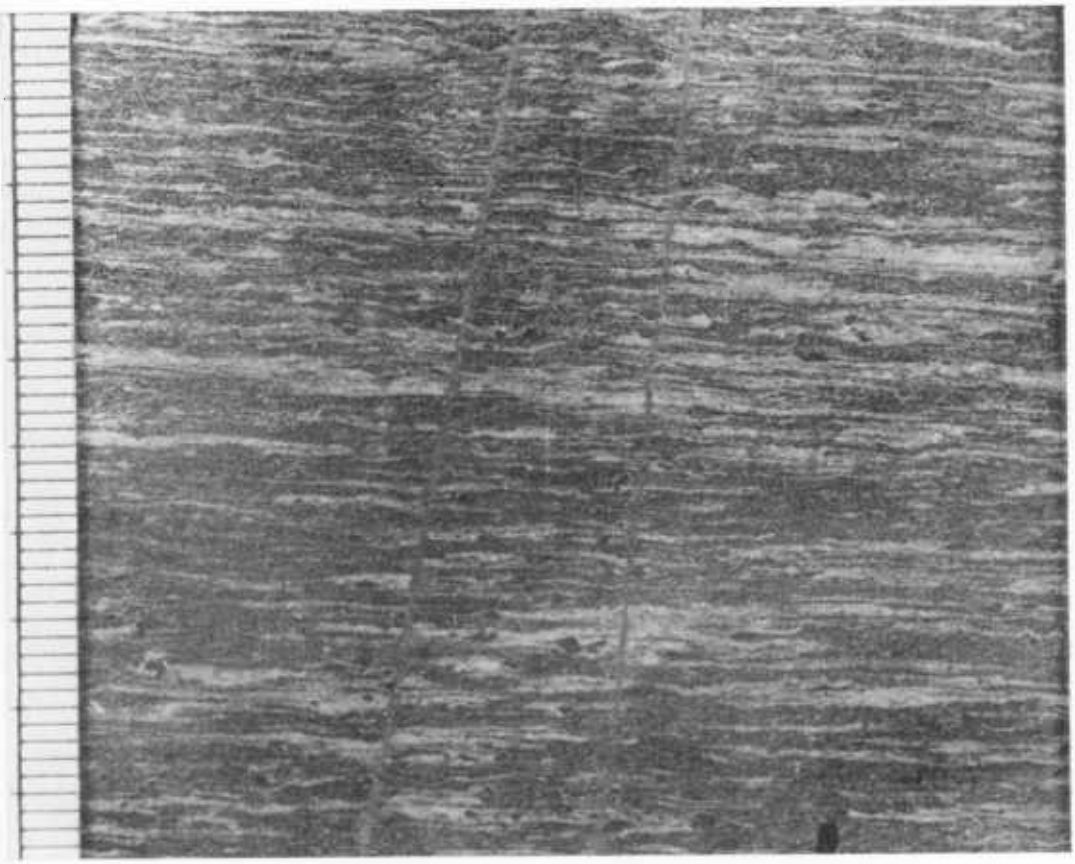
PLATE 44, *Figure 1* (upper). Same specimen as Figure 2 of Plate 43 cut parallel lineation. Strong lineation is self evident. Two dark lines across lineation are fractures. Bear Island, Potomac River.

Figure 2 (lower). Photomicrograph of Figure 2 of Plate 43. A feeble foliation and swirling orientation of dark constituents.



