GEOLOGIC MAP OF THE LOUDOUN COUNTY, VIRGINIA, PART OF THE HARPERS FERRY QUADRANGLE

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INTRODUCTION

Bedrock and surficial deposits were mapped in 1989–90 as part of a cooperative agreement with the Loudoun County Department of Environmental Resources. This map is one of a series of geologic field investigations (Southworth, 1990; Jacobson and others, 1990) in Loudoun County, Va. This report, which includes geochemical and structural data, provides a framework for future derivative studies such as soil and ground-water analyses, which are important because of the increasing demand for water and because of potential contamination problems resulting from the recent change in land use from rural-agricultural to high-density suburban development.

Mapping was done on 5-ft-contour interval topographic maps (Loudoun Co., Va., unpub. maps, 1985) at a scale of 1:12,000, except in areas of complex structure on Blue Ridge where mapping was done at a scale of 1:6,000. The geology was then compiled on a U.S. Geological Survey base at 1:24,000 scale; some contacts, therefore, may not match topography as shown. Some linear features, such as fracture traces and faults, were interpreted from stereoscopic infrared aerial photographs.

The map covers the Loudoun County, Va., part of the Harpers Ferry 7.5′ quadrangle. The area is underlain by rocks of the northeast-plunging Blue Ridge–South Mountain anticlinorium, a late Paleozoic Alleghanian structure (fig. 1). Middle Proterozoic granitoids intruded by Late Proterozoic metabasalt and metarhyolite dikes and sills and Jurassic diabase dikes form the core of the anticlinorium and underlie the area's broad valleys. Outcrops of Middle to Late Proterozoic rocks are best exposed along Dutchman Creek and in the bluffs of the Potomac River. Late Proterozoic metasedimentary and metavolcanic rocks and Lower Cambrian metasedimentary rocks underlie the high ground of Blue Ridge and Short Hill Mountain where topographic relief ranges from 100 to 300 m (300 to 1,000 ft). Alluvium is found along Piney Run and along Dutchman Creek in the valley east of Short Hill Mountain. Colluvium is abundant in the mountainous areas. Numerous springs in the map area indicate that Blue Ridge and Short Hill Mountain are important recharge areas for ground water.

STRATIGRAPHIC NOMENCLATURE

Keith (1894) established the stratigraphy of the area, which contains the type localities of the Catoctin Schist, Loudoun Formation, Weverton Sandstone, and Harpers Shale. He considered the Catoctin to be basement and the Loudoun Formation to include all rocks beneath the Weverton. Jonas and Stose (1939) and Stose and Stose (1946) thought that the Catoctin Metabasalt intruded and overlaid the basement rocks; they subdivided the Loudoun into the Swift Run Tuff, Catoctin Metabasalt, and Loudoun Formation. Nickelsen (1956), who first mapped this area in detail, adopted this stratigraphy and considered the Swift Run and Catoctin Formations to be of Precambrian age and that the Loudoun Formation, Weverton Quartzite, Harpers Formation, and Antietam Quartzite constituted the Chilhowee Group of Early Cambrian age.

For this report, the Loudoun Formation is not recognized. As used by Nickelsen (1956), the Loudoun consisted primarily of phyllite and a local uppermost coarse pebble conglomerate. Phyllite, of probable tuffaceous origin, is largely indistinguishable in the Swift Run, Catoctin, and Loudoun Formations. Phyllite, previously mapped as Loudoun Formation, cannot be reliably separated from phyllite in the Catoctin; thus it is here mapped as Catoctin Formation. The coarse pebble conglomerate previously included in the Loudoun is here mapped as a basal unit of the overlying lower member of the Weverton Formation. The contact of the conglomerate and the phyllite is sharp and is interpreted to represent a major change in depositional environment. Further studies are needed to determine if use of the Loudoun Formation should be restricted.

GEOLOGIC SETTING

Rocks in this area that are interpreted to have fluvial, volcanic, or volcaniclastic (tuffaceous metasediments) origin are both heterogeneous and discontinuous along strike. Some rock units abruptly pinch and swell in thickness, and unconformities are common. For example, the lower member of the Weverton Formation is in contact with quartz syenite at Purcell Knob. Units that are absent here increase greatly in thickness to the north where unique rock types, such as sericitic phyllite and arkose of the Swift Run Formation, are abundant. Field studies have not determined whether this is the result of postdepositional erosion (Jonas and Stose, 1939) or deposition on a paleotopographic surface of moderate relief (Nickelsen, 1956). Unit thicknesses are estimates based on outcrop width of highly cleaved and often folded rock and are only approximate. Where possible, thicknesses were measured in areas of minimal deformation.

The Swift Run Formation lies unconformably on basement rocks and grades conformably up into tuffaceous metasedimentary rocks of the Catoctin Formation. In the Catoctin, tuffaceous metasedimentary rocks lie both above and below extrusive metabasalt and grade up into phyllite. A probable hiatus and period of erosion preceeded the onset of the Chilhowee Group deposition.

The rock types are interpreted to reflect changes in depositional environments related to the change from the rift-to-drift sedimentary regime. Clastic fluvial sediments of the Swift Run mark the beginning of rifting of Grenville basement to form the Iapetus Ocean (Rankin, 1976). Intrusive metabasalt and metarhyolite dikes and sills and extrusive metabasalt were emplaced during the main phase of continental rifting. Conglomerate of the lower member of the Weverton Formation is rift facies, and quartzites of the lower member of the Weverton Formation were deposited in a rift-to-drift transition. Conglomeratic quartzite of the upper member is another rift facies. The drift regime was established during deposition of the Harpers Formation.

MIDDLE PROTEROZOIC ROCKS

The oldest rocks underlying the study area are granitoids of probable intrusive origin. Four distinctive units are recognized: hornblende quartz monzonite, biotite granodiorite gneiss, garnet monzogranite, and quartz syenite. Most rocks, however, are monzogranite (fig. 2). Primary textures and minerals have been substantially altered by deformation and retrograde metamorphism.

Hornblende Quartz Monzonite

Hornblende-bearing quartz monzonite (Yhm) is very resistant to weathering and underlies the south half of Loudoun Valley and the Purcell Knob synformal anticline (ref. loc. 7). The contact between the hornblende quartz monzonite and the garnet monzogranite (Ym) is not exposed but trends northwestward. Hornblende crystals characterize the unit, which is generally massive and has a few zones of mylonitic foliation. Dark hornblende and opaque minerals define the foliation that is cut at a high angle by Paleozoic schistosity, which is characterized by chlorite, epidote, and sericite aggregates. Part of the unit is gneissic and contains perthite and oligoclase augen. Hornblende, biotite, pyroxene (hypersthene?), and perthite suggest hornblende-granulite facies metamorphism during the Middle Proterozoic Grenville orogeny. The unit is lithologically correlative with the Pedlar Formation of the Blue Ridge province. The Pedlar intrudes the 1,138-Ma Old Rag Granite (Gathright, 1976); therefore, if this correlation is correct, the quartz monzonite probably was emplaced during the Grenville event (Clarke, 1984).

