Prepared in cooperation with the National Park Service

Geologic Map of the Shenandoah National Park Region, Virginia

By Scott Southworth, John N. Aleinikoff, Christopher M. Bailey, William C. Burton, E.A. Crider, Paul C. Hackley, Joseph P. Smoot, and Richard P. Tollo

Open-File Report 2009–1153

U.S. Department of the Interior
U.S. Geological Survey

U.S. Department of the Interior
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U.S. Geological Survey
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U.S. Geological Survey, Reston, Virginia 2009
Revised and reprinted: 2009

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Suggested citation:
Southworth, Scott, Aleinikoff, J.N., Bailey, C.M., Burton, W.C., Crider, E.A., Hackley, P.C., Smoot, J.P., and Tollo, R.P., 2009, Geologic map of the Shenandoah National Park region Virginia: U.S. Geological Survey Open-File Report 2009–1153, 96 p., 1 plate, scale 1:100,000.

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Conversion Factors

Multiply	By	To obtain
	Length	
centimeter (cm)	0.3937	inch (in.)
millimeter (mm)	0.03937	inch (in.)
meter (m)	3.281	foot (ft)
kilometer (km)	0.6214	mile (mi)
	Area	
square kilometer (km^2)	0.3861	square mile (mi^2)

Temperature in degrees Fahrenheit (°F) may be converted to degrees Celsius (°C) as follows:
°C=(°F-32)/1.8
Vertical coordinate information is referenced to the North American Vertical Datum of 1988 (NAVD 88)
Horizontal coordinate information is referenced to North American Datum of 1983 (NAD 83)
Altitude, as used in this report, refers to distance above the vertical datum.

Geologic Map of the Shenandoah National Park Region, Virginia

By Scott Southworth,[1] John N. Aleinikoff,[2] Christopher M. Bailey,[3] William C. Burton,[1] E.A. Crider,[1] Paul C. Hackley,[1] Joseph P. Smoot,[1] and Richard P. Tollo[4]

Introduction

The geology of the Shenandoah National Park region of Virginia was studied from 1995 to 2008. The focus of the study was the park and surrounding areas to provide the National Park Service with modern geologic data for resource management. Additional geologic data of the adjacent areas are included to provide regional context. The geologic map can be used to support activities such as ecosystem delineation, land-use planning, soil mapping, groundwater availability and quality studies, aggregate resources assessment, and engineering and environmental studies.

The 1:100,000-scale map was compiled from field data collected at 1:24,000 scale. Geologic mapping was part of a cooperative investigation between the U.S. Geological Survey (USGS), and faculty and students of the College of William and Mary and The George Washington University. Much of the universities' work was through the USGS National Cooperative Geologic Mapping Program's Educational Mapping (EDMAP) component. The index map (fig. 1) shows the sources of geologic data for the 1:100,000-scale geologic map. For each map unit shown on the List of Map Units on plate 1, there is a corresponding expanded description found in the Description of Map Units following this explanatory text.

The map covers two distinct geologic provinces within the central Appalachian region that are defined by unique bedrock and resulting landforms. They are the Blue Ridge province on the east and the Great Valley section of the Valley and Ridge province on the west. The geology is discussed by geologic province and the geologic units within each province are discussed from oldest to youngest.

Previous Studies

Numerous important investigations in the Shenandoah National Park and surrounding area resulted in several geologic maps. King (1950) studied the Elkton area for manganese resources to support military needs during World War II. Reed (1955) studied the Catoctin Formation east of Stanley. County maps and reports included Rockingham (Brent, 1960), Albemarle (Nelson, 1962), Greene and Madison (Allen, 1963), and Page (Allen, 1967). Gathright (1976) used these and more recent data (Rader and Biggs, 1975; Gathright and others, 1977, 1978a,b,c) to produce the first geologic map of the Shenandoah National Park in the form of 1:62,500-scale maps for the northern, central, and southern sections. Historically, studies of the surficial geology in the region were incidental to

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investigations of the bedrock, but specific surficial investigations were conducted by Hack (1965), Kochel and Johnson (1984), Sherwood and others (1987), Kochel (1987, 1990, 1992), Kite (1992), Morgan and others (1999a,b), Whittecar and Duffy (2000), Litwin and others (2001, 2003), Eaton and others (2003a,b), Smoot (2004), Morgan and others (2004), and Wieczorek and others (2004). There are road logs for the geology along the Skyline Drive (Gathright, 1976; Badger, 1999) and numerous field guides to the region (Mitra and Lukert, 1982; Conley, 1989; Reed, 1989; Tollo, Aleinikoff, Borduas, and Hackley, 2004; Bailey and others, 2006).