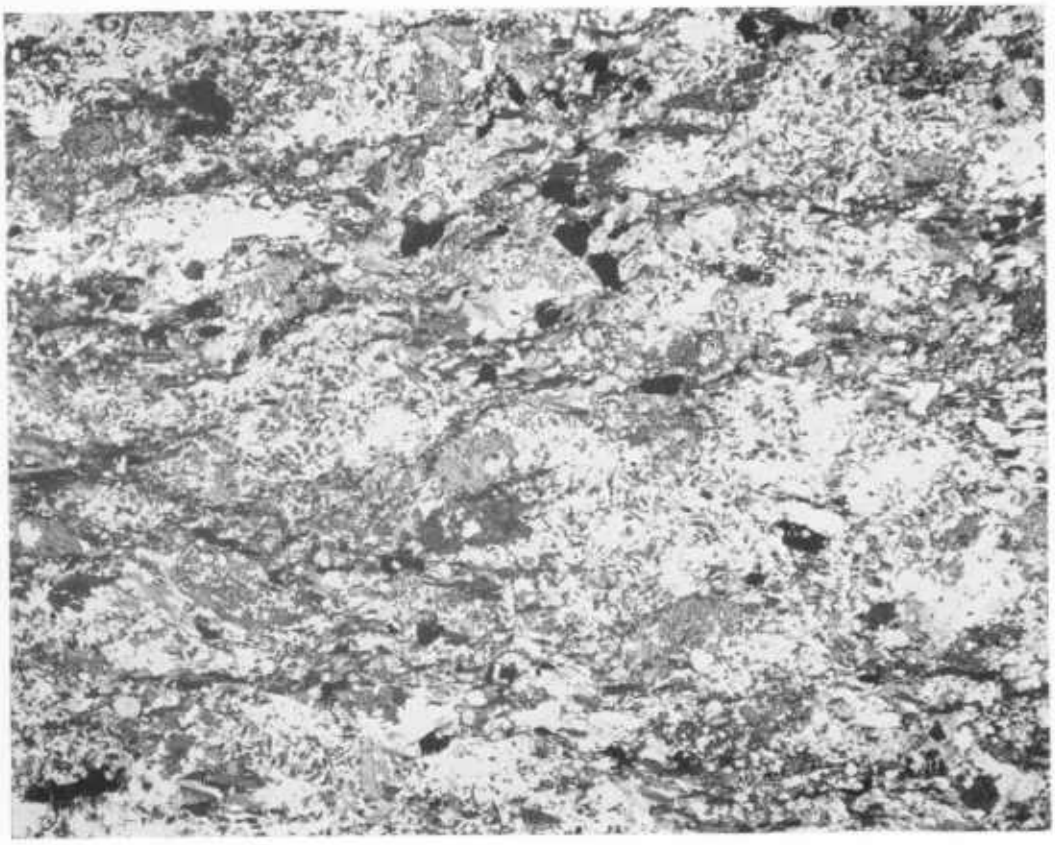
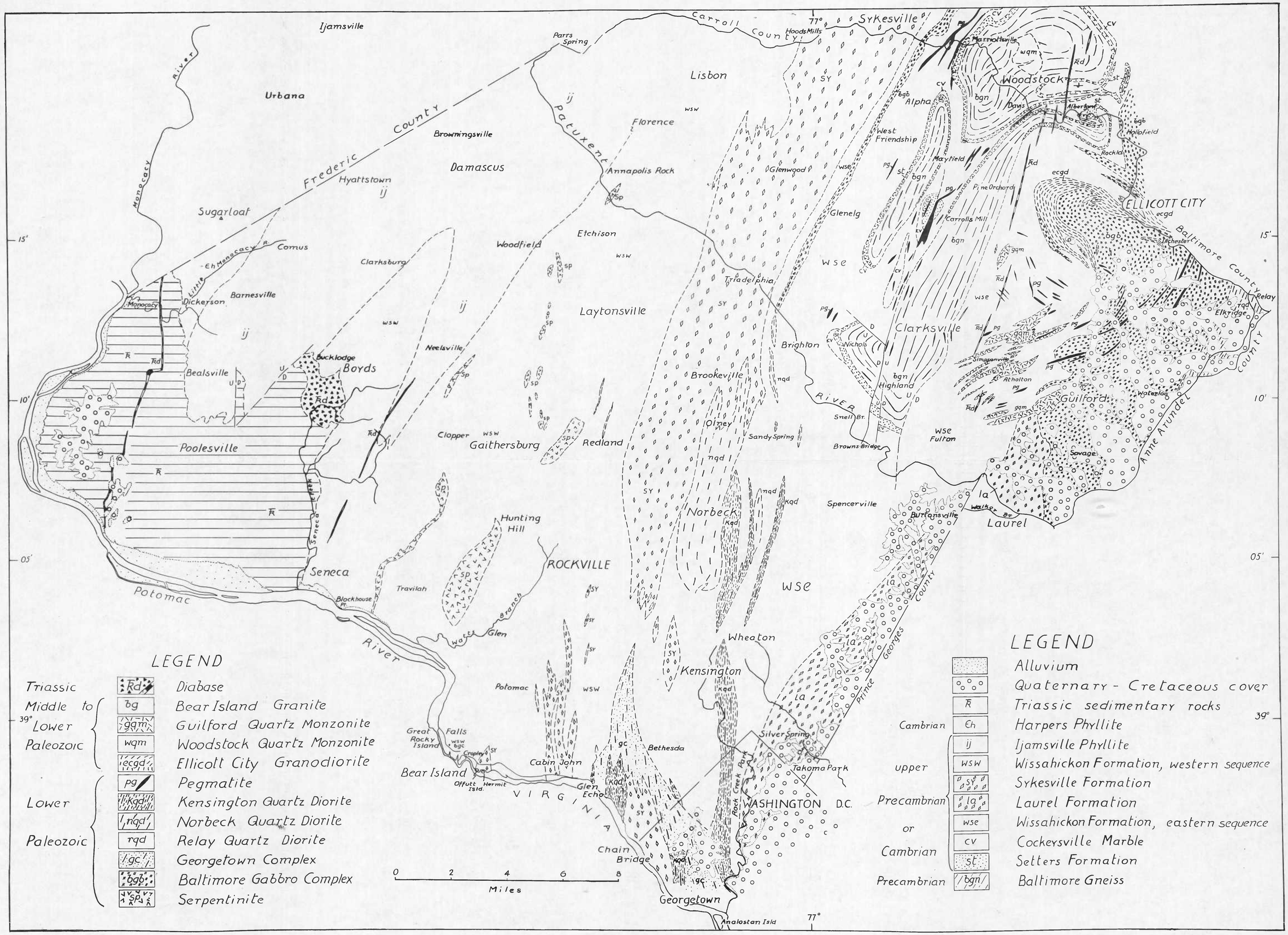