Biotite Granodiorite Gneiss

Biotite granodiorite gneiss (Yg) is well exposed along the Potomac River (ref. loc. 5) where it is intruded by numerous dikes and sills of metabasalt (Zmd) as much as 50 m (164 ft) wide that locally constitute 50 percent of the exposure. The biotite granodiorite gneiss varies from potassium feldspar augen gneiss along the east boundary of the map area to mylonite in the western part of its occurrence. Its contact with the garnet monzogranite (Ym) is a thrust fault that marks the west edge of the 1-km- (0.6-mi-) wide Dutchman Creek shear zone. Southeast of Georges Mill, the contact is inferred because of limited exposure and little float. The rocks within the Dutchman Creek shear zone are highly foliated to mylonitic and contain massive pegmatite bodies as much as 15 m (49 ft) wide. Southeast of Lovettsville, the contact is inferred by limited exposures of sericite-rich mylonite. Roadcuts show that metabasalt dikes and sills are the predominant bedrock here. Foliation of presumed Grenvillian age is well exposed along the Potomac River (ref. loc. 5) and in the creek that parallels State Route 287 (ref. loc. 6) where the foliation is nearly flat; gently undulates to the north, east, and west; and is cut at a high angle by southeast-dipping metabasalt dikes and sills and Paleozoic schistosity. Foliation is seen as dark bands of biotite, some almandine, and opaque minerals. Light bands are quartz and feldspar and minor plagioclase augen. Paleozoic schistosity is represented by discrete ductile deformation zones of chlorite and sericite. Howard (in press) suggests that this unit may correlate with the Marshall Metagranite, located 50 km (31 mi) to the south, which has a $^{207}Pb/^{206}Pb$ age of 1,010 Ma (Clarke, 1984). The biotite granodiorite gneiss most resembles the granodiorite and biotite granite gneiss mapped north of the Potomac River by Stose and Stose (1946). Hypersthene present in those rocks was not observed here.

Garnet Monzogranite

Garnet monzogranite (Ym) is the most abundant basement unit. Garnet monzogranite is well exposed along the Potomac River from Potomac Wayside to just west of Dutchman Creek (ref. loc. 4). South of Elvan, the garnet monzogranite has much biotite and little garnet. Cataclastic mylonite and augen gneiss are locally present, but in general this unit is massive. The massive nature of this unit and the lack of foliation implies that it may be intrusive into the hornblende quartz monzonite (Yhm) and biotite granodiorite gneiss (Yg). Almandine, biotite, and sphene suggest amphibolite-facies metamorphism during the Middle Proterozoic Grenville orogeny. Altered plagioclase and garnet replaced by chlorite and sericite indicate retrograde metamorphism to greenschist facies during the Paleozoic. Blebs of chlorite define a lineation on the schistosity planes where the unit is most affected by greenschist metamorphism and deformation. At these places the rock has a greenish hue and greasy luster.

Light-brown (5YR 5/6) gneiss that has bands of almandine, clots of chlorite, and books of graphite is found near Dutchman Creek and resembles a paragneiss. Outcrops are sparse, so the gneiss is not mapped separately.

Quartz Syenite

Quartz syenite (Ys) is areally restricted to the east limit of the Purcell Knob antiformal syncline (ref. loc. 1) in Loudoun Valley and an area west of Mt. Olivet Church (ref. loc. 2). This unit is adjacent to hornblende quartz monzonite (Yhm) in Loudoun Valley, but no amphibole or pyroxene and only sparse biotite are present in the quartz syenite. The quartz syenite, which may be intrusive to both the hornblende quartz monzonite and the garnet monzogranite (Ym), is restricted to near the basement-to-cover contact. Where involved in Paleozoic deformation, the unit has mylonitic foliation. Mylonitic foliation is seen as fractured books of potassium feldspar megacrysts bounded by seams of chlorite and sericite. No hornblende, almandine, or biotite are present. Paleoregolith of the Swift Run Formation is derived from this unit.

LATE PROTEROZOIC ROCKS

Swift Run Formation

Rocks of the Swift Run Formation were called Loudoun Formation by Keith (1894) and later the Swift Run Tuff (Stose and Stose, 1946). Nickelsen (1956) mapped the Swift Run Formation as quartz sericite phyllite and blebby phyllite; the blebby phyllite was of probable tuffaceous origin.

The Swift Run Formation consists of a heterogeneous sequence of fluvial metasedimentary rocks that generally fine upward. The discontinuous, heterogeneous rocks of the Swift Run may be attributed to (1) original deposition on a paleotopographic erosion surface, (2) deposits that strike orthogonally to the present-day mountain trend, and (or) (3) differences in degree of deformation. It unconformably overlies basement rocks and transitionally grades up into the overlying tuffaceous rocks of the Catoctin Formation.

Five units are mapped in the Swift Run Formation: paleoregolith (Zsr), sericitic metasandstone (Zssm), quartz sericite schist (Zss), sericitic phyllite and arkose (Zsp), and marble (Zsm). Paleoregolith is recognized north of Purcell Knob (ref. loc. 3) and south of Mt. Olivet Church (Nickelsen, 1956) and represents a local basement-to-cover transition. This unit was derived from the quartz syenite (Ys).

Sericitic metasandstone, which forms the base of the Swift Run Formation, unconformably overlies the basement and is locally transitional with the paleoregolith. Nowhere is the basement contact exposed. Outcrops of sericitic metasandstone, which are as much as 10 m (33 ft) thick and contain cobbles and pebbles are best seen from north of Purcell Knob to Loudoun Heights (ref. loc. 8), and on the east side of Short Hill Mountain, southwest of Mt. Olivet Church (ref. loc. 9).

Quartz sericite schist grades laterally and up into the sericitic metasandstone. The quartz sericite schist has rounded quartz pebbles and coarse sand in a matrix of sericite. The schist often has a protomylonitic texture. Quartz sericite schist is interpreted to lie between discontinous outcrops of the sericitic metasandstone. The unit is best seen east of Short Hill Mountain at the Potomac River (ref. loc. 10) and in the Dutchman Creek (ref. loc. 11) and Mt. Olivet synclines (ref. loc. 12). Quartz sericite schist in the synclines is distinctively different in color and texture from outcrops at the base of Short Hill Mountain; these schistose metasedimentary rocks are believed to be an eastern facies of the unit. Quartz sericite schist and fine pebble conglomerate is also seen 2.5 km (1.6 mi) north of Purcell Knob (ref. loc. 13), where they dip under the hornblende quartz monzonite (Yhm) in the Purcell Knob synformal anticline.

Sericitic phyllite interbedded with thin arkose is present above the sericitic metasandstone at the Purcell Knob antiformal syncline (ref. loc. 14). Polyphase deformation has tectonically mixed these two units here, but some outcrops show the original stratigraphy. A small outcrop and float of similar composition is exposed in a sackung bedrock landslide at the end of Virginia Route 853 on the east side of Short Hill Mountain. Minerals observed in thin section are consistent throughout the sericitic phyllite and include quartz, sericite, feldspar, apatite, zircon, tourmaline, magnetite, hematite, leucoxene, and opaque minerals.

A lens of marble, less than 2 m (6 ft) thick and discontinuous along strike, marks the top of the Swift Run at the Potomac River east of Short Hill Mountain (ref. loc. 15). Marble float was found near the inverted contact of the sericitic metasandstone and the sericitic phyllite and arkose north of Purcell Knob. The marble may have been deposited in freshwater lakes rather than during a marine transgression. Volcaniclastic rocks and metabasalt of the Catoctin Formation are above the marble.

The Swift Run Formation (and the Catoctin) consists of lithofacies that may be time transgressive throughout the Late Proterozoic in this region. King (1950), Bloomer and Werner (1955), and Reed (1955) recognized that metabasalt was interbedded with the Swift Run Formation and in places overlapped onto basement; this same relation can be seen north of Purcell Knob.

Age constraints on the Swift Run Formation are poor. The Swift Run rests unconformably on the biotite granodiorite gneiss (Yg), which may correlate with the 1,010-Ma Marshall Metagranite; it underlies the Catoctin Formation, which has an age of 570±36 Ma (Badger and Sinha, 1988).

Metabasalt Dikes and Sills

Numerous metabasalt dikes and sills (Zmd), compositionally similar to metabasalt of the Catoctin Formation (Zcm), intrude the basement rocks; some also cut the Swift Run Formation. Metabasalt dikes and sills are so abundant that Keith (1894) and Cloos (1951) thought that the granite intruded the metabasalt. Keith (1894) stated that "the two were too closely and universally interbedded for separate representation". Espenshade's (1986) statement that 15 to 20 percent of the basement core of the Blue Ridge is intruded by metabasalt is considered a conservative estimate, on the basis of the number of dikes and sills present in basement rocks exposed along the Potomac River.