General Geologic Setting and Physiography

The study area is centered on the Shenandoah National Park, which is mostly situated in the western part of the Blue Ridge province. The map covers the central section and western limb of the Blue Ridge-South Mountain anticlinorium. The Skyline Drive and Appalachian National Scenic Trail straddle the drainage divide of the Blue Ridge highlands. Water drains northwestward to the South Fork of the Shenandoah River and southeastward to the James and Rappahannock Rivers. East of the park, the Blue Ridge is an area of low relief similar to the physiography of the Piedmont province (Fenneman, 1938). The Great Valley section of the Valley and Ridge province is west of Blue Ridge and consists of Page Valley and Massanutten Mountain. The distribution and types of surficial deposits and landforms closely correspond to the different physiographic provinces and their respective bedrock.

The Shenandoah National Park is underlain by three general groups of rock units: (1) Mesoproterozoic granitic gneisses and granitoids, (2) Neoproterozoic metasedimentary rocks of the Swift Run Formation and metabasalt of the Catoctin Formation, and (3) siliciclastic rocks of the Lower Cambrian Chilhowee Group. The gneisses and granitoids mostly underlie the lowlands east of Blue Ridge but also rugged peaks like Old Rag Mountain (996 m). Metabasalt underlies much of the highlands, like Stony Man (1,200 m). The siliciclastic rocks underlie linear ridges from 800 to 400 m in altitude. The Page Valley is underlain by Cambrian and Ordovician carbonate rocks. Siliciclastic rocks are mostly west of the South Fork of the Shenandoah River and underlie Massanutten Mountain. Surficial deposits in the highlands include colluvium and debris fans. The lowlands have broad alluvial fans, alluvial plains, and fluvial terraces. Ridges underlain by siliciclastic rocks have abundant boulder fields. Numerous sinkholes and caves are due to the dissolution of the carbonate bedrock.

General Geologic History

Mesoproterozoic granitic gneisses and granitoids were emplaced, metamorphosed, and deformed during several phases of the late Mesoproterozoic Grenvillian orogeny (1,183 to 1,028 million years, or mega-annum (Ma); Tollo and others, 2006). These rocks were uplifted and eroded, and early Neoproterozoic sedimentary rocks were locally deposited (post-960 Ma), buried, and metamorphosed (Southworth and others, 2008). The rocks were intruded by granitoids of the Neoproterozoic (730–700 Ma) Robertson River Igneous Suite (Tollo and Aleinikoff, 1996). Local volcanic rocks associated with the Robertson River Igneous Suite accumulated and were intercalated with sediment (Tollo and Hutson, 1996; Bailey, Peters, and others, 2007). Unconformably overlying the gneisses, granitoids, paragneisses, and igneous rocks are Neoproterozoic sedimentary rocks of the Mechum River and Swift Run Formations (Bailey, 2007). They were succeeded by volcanic rocks of the Catoctin Formation (about 570 Ma), which were mostly basalt flows fed by a swarm of diabase dikes (Lukert and Nuckols, 1976) during continental extension (Rankin, 1976). Siliciclastic sedimentary rocks of the newly formed continental margin were deposited upon an irregular topographic surface that had developed on the Mesoproterozoic and Neoproterozoic rocks. As the rifted continental margin stabilized and developed into a passive margin during the early Paleozoic, carbonate rocks were deposited on a broad continental shelf. The early Paleozoic carbonate platform became unstable in response to the outboard Ordovician Taconian orogeny. Deposition changed from carbonate rocks to the siliciclastic rocks (Ordovician to Devonian) that underlie the Massanutten synclinorium and Massanutten Mountain. Greenschist-facies metamorphism and deformation associated with the Carboniferous to Permian Alleghanian orogeny resulted in folded and cleaved strata cut by numerous thrust faults. The cross sections on plate 1 illustrate the regional structure of the Blue Ridge anticlinorium and Massanutten synclinorium. All of the rocks were transported westward along the North Mountain thrust fault, which crops out on the west side of the Great Valley. About 80 million years (m.y.) after the Alleghanian orogeny, early Mesozoic continental rifting resulted in the formation of the Culpeper basin to the east of the map area. Basaltic dikes were intruded at about 200 and from 157 to 155 Ma, during the opening of the Atlantic Ocean. Sediment derived from erosion of the rocks of the Blue Ridge highlands was deposited into the adjacent lowlands.