PLATE 45. Photomicrograph parallel Figure 1 of Plate 44. Strong lineation is shown as mineral orientation. The amphibolite is a lineated rock with almost no foliation.


LEGEND
Triassic
Rd Diabase
Middle to Lower Paleozoic
bg Bear Island Granite
gqm Guilford Quartz Monzonite
wqm Woodstock Quartz Monzonite
ecgd Ellicott City Granodiorite
Lower Paleozoic
pg Pegmatite
kqd Kensington Quartz Diorite
nqd Norbeck Quartz Diorite
rqd Relay Quartz Diorite
gc Georgetown Complex
bgb Baltimore Gabbro Complex
sp Serpentinite
LEGEND
Alluvium
Quaternary - Cretaceous cover
R Triassic sedimentary rocks
Cambrian
Єh Harpers Phyllite
upper Precambrian or Cambrian
ij Ijamsville Phyllite
wsw Wissahickon Formation, western sequence
sy Sykesville Formation
la Laurel Formation
wse Wissahickon Formation, eastern sequence
cv Cockeysville Marble
st Setters Formation
Precambrian
bgn Baltimore Gneiss
0 2 4 6 8
Miles
Ijamsville
Urbana
Frederic County
Monocacy River
Sugarloaf
Hyattstown
Damascus
Browningsville
Parrs Spring
Carroll County
Hoods Mills
Sykesville
Lisbon
Patuxent
Florence
Annapolis Rock
Woodstock
Alpha
West Friendship
Mayfield
Pine Orchard
Carrolls Mill
Glenwood
Glenelg
ELLICOTT CITY
Baltimore County
Relay
Elkridge
Comus
Little Monocacy R.
Clarksburg
Woodfield
Etchison
Triadelphia
Barnesville
Dickerson
Monocacy
Laytonsville
Clarksville
Brighton
Nichols
Highland
Simpsonville
Atholton
Guilford
Waterloo
Anne Arundel County
Bealsville
Buckledge
Boyds
Neelsville
Brookeville
Olney
Sandy Spring
Snell Br.
Browns Bridge
Fulton
Savage
Poolesville
Clopper
Gaithersburg
Redland
Norbeck
Spencerville
Burtonsville
Walker Br.
Laurel
Hunting Hill
ROCKVILLE
Seneca
Seneca Creek
Potomac River
Blockhouse Pt.
Travilah
Watts Branch
Glen
Wheaton
Kensington
Prince Georges County
Potomac
Great Falls
Rocky Island
Cropley
Bear Island
Offutt Isld.
Hermit
Cabin John
Glen Echo
Bethesda
Silver Spring
Takoma Park
Rock Creek Park
WASHINGTON D.C.
VIRGINIA
Chain Bridge
Georgetown
Analostan Isld
77°
15'
10'
05'
39°

P. 357 B

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