The metabasalt dikes have long been interpreted as feeder dikes for the extrusive metabasalt of the Catoctin Formation (Stose and Stose, 1946), and chemical data (table 1) support this idea. No metabasalt dikes were observed to feed or cut the metabasalt flows, and, in general, metabasalt dikes are sparse adjacent to these flows.

Metarhyolite Dikes

A metarhyolite dike (Zrd) as much as 50 m (164 ft) wide and over 8 km (5 mi) long intrudes the garnet monzogranite (Ym) and the Swift Run Formation. Tuffaceous metarhyolite containing feldspar phenocrysts, quartz-filled amygdules, and clasts of metarhyolite is found as float adjacent to the dike along the Potomac River (ref. loc. 16). The metarhyolite dikes are similar to felsic dikes (Stose and Stose, 1946; Gathright and Nystrom, 1974) and rhyolite of the Catoctin Formation (Fauth, 1981) examined in the region. The metarhyolite dikes may represent feeder dikes to the Catoctin flows.

Catoctin Formation

The Catoctin Formation has been called the Catoctin Schist (Keith, 1894), Catoctin Metabasalt (Stose and Stose, 1946), and Catoctin Greenstone (King, 1950). Nickelsen (1956) mapped four units in the Catoctin Formation: light-green phyllite, metabasalt, tuffaceous sedimentary rocks, and Catoctin dikes.

The Catoctin Formation in the map area includes metabasalt and tuffaceous metasedimentary rocks. In addition, a local, dark phyllite, previously assigned to the Lower Cambrian Loudoun Formation (Nickelsen, 1956) is here mapped as part of the Catoctin. This interpretation is consistent with that of Reed (1955), Edmundson and Nunan (1973), Gathright and Nystrom (1974), Rader and Biggs (1975), and Lukert and Nuckols (1976). The Catoctin is Late Proterozoic in age; Badger and Sinha (1989) cited a Rb-Sr whole-rock and pyroxene age at 570±36 Ma.

Three units are mapped in the Catoctin Formation: tuffaceous metasedimentary rocks (Zct), metabasalt (Zcm), and phyllite (Zcp). Tuffaceous metasedimentary rocks include phyllite, schist, mud-lump breccia, and thin-bedded metabasalt. The metabasalt unit includes massive metabasalt, agglomeratic metabasalt breccia that contains epidosite, and schistose metabasalt. The metabasalt unit is contained within and locally interlayered with the tuffaceous metasedimentary rocks unit; the interlayering suggests that the volcaniclastic sediments were incorporated into metabasalt flows (Reed, 1955).

Along the Potomac River at Short Hill Mountain, the Catoctin Formation is 84 m (276 ft) thick. The lowermost part of the Catoctin consists of tuffaceous phyllite and thin-bedded metabasalt (Zct). A 50-m (164-ft) section of the metabasalt unit (Zcm) consists of four belts of massive to schistose metabasalt, each ranging in thickness from 6 to 20 m (20 to 66 ft). The metabasalt has epidosite bodies as much as 0.5 to 1 m (1.5 to 3 ft) in diameter, which are bounded on either side by mud-lump breccia (Reed, 1955) and tuffaceous metasedimentary rocks (ref. loc. 19). The metabasalt unit, as well as the thin-bedded metabasalt within the tuffaceous metasedimentary rocks unit is interpreted as flows. Chemical analyses of the upper and lower flows (map Nos. 17 and

18, respectively) are shown in table 1 and figure 3. The metabasalt forms prominent outcrops in the Potomac River but is thin to absent north of the river for at least 7 km (4 mi).

The metabasalt unit is also observed on the Loudoun Heights trail. Here, as much as 61 m (200 ft) of amygdaloidal, massive, aphanitic to schistose, gray-green (5G 5/2) metabasalt is exposed. A chemical analysis of this rock (map No. 16) is shown in table 1 and figure 3. Deformation here is intense, but some flow structures and possible columnar joints are observed. Light-gray metarhyolite tuff is interbedded with the metabasalt.

In the upper part of the Catoctin, tuffaceous metasedimentary rocks increase in abundance upward and are transitional with the phyllite (Zcp); this relation is most clearly seen near the Loudoun Heights trail on Blue Ridge (ref. loc. 17). Here, approximately 10 m (33 ft) of phyllite containing very fine pebbles of quartz is found between the tuffaceous metasedimentary rocks and the basal conglomerate of the lower member of the Weverton (Ꞓwlc). No phyllite is recognized in this stratigraphic position on the east slope of Short Hill Mountain.

The largest exposure of the Catoctin Formation phyllite (Zcp) is north of Purcell Knob on the limbs of inverted folds (ref. loc. 18). The contact with the basal conglomerate of the lower member of the Weverton is abrupt. The phyllite is gray black (N2) and grades downward to light olive gray (5Y 6/1) and light gray (N7) and contains pale-red-purple (5RP 6/2) lithic clasts in a dusky-blue (5PB 3/2) matrix. Medium- to light-gray (N6) phyllite contains pinkish-gray (5YR 8/1), very light gray (N8), and pale-red-purple (5RP 6/2) amygdules elongated in the slaty cleavage plane, which indicates a volcanic origin. Some of the dark phyllite may be an iron-rich paleosol (Nickelsen, 1956). The phyllite is interbedded with thin layers of metabasalt at the Potomac River.

PALEOZOIC ROCKS

Chilhowee Group

In the map area the Chilhowee Group consists of the Weverton Formation and Harpers Formation. Nickelsen (1956) placed the Loudoun Formation at the base of the Lower Cambrian Chilhowee Group. The stratigraphic relation between the Chilhowee Group and the underlying Late Proterozoic Catoctin Formation has been interpreted as transitional (Nickelsen, 1956) or unconformable (King, 1950; Reed, 1955; Edmundson and Nunan, 1973; Gathright and Nystrom, 1974). Rounded clasts of red jasper and metabasalt (King, 1950; Reed, 1955; and Rader and Biggs, 1975) suggests a period of erosion between the two formations. Simpson and Sundberg (1987) recognized Early Cambrian fossils in the correlative Unicoi Formation in southwestern Virginia; the fossil beds are above a lens of metabasalt within the Unicoi. The Chilhowee is interpreted to be a marine transgressive sequence of a rift-to-drift transitional regime.

Weverton Formation

The Weverton Formation was called the Weverton Sandstone by Keith (1894) and Weverton Quartzite by Nickelsen (1956). The type locality (Keith, 1894) is "the prominent outcrops of South Mountain, near Weverton, Maryland." Nickelsen (1956) mapped lower, middle, and upper quartzite members.

In the map area, the Weverton Formation is divided into a lower member (Ꞓwl) and an upper member (Ꞓwu). The lower member consists of quartzite interbedded with metasiltstone and locally has a basal coarse pebble conglomerate (Ꞓwlc). The upper member consists of pebble conglomerate and metasiltstone. The base of the Weverton Formation may be unconformable with phyllite of the Catoctin Formation (Zcp). The upper member grades into the thin-bedded metasiltstone, arkose, and fine pebble conglomerate of the Harpers Formation.

The Weverton is interpreted as alluvial sediments (Schwab, 1986) at the base of a marine transgressive sequence and reflects a change from a volcaniclastic environment to a predominately fluvial one (Nickelsen, 1956). Current directions suggest a source from the west (Whitaker, 1955). Both the upper and lower members are fining-upward sequences. The upper member is interpreted to be a rift facies.

Lower Member and Basal Conglomerate

The lower member (Ꞓwl) contains a discontinuous basal, coarse, quartz-pebble conglomerate (Ꞓwlc that may represent channel fill cut into the Catoctin Formation. Rip-up clasts of phyllite and metabasalt of the Catoctin and rounded clasts of red jasper (King, 1950; Reed, 1955, Rader and Biggs, 1975) in the basal conglomerate suggest a period of erosion between the two formations. King (1950), Reed (1955), and Rader and Biggs (1975) (fig. 1) mapped similar basal conglomerates in the Weverton Formation.