Methodology

The focus of geologic mapping and research was within the Shenandoah National Park and the Blue Ridge east of the park. The geology of the Great Valley was compiled from Rader and Gathright (2001a,b). The research priorities within the park were (1) the Mesoproterozoic rocks, (2) the relations between basement and cover rocks, (3) delineating faults, (4) stratigraphy of the Chilhowee Group, and (5) surficial geologic units. The Mesoproterozoic rocks and some of the Neoproterozoic rocks were classified based on whole-rock chemistry (Tollo, Aleinikoff, Borduas, and Hackley, 2004; Tollo and others, 2006). Age determinations of the Mesoproterozoic rocks were determined using the sensitive high-resolution ion microprobe (SHRIMP) for U-Pb isotopic analyses of zircons (Tollo, Aleinikoff, Borduas, and Hackley, 2004; Tollo, Bailey, Borduas, and Aleinikoff, 2004; Tollo and others, 2006; Southworth and others, 2008). East of the park, the extents of the Robertson River Igneous Suite (Tollo and Lowe, 1994) and Mechum River Formation (Bailey, Peters, and others, 2007) were compiled from existing 1:100,000-scale maps. Geologic maps at 1:24,000 scale published between 1975 and 1980 exist for only about 30 percent of the map area. New 1:24,000-scale mapping was conducted in about 14 quadrangles through the Educational Mapping (EDMAP) component of the USGS National Cooperative Geologic Mapping Program (see fig. 1). Approximately seven published 7.5-minute quadrangle maps by the Virginia Division of Geology and Mineral Resources were modified and incorporated into the final map. Rock units were mapped on 1:24,000-scale topographic base maps. Elsewhere, field mapping was directed to solve known specific problems and to obtain reconnaissance data in unmapped areas. Structural data acquired from this study were combined with data compiled from all available sources and are presented on the map.

Surficial units of Gathright and others (1977, 1978a,b,c), Rader and Conley (1995), Eaton and others (2001), and Morgan and others (2004) were revised by new mapping. Surficial deposits across the region were mapped along selected field traverses. Conceptual models generated from the traverses allowed for the interpretation of landforms and surficial units elsewhere. Surficial deposits have distinctive geomorphic characteristics with mappable contacts, and consist of material that is distinct from the surrounding and underlying units. Some surficial units and landforms were interpreted by their expression on digital elevation models (DEMs), topographic maps, and aerial photographs. Exposures of the surficial units included those provided by landslides and stream-cut banks as well as roadcuts and excavations. Sinkholes and mines in the Great Valley were compiled from Stose and others (1919), King (1950), and Gathright and others (1977, 1978a,b,c). The surficial units were compiled, inked on mylar registered to the latitude and longitude coordinates of 7.5-minute quadrangles, scanned, and plotted at a scale of 1:50,000. A slope map derived from a 10-meter (m)-resolution DEM (Chirico and Tanner, 2004) was used to refine some of the contacts. Rock outcrops that are too small to portray on the map are denoted by structural symbols within surficial deposits, such as alluvium or debris fans.

The base map is a mosaic of the 1:100,000-scale Front Royal and Charlottesville 30′×60′ quadrangles with 20-m contour intervals. Because the metric contours and map elements were manually derived from the 1;24,000-scale 7.5-minute topographic base maps with 40-foot (ft) contour intervals, some of the unit polygons may not correspond directly with the metric base map.

Blue Ridge Province

Mesoproterozoic Rocks

Introduction

The Mesoproterozoic rocks in the core of the Blue Ridge anticlinorium are some of the most lithologically and structurally complex rocks in eastern North America. They are predominantly orthogneisses and metamorphosed granitoids that are included in three groups of rock that correspond to magmatic intervals as determined by SHRIMP U-Pb zircon ages (Tollo, Aleinikoff, Borduas, and Hackley, 2004; Tollo, Bailey, Borduas, and Aleinikoff, 2004; Aleinikoff and others, 2005; Tollo and others, 2005, 2006; Tollo and Kentner, 2007). Some outcrops show intrusive crosscutting relations. There are several noteworthy outcrops of granitoids that contain xenoliths of gneisses (Tollo, Aleinikoff, Borduas, and Hackley, 2004; Tollo, Bailey, Borduas, and Aleinikoff, 2004; Bailey and others, 2006). Collectively, the Mesoproterozoic rocks constitute a large melting pot of magma. The rocks crystallized, were metamorphosed and deformed, and further melted and recrystallized, over an extended period of time.