The basal conglomerate is best exposed north of Purcell Knob (ref. loc. 20). No conglomerate is observed on the east side of Short Hill Mountain, but it is present on South Mountain, Md., approximately 20 km (12 mi) north of the map area. North of Purcell Knob, the basal conglomerate is exposed in vertical contact with the phyllite of the Catoctin Formation and quartzite of the lower member of the Weverton. Here, the quartz-pebble conglomerate contains rounded white and purple quartz clasts as large as 6 cm (2 in.) in diameter. The contact with the phyllite is sharp and the conglomerate contains rip-up clasts of phyllite. The matrix of the conglomerate is rich in hematite. The basal conglomerate is arkosic, some beds are coarse grained and others are cross stratified.

In places, the upper 20 cm (8 in.) of the phyllite of the Catoctin contains small matrix-supported pebbles that may suggest a transitional contact. In one exposure, a 1-m- (3-ft-) thick, light-gray, massive quartzite that contains blue-black cross-stratified beds is between the conglomerate and the phyllite. On the Loudoun Heights trail on the Blue Ridge (ref. loc. 17), coarse pebbles to small cobbles of rounded to subangular quartz and red jasper are found within the quartzite of the lower member. The quartzite is arkosic but grades into clean quartzite within 0.5 m (1.5 ft).

The majority of the lower member is well-sorted, mature quartzite that grades upward into quartzite interbedded with light-colored metasiltstone. The lower member is most clearly exposed at its type locality. Good exposures of the lower member can be seen north of Purcell Knob, at the Potomac River gorge at Blue Ridge, and along the crest of Short Hill Mountain as far north as the Potomac River (ref. loc. 22). At Purcell Knob, the lower member is arkosic and rests unconformably on quartz syenite (Ys). Stose and Stose (1946) recognized this relation and interpreted doming and erosion between the time of deposition of the Catoctin and Weverton Formations. The Swift Run and Catoctin Formations are thin to the north and south of Purcell Knob. Exposures of the Weverton Formation unconformable on basement rocks were also described by King (1950) and Bloomer and Werner (1955). At the Potomac River gorge at Blue Ridge, the lower member is highly folded and the stratigraphy is obliterated by dynamic recrystallization.

The contact with the upper member is a rarely exposed dark metasiltstone. The dark metasiltstone is best exposed at the type locality section and immediately north of the Radio Tower on Short Hill Mountain (ref. loc. 21). The dark metasiltstone is similar in appearance to metasiltstones in the lower part of the Harpers Formation (King, 1950); this may have led Cloos (1951), Nickelsen (1956), and Burford and others (1964) to interpret folds and thrust faults within the Weverton Formation.

Upper Member

The upper member (Ꞓwu) is a diagnostic "gun-metal blue" (Nickelsen, 1956) to green-gray pebble conglomerate. Blue quartz, magnetite, and opaque minerals make the pebble conglomerate distinctively dark blue. Only sparse outcrops of the upper member are exposed on the west slope of the Blue Ridge. Exposures on the Blue Ridge at the Potomac River are so deformed by folding and attendant cleavage that Chimney Rock (local usage) was interpreted by Woodward (1949) to be the Harpers Formation. The upper member crops out well on Short Hill Mountain where the Hillsboro syncline preserves up-right and overturned sections, whose tops are being shown by crossbeds and graded beds (ref. loc. 23). The upper member is more poorly sorted than the lower member and contains pebbles of blue and red quartz, opaque minerals, and blue-green phyllite clasts 0.5 to 15 cm (0.2 to 6 in.) long. As much as 5 m (16 ft) of metasiltstone at the base of the upper member grades up into 4 m (13 ft) of clean gray-green conglomeratic quartzite. Over 8 m (26 ft) of gun-metal blue pebble conglomerate grades upward into 20 m (64 ft) of green-gray pebble conglomerate at the top. A total section of 32 m (105 ft) (ref. loc. 21), is in accordance with the 23 to 39 m (75 to 128 ft) measured by Nickelsen (1956).

Harpers Formation

King (1950) renamed the Harpers Shale of Keith (1894) the Harpers Formation because siltstone and sandstone are the dominant rock types. The type locality of the Harpers Formation is the "fine exposures on the Potomac at Harpers Ferry" (Keith, 1894).

In the map area, the Harpers Formation (Ꞓh) is a monotonous sequence of phyllitic metasiltstone interbedded with meta-arkose and pebble conglomerate at the base and meta-sandstone in the upper part of the formation. These units were not differentiated because of lack of exposure. *Skolithos* tubes (trace fossil) were seen in the Harpers Formation on South Mountain 10 km (4 mi) north of the map area, but none were recognized in the map area. *Skolithos* tubes support the interpretation that the Harpers Formation represents a fluvial to deltaic marine environment of the transgressive Chilhowee Group.

Deformation of the Harpers Formation is intense: outcrops are dominated by slaty cleavage and bedding rarely can be observed. The relation of bedding to cleavage is seen along the Shenandoah River near the type locality and in the Hillsboro syncline on Short Hill Mountain. Deformation is less intense west of the U.S. Route 340 bridge at Bolivar, W.Va. Here bedforms are seen in light-gray (N7) sandstone and metasiltstone. The lower part of the Harpers Formation is best seen in Coon Hollow (local usage), 1 km (0.6 mi) south of the Potomac River on the west side of Short Hill Mountain (ref. loc. 24). Metasiltstone of the Harpers Formation is phyllonitic in the immediate upper plate of the Short Hill thrust fault. A traverse up the inverted section shows a wide range of metasiltstone, calcareous metasiltstone, sandy metasiltstone, and magnetite-rich arkose. At the north end of Short Hill Mountain along the Potomac River, the lower part of the Harpers is a quartz-rich metasiltstone containing thin-bedded arkosic metasandstone that grade down into the upper member of the Weverton. On the north side of the Potomac River, there is a 1-m- (3-ft-) thick, dark greenish-gray (5GY 4/1) fine pebble conglomerate.

Jurassic Diabase Dikes

Near-vertical, northwest-trending diabase dikes of Jurassic age intrude the Blue Ridge–South Mountain anticlinorium. The diabase is fine- to coarse-grained, massive, dense, and composed of plagioclase, augite, olivine, and (or) quartz. Porphyritic and granophyric diabase is seen along the Potomac River (ref. locs. 25 and 26). Diabase dikes are most numerous east of Short Hill Mountain. Several linear to gently curving near-vertical dikes as much as 61 m (200 ft) wide are near Georges Mill, (ref. loc. 25). The dikes are discontinuous and are en echelon. An $^{40}Ar/^{39}Ar$ radiometric age of 200 Ma (M.J. Kunk, U.S. Geological Survey, oral commun., 1990) was obtained from the 4.5-m- (15-ft-) wide diabase dike on the Baltimore & Ohio Railroad cut between Weverton and Sandy Hook, Md. (the northwest extension of ref. loc. 26).

Diabase dikes are barriers to ground-water movement. They can be resistant ridges (Georges Mill) or topographic swales. Commonly the dikes are expressed as rounded boulders in a sandy clay soil. Diabase is easily distinguished from metabasalt because it lacks cleavage and weathers to spheroidal boulders.

METAMORPHISM

The Middle Proterozoic rocks contain mineral assemblages indicative of upper amphibolite to granulite-facies metamorphism. Hornblende, pyroxene, almandine, biotite, and sphene suggest amphibolite- to hornblende-granulite-facies metamorphism during the Grenville orogeny. During the Paleozoic these rocks were metamorphosed into greenschist facies along with their cover rocks. This event sausseritized the plagioclase and chloritized the amphibole, pyroxene, garnet, and biotite. Bodies of epidosite within metabasalt of the Catoctin Formation (Zcm) also formed during the Paleozoic event (Reed and Morgan, 1971). It is uncertain whether the rocks containing garnet and biotite were once in granulite facies and later retrograded, as suggested by Bailey (1990).

Mitra and Elliott (1980) suggested that deformation and greenschist-facies metamorphism were coeval. Elliott (1973) found that the anticlinorium formed under greenschist-facies conditions at about 350 °C and 3.5 kbar pressure corresponding to an overburden of 10.6 km (35,000 ft). Late Proterozoic to Cambrian age rocks have mineral assemblages stable in greenschist facies. Albite, chlorite, epidote, and magnetite octahedra are found in all cover rocks; in addition, actinolite is present in metabasalt and sericite is abundant in aluminosilicate-rich rocks. Epidosite is most common in metabasalt flows but is also found in metabasalt dikes and sills and basement rocks. Epidosite is believed to be formed by circulating fluids after deposition (Espenshade, 1986). Morgan (1972) interpreted the Cambrian and Ordovician sequence immediately west of Harpers Ferry, W.Va., to be subgreenschist facies, but Gathright and Nystrom (1974) showed that rocks of the Lower Cambrian Chilhowee Group contain mineral assemblages stable in greenschist facies.

White metamorphic quartz veins are common in all units. The Purcell Knob antiformal syncline exposes sericitic phyllite of the Swift Run Formation that contains isoclinally folded quartz veins.

The fold axial planes parallel a pervasive slaty cleavage, and a later crenulation deforms the fold limbs into upright folds. The White Rock thrust fault is parallel to bedding in the Weverton Formation and is marked by cataclastic white metamorphic quartz, which constitutes as much as 80 percent of the exposure. Abundant residual blocks and boulders of white metamorphic quartz are common in areas underlain by basement rocks.

STRUCTURAL GEOLOGY

The Harpers Ferry quadrangle covers the west half of the Blue Ridge–South Mountain anticlinorium; the anticlinorium is overturned to the northwest, plunges gently to the northeast, and is allochthonous above one or more blind thrust faults (Mitra and Elliott, 1980; Mitra, 1987). It is a classic fault-bend fold. Structural geology in this area has been the subject of research in fold and thrust fault mechanics by Cloos (1947, 1971), Elliott (1970, 1973), and several generations of graduate students at Johns Hopkins University (Nickelsen, 1956; Gautum Mitra, 1978; Shankar Mitra, 1978, 1979). Structural elements include foliation and schistosity in Middle Proterozoic rocks, lineations, at least three orders of folds, bedding, slaty and crenulation cleavage, shear zones, both normal and thrust faults, and joints.

Blue Ridge and Short Hill Mountain are both overturned west limbs of major folds of the Blue Ridge–South Mountain anticlinorium that are separated by the Short Hill thrust fault. The Short Hill thrust fault is here interpreted to be a low-angle thrust fault that transported the South Mountain fold northwestward onto the core of the Blue Ridge fold. Lower hemisphere, equal-area projections of structural data (fig. 4) are presented for the terranes both east and west of the Short Hill thrust fault, because they may have had different tectonic histories. The terrane east of the fault is here called Short Hill–South Mountain terrane and the area west of the fault is here called Blue Ridge terrane. Maxima of hand-contoured poles to bedding, slaty cleavage, and foliation and schistosity demonstrate that the basement and cover rocks were deformed together to produce the Blue Ridge–South Mountain anticlinorium (fig. 4).

Foliation and Schistosity

Foliation of basement rocks is expressed as the planar arrangement of dark- and light-colored minerals. The dark minerals include biotite, almandine, hornblende, pyroxene, chlorite, and opaque minerals. The light-colored minerals are quartz, plagioclase, and alkali feldspar. The foliation is presumed to be of Grenvillian age. Schistosity of Paleozoic deformation cuts the foliation. Schistosity is seen as sericite and chlorite in thin ductile-deformation zones. Foliation is most clearly seen in the biotite granodiorite gneiss (Yg) and in gneissic units of the hornblende quartz monzonite (Yhm). Foliation is less apparent in the garnet monzogranite (Ym) and quartz syenite (Ys). Foliation is gentle to steep and often dips west in gneissic units of the hornblende quartz monzonite (Yhm); foliation in biotite granodiorite gneiss (Yg) dips gently to the north, east, and west (ref. loc. 6). Southeast-dipping schistosity cuts the foliation and is coplanar to cleavage in the cover rocks. Poles of foliation compared to schistosity (fig. 4*D*) show that the intersection angle is steeper in the Blue Ridge terrane than in the Short Hill–South Mountain terrane.

Lineations

Garnet, chlorite, and biotite elongated down the schistosity plane of basement rocks were mapped as lineations. The lineations are clearly observed in pavement outcrops of hornblende quartz monzonite (Yhm) in Loudoun Valley. The lineations consistently plunge west to northwest (fig. 4*F*). Slaty cleavage is marked with down-dip lineations of aligned quartz grains, metamorphic minerals, stretched lithic fragments of phyllite, vesicles, and amygdules. The intersections of bedding with slaty and spaced cleavage were not systematically recorded, but they roughly parallel the axis of the Blue Ridge–South Mountain anticlinorium (Nickelsen, 1956).

Mylonitic Fabric

The deformed basement rocks commonly display textures (Higgins, 1971) ranging from protomylonite to ultramylonite; some blastomylonite is present. All are mapped as mylonite. All Middle Proterozoic rocks contain mylonite, but hornblende quartz monzonite (Yhm) contains the least. Garnet monzogranite (Yg) is generally quite massive, but mylonite zones are present. Most mylonite zones are no wider than several meters and most are less than 20 cm (8 in.) wide. Poles to mylonitic foliation (fig. 4*F*) suggest that it is associated with schistosity in the basement rocks in the Short Hill–South Mountain terrane. Mylonitic foliation in the Blue Ridge terrane is gently dipping and appears to be associated with east- and west-dipping schistosity that defines folds (fig. 4*F*). Poles to mylonitic foliation (fig. 4*F*) are coplanar with poles to slaty cleavage of Late Proterozoic dikes and sills (fig 4*E*).

The largest mylonite zone is the 1-km- (0.6-mi-) wide Dutchman Creek shear zone (ref. loc. 33). Along the Potomac River, more than seven discrete mylonite zones, ranging from 0.5 m to 3 m (1.5 to 10 ft) wide, define the shear zone. The rocks are predominantly ultramylonite and blastomylonite. Sense-of-shear indicators of asymmetric porphyroclastic augen, fractured books of feldspar, pressure shadows of quartz, sericite fish, and S-C fabrics, consistantly show east over west. The shear zone contains abundant pegmatites, some of which are mylonitized. The biotite granodiorite gneiss (Yg) grades into augen gneiss that, in turn, grades into mylonite as the shear zone is approached. The basement rocks are poorly exposed south of the Potomac River, and it is difficult to assess the regional importance of the mylonite zones.

The garnet monzogranite (Ym) and biotite granodiorite gneiss (Yg) have zones of quartz sericite schist. Keith (1894), Stose and Stose (1946), and Nickelsen (1956) recognized these zones of intense shearing and all stated that they closely resembled the phyllitic units of the Swift Run Formation. In general, the quartz sericite schist of the Swift Run Formation (Zss) has a protomylonite texture because of ductile deformation near its contact with the basement rocks. The quartz sericite schist of the Swift Run Formation and adjacent protomylonite of basement rock result from ductile deformation in the basement-to-cover transition zone. This relation can best be seen at the margins of the Mt. Olivet syncline and along the base of the east slope of Short Hill Mountain.

Dikes and Sills

Late Proterozoic Dikes and Sills

Many Late Proterozoic metabasalt dikes and sills (Zmd) intruded the Middle Proterozoic basement. One metarhyolite dike (Zrd) and some metabasalt sills intruded both Middle Proterozoic rocks and the Swift Run Formation. None of these dikes and sills can be seen to feed the metabasalt and metarhyolite flows of the Catoctin Formation (Zcm); however, geochemical data (table 1; fig. 3) support this relation. The dikes and sills were generated during the rifting leading to the continental breakup and the opening of the Iapetus Ocean (Rankin, 1976). The distribution of dikes and sills, therefore, indicates the zone of rifting.