Previous Work

Jonas (1928, 1934) first designated Mesoproterozoic rocks in Virginia as the Lovingston Granite Gneiss, Marshall Granite, and hypersthene granodiorite for one of the first geologic maps of Virginia. Because all other rocks were deposited on top of them, they have been referred to as the “basement rocks.” They have also been called the injection complex (Jonas and Stose, 1938; King, 1950), the basement complex (Bloomer and Werner, 1955), the Blue Ridge Basement Complex (Rader and Evans, 1993), Grenville massifs (Bartholomew and Lewis, 1984), and Grenville terrane (Sinha and Bartholomew, 1984) (the latter two because the rocks were the products of the late Mesoproterozoic Grenvillian orogeny). In the Shenandoah National Park area, Furcron (1934) named and described the Old Rag Granite. Nelson (1962) described the Lovingston Gneiss, Marshall Granite, Crozet Granite, and the Virginia Blue Ridge Complex; his Virginia Blue Ridge Complex included granodiorite and hypersthene granodiorite which he considered to be the equivalent of the Pedlar Formation (Bloomer and Werner, 1955), whose type locality is southwest of the map area. Allen (1963) described the Lovingston Formation, Old Rag Granite, Marshall Formation, and Pedlar Formation. Rocks of the Pedlar Formation in the park area were described as hypersthene granodiorite (Allen, 1963), hypersthene-bearing granulites (Reed, 1969), and charnockites (Herz and Force, 1984). Gathright (1976) suggested that the Old Rag Granite and Pedlar Formation were gradational to one another. The Pedlar Formation and Virginia Blue Ridge Complex, as previously used, contain a variety of rocks, and modern U-Pb zircon geochronology demonstrates that their ages span a period of 154 m.y. (Aleinikoff and others, 2000; Tollo, Aleinikoff, Borduas, and Hackley, 2004; Tollo, Bailey, Borduas, and Aleinikoff, 2004; Tollo and others, 2006). The Blue Ridge Basement Complex on the Geologic Map of

Virginia (Virginia Division of Mineral Resources, 1993) consists of an assortment of rock units that included the Old Rag Granite, Crozet Granite, Flint Hill Gneiss, and a host of rocks that were considered to be the Pedlar Formation (leucocharnockite, porphyritic leucocharnockite, charnockite, charnockite gneiss, megacrystic charnockite, leucocratic granulite and gneiss, layered pyroxene granulite, layered porphyroblastic pyroxene granulite, and layered biotite granulite and gneiss).

The Mesoproterozoic rocks are difficult to distinguish from one another in the field; thus they require chemical, isotopic, and thin-section analyses to differentiate them (Tollo, Aleinikoff, Borduas, and Hackley, 2004; Tollo, Bailey, Borduas, and Aleinikoff, 2004; Tollo and others, 2006). In the early 1950s, some of the first isotopic dating techniques were applied to rocks in the Shenandoah National Park. The rocks exposed south of Thornton Gap at the south portal of the tunnel near Marys Rock were initially determined to be approximately 1,300 Ma (Tilton and others, 1960). New and improved isotopic techniques and analyses indicate that these rocks are 1,159±14 Ma. The Old Rag Granite was determined to be 1,115 Ma by Lukert (1982), but our analyses establish that it is 1,060±5 Ma.

Terminology

All Mesoproterozoic rocks were metamorphosed at upper amphibolite to granulite facies during the Mesoproterozoic and contain variably developed foliations. The map-unit name and lithologic nomenclature include the diagnostic minerals (biotite and orthopyroxene), texture (megacrystic, lineated, gneiss) and petrologic normative compositions (from Le Maitre and others, 1989) as determined by chemical plots (Tollo, Aleinikoff, Borduas, and Hackley, 2004; Tollo, Bailey, Borduas, and Aleinikoff, 2004; Tollo and others, 2006). All of the rocks are igneous in origin on the basis of primary mineral assemblages, chemical compositions, igneous textures, crosscutting relations, and xenoliths (Tollo, Aleinikoff, Borduas, and Hackley, 2004; Tollo, Bailey, Borduas, and Aleinikoff, 2004; Tollo and others, 2006). They are all granitoids, with percentages of quartz, alkali feldspar, and plagioclase used to determine the International Union of Geological Sciences' (IUGS) nomenclature. The descriptor "orthopyroxene" is used instead of the term "charnockite," which has both strict and more general connotations (Le Maitre and others, 1989). The term "gneiss" is restricted to rocks that contain compositional layering and a well-developed foliation. Leucocratic granites (or leucogranites) are light-colored, quartz-rich granites. Megacrystic and porphyroblastic granites are very coarse grained rocks.