Metabasalt dikes cut the Grenville foliation at a high angle but parallel Paleozoic schistosity and mylonitic foliation. Slaty cleavage of the rocks is parallel to the contacts and dips southeastward coplanar to the so-called South Mountain cleavage. At many places the slaty cleavage gives the rock an anastamosing fabric. Poles to slaty cleavage show that the maximum for Short Hill–South Mountain terrane dikes and sills is N. 26° E., 22° SE. The maximum for Blue Ridge dikes and sills is N. 32° E., 19° SE. (fig. 4*E*). Dikes and sills that dip away from the southeast are those associated with shear zones and the Purcell Knob inverted folds (fig. 5). Changes in the strike and dip of dikes and sills are interpreted to reflect broad folds of Paleozoic deformation of the basement. Mylonite records the orientation of greatest strain in the basement rocks; therefore, poles to mylonitic foliation (fig. 4*F*) compare to contoured poles to Late Proterozoic dikes and sills (fig. 4*E*).

Some dikes and sills contain xenoliths of basement rocks. Dikes and sills show boudinage and pinch-and-swell structure near contact zones with basement rocks. The metabasalt is phyllonitic near the contact and intrafolial folds are in the plane of slaty cleavage. Basement rocks obviously utilized the dikes and sills as planes of simple shear during deformation of the core.

Two distinct trends of metabasalt dikes (N. 55° E. and N. 15° E.) are seen in Piney Run near Loudoun Heights (ref. loc. 32). Dikes trending N. 15° E. cut garnet monzogranite schistosity of N. 55° E. N. 55° E.-trending dikes cut schistosity of N. 15° E.

Jurassic Dikes

Northwest-trending Jurassic diabase dikes were emplaced during the extensional event that led to the opening of the Atlantic Ocean. These near-vertical dikes are discontinuous and are en echelon, probably because they were emplaced in subparallel fractures. Diabase dikes are most numerous east of Short Hill Mountain. Their north-northwest trend suggests extension to the north-northeast. One dike trends somewhat more westerly than the other dikes. It cuts the Short Hill fault but ends approximately 10 m southeast of the fault. A smaller dike mapped on float and magnetometer surveys, is located about 152 m (499 ft) to the south. There is no proof that the dikes are the same. Either the dike has been offset by minor normal movement on the upper plate rocks east of the Short Hill thrust fault, or the dikes were emplaced en enchelon along parallel fractures. Magnetometer traverses 7 km (4 mi) north of the Potomac River show that there a diabase dike cuts the fault and is not displaced.

The Jurassic diabase dikes are more than 14 km (9 mi) west of the Bull Run Mountain border fault of the Culpeper Basin of Mesozoic age; they demonstrate that the Blue Ridge–South Mountain anticlinorium experienced the extensional tectonics leading to the opening of the Atlantic Ocean.

Folds

A series of complex refolded folds in hornblende quartz monzonite (Yhm) was mapped north of Purcell Knob in Loudoun Valley. The Purcell Knob antiformal syncline and synformal anticline, as well as several inverted parasitic folds, are well constrained by 1:6,000-scale mapping of stratigraphy of the cover rocks for this study. North of Purcell Knob, younger rocks that have slaty cleavage dip westward beneath older rocks. Slaty cleavage dips west and is cut by crenulation cleavage that dips east. The axes of the synformal anticline and antiformal syncline rise and fall along strike. The culmination of the antiformal syncline exposes metabasalt of the Catoctin Formation (table 1, map No. 15) beneath the Swift Run. The rocks are thickened in fold hinges and thinned on inverted limbs, which is consistent with supratenuous folds of passive flow. Early metamorphic quartz veins and compositional layering are isoclinally folded so that they parallel the west-dipping slaty cleavage; fold axes plunge either southwest or northeast and are refolded by structures that have an axial surface parallel to the southeast-dipping crenulation cleavage. Deformation of the hornblende quartz monzonite (Yhm) in the synformal anticline is heterogeneous. Mylonitic foliation is seen at the contact of the inverted limb, but the basement rock is massive in the interior of the erosional window. Foliation and schistosity in hornblende quartz monzonite in Loudoun Valley dip both northwest and southeast and may indicate tight folds in the basement rocks.

Folds in the cover rocks range in size from the first-order Blue Ridge–South Mountain anticlinorium to microscopic interfolial folds. The Blue Ridge and Short Hill–South Mountain are overturned west limbs of anticlines that are separated by the Short Hill fault. The Blue Ridge defines a tectonite front (Mitra, 1987). Deformation is much more extreme on the Blue Ridge than on Short Hill–South Mountain. Mesoscopic and macroscopic folds are overturned to the northwest, are disharmonic, and are of similar fold geometry because they have thickened crests and thinned limbs. Macroscopic folds that have wavelengths of 1 to 10 m (3 to 33 ft) plunge gently northeast or southwest. Slickenlines, stretched quartz veins, and boudinage plunge down the dip of the bedding and indicate flexural slip folding.

Short Hill–South Mountain is underlain by the second-order Hillsboro syncline. It is overturned to the northwest and is cored by rocks of the Harpers Formation. Stereographic determination using poles to bedding (fig. 4*A*) suggests that the Hillsboro syncline probably plunges about 11° to 16° toward N. 19° E. to N. 36° E. In places, the overturned east limb shows terrace steps of a monocline. Local third- and fourth-order folds were mapped near the Radio Tower on Short Hill Mountain. A third- or fourth-order anticline in this area has a wavelength of several meters plunges gently south.

The Weverton Formation is intensely folded on Blue Ridge. Second- and third-order, asymmetrical, similar folds have wavelengths of 1 to 100 m (3 to 328 ft) and plunge gently southwest. Two recumbent folds, exposed in the cliffs on the Maryland side of the Potomac River gorge, plunge 10° SW. Parasitic folds on these larger structures are clearly seen on Blue Ridge (ref. loc. 34). The rocks are extremely thickened by folding. In the Harpers Ferry National Historical Park, several anticlines, synclines, and antiformal synclines in the Harpers Formation are recognized by cleavage fans. Anticlinal fold axes are shown on contoured stereonets of poles to bedding (fig. 4*A*).

Two synclines in quartz sericite schist of the Swift Run Formation (Zss) are located east of Short Hill–South Mountain. The synclines are short and probably shallow. The Dutchman Creek syncline is in the upper plate of a shear zone that separates two distinctive Middle Proterozoic units. The Mt. Olivet syncline has a complex outcrop pattern due to a basement-to-cover transition zone and possible refolded folds.

Bedding Orientation

Bedding can be recognized in most rocks above the Middle Proterozoic basement. Graded beds and crossbeds are most visible in the Weverton Formation and are present at many places in the Swift Run Formation. Bedding is less obvious in the Harpers Formation and is difficult to find in phyllite and schist of the Swift Run and Catoctin Formations. Possible igneous flow structures can be seen locally in the Catoctin Formation. Bedding can only consistently be traced along strike in the Weverton Formation on Short Hill Mountain. Elsewhere, bedding changes attitude over short distances, which suggests that significant folding has occurred.

Poles to bedding of the Short Hill–South Mountain terrane have a maximum of N. 16° E., 35° SE. (fig. 4*A*). More than 88 percent of the 164 observations are from the Weverton Formation, and crossbeds show that bedding is overturned in 93 percent of the observations. Sparse west-dipping data are upright fold limbs. Strike variations due to broad bending of folds is seen in the spread of the 1, 2, and 4 percent contours (fig. 4*A*). Poles to bedding in rocks of the Blue Ridge terrane have a maximum of N. 39° E., 42° SE. (fig. 4*A*). More than 77 percent of the 162 observations were measured on the Weverton Formation and 72 percent of these were overturned.

Cleavage

All Late Proterozoic and Paleozoic rocks in this area have a penetrative slaty cleavage that dips consistently to the southeast. This South Mountain cleavage (Cloos, 1947; 1951) is axial planar to the Blue Ridge–South Mountain anticlinorium. Slaty cleavage development depended on rock competence. Crenulation cleavage is present in the phyllites and schists and is probably related to late folding.

Slaty Cleavage

All slaty cleavage dips southeast and is axial planar to the Blue Ridge–South Mountain anticlinorium except in the Purcell Knob inverted folds. Slaty cleavage (S_1) is best developed in fine-grained phyllite and schist of the Swift Run, Catoctin, and Harpers Formations, where it is the dominant structural element because bedding has been largely obliterated. Schistosity of the sericitic schist of the Swift Run Formation is mapped as slaty cleavage. Slaty cleavage is less developed in sericitic metasandstone of the Swift Run and quartzite and conglomerate of the Weverton Formation. Slaty cleavage surfaces have a shiny luster produced by the synkinematic growth of sericite and chlorite flakes. Quartz grains are elongated and flattened in the cleavage.

Poles to slaty cleavage of Short Hill–South Mountain have a maximum of N. 22° E., 26° SE. (fig. 4*B*). Poles to slaty cleavage of Blue Ridge have a maximum of N. 30° E., 19° SE. (fig. 4*B*). Some poles to crenulation cleavage are included in this plot.

In the Purcell Knob inverted folds, poles to slaty cleavage have a maximum of N. 16° E., 30° NW. (fig. 5). The northwest-dipping slaty cleavage is cut at a high angle by southeast-dipping crenulation cleavage. In finer grained rocks, bedding is transposed into slaty cleavage where the rock is blebby and variegated on the slaty cleavage plane. Poles to slaty and crenulation cleavage in dikes and sills and schistosity in hornblende quartz monzonite in the inverted folds indicate that "the basement participated in the deformation but less willingly" (Cloos, 1951).

Crenulation cleavage

A late crenulation cleavage that folds slaty cleavage is most clearly developed in fine-grained phyllite and schist. The crenulation cleavage (S_2) is 1-mm- to 1-cm- (.04- to .4-in.-) wide spacings that intersect the slaty cleavage at high angles. At the Purcell Knob inverted folds, the crenulation cleavage uniformly dips to the southeast (fig. 5) and approximately parallels the regional slaty cleavage of the rest of the quadrangle (fig. 4*B*). Poles to S_2 show a maximum of N. 16° E., 30° SE. (fig. 5). Where crenulation cleavage is well developed, the slaty cleavage shows intrafolial folds and transposed compositional layering in the slaty cleavage plane. Crenulation cleavage is best developed along Blue Ridge. Crenulation cleavage in the Swift Run and Catoctin Formations on Short Hill Mountain may represent a late stage of minor folding. Crenulation cleavage on Blue Ridge consistently dips to the southeast, and on Short Hill Mountain it dips both southeast and northwest.

Faults

Faults in the map area consist of the Short Hill thrust fault, the White Rock thrust fault, the Dutchman Creek shear zone, and several small normal faults.

The Short Hill fault was interpreted by Cloos (1951) and Nickelsen (1956) to be a Triassic normal fault that down-dropped the Short Hill–South Mountain terrane relative to the crest of the Blue Ridge. Nickelsen (1956) recognized that the fault truncates folds in the Weverton and Harpers Formations, and interpreted these rocks to be in contact with granodiorite gneiss on a normal fault dipping 50° to 60° E. Reconnaissance mapping (C.S. Southworth, unpub. mapping, 1990) shows that the fault is very complicated over a length of 50 km (31 mi). Excavations along U.S. Route 340 in Maryland and cuttings from water wells suggest that the fault may dip as little as 10° SE. Metasiltstone of the Harpers Formation above the fault is protomylonite, which has S-C fabrics that suggest east-over-west movement. Thus, the Short Hill fault is here interpreted to be a low-angle thrust fault that places the overturned west limb of the South Mountain fold onto the core of the overturned Blue Ridge anticline. The fault cuts up-section to the north and parallels fold axial surfaces in the upper plate.

The White Rock thrust fault is interpreted to be a detachment in quartzite of the lower member of the Weverton Formation on the crest of Short Hill Mountain. The thrust fault is marked by a zone of quartz breccia as much as 10 m (33 ft) thick. In places, slaty cleaved quartz constitutes 80 percent of the fault zone in which 0.25- to 0.5-m- (0.7- to 1.5-ft-) thick slabs of quartzite float in the white quartz matrix (ref. loc. 35). The trace of the fault defines a bow where there is maximum displacement. Here, the dip flattens and the fault cuts out the upper member of the Weverton. South of White Rock, the east slope of Short Hill Mountain has scattered blocks of deformed quartz. This area is interpreted to be the sole of an eroded thrust sheet. A small normal fault was mapped in this area by Nickelsen (1956). The stratigraphy in the area is not duplicated; therefore, the structure is interpreted to be a sackung bedrock landslide that displaced rocks of the sole thrust downslope to the east.

As previously discussed, the Dutchman Creek shear zone is a 1-km- (0.6-mi-) wide fault zone that separates two distinctive basement units. Garnet is restricted to rocks west of the fault, whereas biotite predominates to the east. Sericitic quartz schist of the Swift Run Formation (Zss) in the Dutchman Creek syncline is in the upper plate of the shear zone; thus it may represent an eastern facies of the Swift Run Formation that was transported westward.

Northwest-trending, steep, northeast-dipping shear zones in garnet monzogranite along Potomac Wayside at the bluffs of Loudoun Valley are interpreted to be normal faults. The fault rock has a phyllonitic to pseudotachylitic texture. Cloos (1950) recognized that one of the dikes was vertical and crosscut all foliations (ref. loc. 36). The geochemistry of this pseudotachylite fault rock suggests that it is a Jurassic diabase dike (table 1, map No. 21; fig. 3) sheared in a down-to-the-southwest sense.

Near the Virginia-West Virginia State line west of Potomac Wayside, northwest-striking slickensides in the Weverton Formation are marked with slickenlines that dip 35° to 40° SE. They are about parallel to the normal faults and may result from strike-slip faulting along the Potomac River water gap. Shear zones that have the same orientation are observed north of the Potomac River in basement rocks.

Joints

Both open and annealed, quartz-filled joints are common in all rocks of the quadrangle. Joints are most abundant in quartzite of the Weverton Formation and can be interpreted to be cross, longitudinal, and oblique joints formed during the folding of these rocks. Poles to joints in the Weverton Formation on Short Hill–South Mountain have a maximum of N. 73° W. that is nearly perpendicular to the strike of the rocks where the data were collected (fig. 4*G*). The joints are spaced at a minimum of 1 cm apart where folds are deflected or where the rocks are adjacent to bedding-plane thrust faults. Joints are less common in phyllite and schist. Joints are common in the Middle Proterozoic rocks. Poles to joints in garnet monzogranite (Ym) of the Blue Ridge plot in a maxima of N. 72° W. and N. 49° E.; the N. 72° W. set is parallel to the inferred cross joints in the cover rocks (fig. 4*G*). Poles to joints in biotite granodiorite gneiss (Yg) east of Short Hill–South Mountain plot in a maximum striking about due north (fig. 4*G*). The set dips moderately west almost perpendicular to east-dipping cleavage.

Fractures

A total of 522 linear fracture traces were interpreted from stereoscopic color-infrared aerial photographs and 5-ft-contour contour topographic maps. Trends that had two or more fracture traces were plotted as poles, assuming a near-vertical inclination (fig. 4*H*). Poles to traces of fracture planes of Short Hill–South Mountain show maxima of N.–S. and N. 48° E.; the highest frequency of fractures is 21. They trend N.–S. The N.–S.-fracture set trends parallel to the maximum of joints in biotite granodiorite gneiss. Poles to traces of fracture planes in the Blue Ridge have a maximum of N. 40° W. The highest frequency is 13. They trend N. 5° E. The N. 5° E.-trending fracture planes parallel a maximum of joints in garnet monzogranite, whereas the N. 40° W. set parallels the cross joints in the Weverton Formation on Short Hill Mountain. Maxima of poles to joints in garnet monzogranite (fig. 4*G*) and poles to fracture planes in the Blue Ridge (fig. 4*H*) are parallel with Jurassic dikes and shear zones related to Mesozoic extension. Major fracture traces on Short Hill–South Mountain are parallel to Jurassic dikes, but they do not define the pole maximum. Major fracture traces show a 1:1 correlation with poles to joints of the same area. The correlation of joints to fracture traces in the basement rocks is good for site specific areas. An attempt to relate regional fracture traces to limited joint data would probably give random results.

BEDROCK LANDSLIDES

Several large bedrock landslides (Schultz and Southworth, 1989) are recognized and include rock block slump and sackung. The bedrock landslides were caused by both gravity and ground water and are probably prehistoric in age. A rock block slump of the Catoctin Formation is at the end of Virginia Route 852 near the Potomac River. Approximately 25,000 m^3 of tuffaceous phyllite and schistose metabasalt had rotational movement 20 m (66 ft) downslope. The rock block slump exposes west-dipping slaty cleaved rock at the toe; detachment was along the slope parallel to the schistosity in metabasalt at the head. Hemlocks trees grow on the rock block slump and suggest that it has been stable for over a century.

Gravitational sagging, or sackung (Zischinsky, 1969) was recognized along the east slope of Short Hill Mountain east of White Rock and Buzzard Rock. Sackung is continuous shear deformation distributed among numerous fractures that have no basal sliding surface (Varnes, 1978). Sackung are common in the western Cordillera (Varnes and others, 1989) but were only recently recognized in inclined strata in the Appalachian Valley and Ridge province (Southworth and Schultz, 1988; Schultz and Southworth, 1989).

A 0.5-km- (0.3-mi-) long by 100-m- (328-ft-) wide sackung of quartzite of the lower member of the Weverton Formation is exposed west of the intersection of Virginia Routes 758 and 852 (ref. loc. 30). The center of the sackung has been breached to form a gap. The gap exposes 10 to 13 m (33 to 43 ft) of northwest-dipping right-side-up quartzite on top of southeast-dipping overturned quartzite. Southeast-dipping overturned quartzite forms the dip slope west of the gap and a continuous dip slope near the contact with the Catoctin Formation. The lip of the sackung can be traced continuously from end to end; the strike of west-dipping right-side-up beds forms an arc above a linear northeast-striking belt of overturned quartzite on the dip slope. One quartzite block 3 m (10 ft) thick and 20 m (66 ft) long is flat lying in the sag north of the gap and is interpreted to be a gravitational klippe.

A spring in the gap discharges over 200,000 gal/d for the Brunswick, Md., municipal water supply (Southworth, 1990). A Jurassic diabase dike may also cut the gap; magnetometer profiles provide insufficient data to connect the dike segments as shown. A causal relation between ground-water impediment by the diabase dike and increased pore pressure along bedding and joint planes may have facilitated sackungen.

A small normal fault interpreted by Nickelsen (1956) southeast of White Rock (ref. loc. 31) is reinterpreted here as a sackung that developed along the sole of a bedding-parallel thrust fault. Quartzite of the lower member of the Weverton Formation is displaced eastward over a continuous strike belt of the Swift Run and Catoctin Formations that forms the hogback ridge. Bulldozer excavation and cuttings from a 61-m- (200-ft-) deep water well (John Rickard, oral commun., 1989) showed quartzite saprolite over tuffaceous phyllite of the Catoctin Formation. This area is locally known as the "sand pit," and the water well discharges over 75 gal/min (Brutus Cooper, oral commun., 1990). The entire east slope is colluvium of quartzite and metamorphic quartz and sandy saprolite, both derived from the sole of a bedding-parallel thrust fault within the Weverton Formation. Bedrock is covered along the front of the sackung, but west-dipping slaty cleavage in the Swift Run Formation to the north supports the interpretation that gravitational sagging of bedrock is common along the base of the dip slope. The presence of sackung near the Potomac River suggests that hydrologic gradients increase toward base level and contribute to the gravitational sags.

SURFICIAL GEOLOGY

The surficial geology of the map area was mapped by Jacobson and others (1990) at a scale of 1:12,000. Jacobson and others (1990) recognized four units on the basis of depositional processes and material properties and dimensions.

For this report, two surficial units are mapped: alluvium and fine colluvial debris, undifferentiated, (Qal) and colluvium (Qc). The alluvium and fine colluvial debris unit is a mixture of clay, sand, pebbles, gravel, cobbles, and some boulders that underlie sinuous flood plains along creeks and rivers. Alluvial terraces as much as 6 m (20 ft) above base level are included. Fine colluvial debris from side slopes is transitional with alluvium. Alluvium is well sorted and is found in fining-upward sequences as much as 6 m (20 ft) thick. High alluvial terraces of the Shenandoah and Potomac Rivers are not recognized along the cutbank in Virginia but were previously mapped in Virginia (Gathright and Nystrom, 1974) and Maryland (Stose and Stose, 1946). Coarse cobbles, boulders, and blocks of quartzite and some metabasalt and granitoid rocks, which were transported by colluvial processes, are found as debris fans and debris terraces along the Potomac River at Blue Ridge and Short Hill Mountain. The tops of these debris deposits are planar, which suggests modification by fluvial reworking. A spectacular example of this process can be seen near the Short Hill thrust fault on the Potomac River (ref. loc. 29).

Coarse colluvium (Qc), primarily quartzite of the Weverton Formation, is abundant on Short Hill Mountain and Blue Ridge. The downslope extent of this material is shown on the map as a heavy dotted line.

Topographic relief and surface-water runoff play an important role in depositional processes and resultant deposits and landforms. Depositional processes include gravity and freeze-thaw (colluvium), fluvial transport (alluvium), and debris flow (debris); debris flow involves both gravity and fluvial transport and, thus, is a transitional process. Colluvium and debris grade laterally and downslope from coarse to fine; conversely, alluvium grades upward from coarse to fine. These units are spatially distributed from upper to lower slopes, and each has a characteristic landform. Coarse colluvium includes boulder streams (ref. loc. 26), boulder fields (ref. loc. 27), and talus. Colluvium is abundant on all mountain slopes, but coarse colluvium is concentrated in hillslope depressions and hollows. Coarse colluvium of quartzite on west-facing outcrop slopes is blocky due to mechanical weathering along longitudinal and cross joints; on east-facing dip slopes, the colluvium is slablike along cross joints and bedding planes. Lateral migration of drainage adjacent to block streams and debris flows produces a W-shaped hollow as a result of gully gravure (Mills, 1981). Gully gravure is most common on outcrop slopes where quartzite colluvium lies in hollows incised in residuum of the Harpers Formation; this is best seen near the Potomac River, west of the Short Hill thrust fault (ref. loc. 28). Fine colluvium, debris, and alluvium is a continuum of matrix-supported diamicton to matrix-supported clasts. These units are transitional at the base of the slope. Particle dimensions decrease, clasts become rounded, and the amount of matrix increases as a function of both slope and surface-water runoff.

Ridgecrests and dip slopes have outcrop and (or) shallow bedrock and are thus thin colluvium over residuum, which is generally 0 to 2 m (0 to 7 ft) thick (Jacobson and others, 1990), and is characterized by broad convex slopes.

GROUND WATER

Ground-water behavior is a function of bedrock lithology, structural fabric, and the textures of overlying surficial deposits. Ground-water recharge occurs high on the slope where inclined strata of quartzite parallel the slope. Massive quartzite beds have low porosity and permeability; thus, the dip slope is characterized by rapid overland flow. Interflow increases in areas of thin colluvium over residium and becomes greatest where convex slopes are incised by hollows near the base of the slope. Coarse colluvium and debris is concentrated in the hillslope hollows where interflow is highest. Contact springs on mountain slopes define broad areas of ground-water discharge (Southworth, 1990). Springs discharge at the heads of many hollows where near-vertical cross joints in quartzite provide conduits to the ground water. Coarse colluvium in hollows along the lower slope serve as catchment basins for surface and shallow ground water. Thus, hollows and linear stream reaches are controlled by cross joints through a combination of spring sapping and the mechanical weathering along joint and (or) bedding planes.

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