The names and symbols of some units have been modified from those used by Tollo and others (2006). For example, several units that were previously called "gneisses" are now considered to be "metagranitoids" and, where chemical data are available, IUGS nomenclature is used instead of the default term "metagranitoid." These rocks have been examined over a period of 80 years by many geologists with varying degrees of sophistication, which inevitably has resulted in conflicts with names. A few formally named units have been retained, including (1) Old Rag Granite (Furcron, 1934; Gathright, 1976; Lukert, 1982), (2) Crozet Granite (Nelson, 1962), and (3) Flint Hill Gneiss (Lukert and others, 1977).

Lithologies

The oldest rocks in the Shenandoah National Park region are strongly foliated gneisses. The youngest rocks are weakly foliated to nonfoliated metagranitoids that include leucocratic, orthopyroxene-bearing, and biotite-bearing varieties. Five leucogneisses are interlayered at map scale with various orthopyroxene-bearing nonleucocratic gneisses. The 19 rock units consist of 4 general lithologies: (1) coarse-grained metagranitoid, (2) leucocratic gneiss, (3) orthopyroxene-bearing nonleucocratic metagranitoid, and (4) biotitic metagranitoid. Most of the units have been described previously by Tollo and others, (2004b,c; 2006), although five units are newly or more thoroughly described here; they are granodiorite gneiss (Ygd), orthopyroxene-bearing gneiss of monzogranite-quartz monzodiorite composition (Yomg), lineated leucogranite gneiss (Yll), the megacrystic Crozet Granite (Ycg), and biotite monzogranite-quartz monzodiorite (Ybg).

Groups

Introduction

SHRIMP crystallization ages for Mesoproterozoic rocks in the park range from 1,183±11 to 1,028±9 Ma. The tripartite subdivision of Mesoproterozoic rocks in the northern Blue Ridge (Aleinikoff and others, 2000; Tollo, Aleinikoff, Borduas, and Hackley, 2004; Tollo, Bailey, Borduas, and Aleinikoff, 2004; Tollo and others, 2006) is herein revised. Group 1 consists of strongly foliated orthogneisses and metagranitoids that crystallized from 1,183 to 1,144 Ma. Group 2 consists of variably foliated metagranitoids that crystallized from 1,143 to 1,111 Ma. Group 3 consists of 13 mostly massive metagranitoids that crystallized from 1,078 to 1,028 Ma. Within the limitations of the 2-σ uncertainties, the group boundaries are transitional, suggesting that there was a near continuum of magmatism interspersed with episodes of deformation. A time gap of 8 m.y. remains between 1,095 and 1,087 Ma where no igneous activity has yet been documented.

Group 1

Leucogranite Gneiss

Three rock types are included within the leucogranite gneiss unit (Ylg), and they are some of the oldest dated rocks in the region: two leucogranites (1,183±11 and 1,171±22 Ma) and leucogranite dikes (1,175±11 Ma) (Tollo, Aleinikoff, Borduas, and Hackley, 2004; Tollo, Bailey, Borduas, and Aleinikoff, 2004; Tollo and others, 2006) (fig. 2*A*). The reference locality of the unit (table 1, sample 36) is in the chute of a debris flow in Deep Hollow, west of Virginia State Route 651, about 2.5 km southwest of Aylor in the Madison 7.5-minute quadrangle (Bailey and others, 2003; Tollo, Aleinikoff, Borduas, and Hackley, 2004; Tollo, Bailey, Borduas, and Aleinikoff, 2004; Bailey and others, 2006). The leucogranite gneiss and dikes constitute a composite pluton. A large xenolith of leucogranite (1,171±22 Ma) was sampled along the Covington River (table 1, sample 35; Washington 7.5-minute quadrangle; Tollo and others, 2006).

The composite gneissic felsic pluton in Deep Hollow is composed of megacrystic leucocratic granite; medium-grained, equigranular leucogranite; and coarse-grained leucogranite pegmatite dikes that have important crosscutting relations (Bailey and others, 2003; Tollo, Aleinikoff, Borduas, and Hackley, 2004; Tollo, Bailey, Borduas, and Aleinikoff, 2004; Tollo and others, 2006). The megacrystic leucogranite is composed of alkali-feldspar-dominant mesoperthite (containing both microcline and orthoclase), quartz, and intergrown biotite. The medium-grained leucogranite is composed of alkali feldspar (chiefly microcline), plagioclase, and quartz, with minor secondary biotite. The leucogranite pegmatite is composed of alkali feldspar (microcline), minor plagioclase, and quartz.

The megacrystic leucogranite was intruded by leucogranite dikes. Pegmatite derived through differentiation of the magma (dated at about 1,175 Ma) forms dikes (which exhibit boudinage) cutting both rock types. A 30- to 50-cm-wide dike of fine- to medium-grained biotite granodiorite intruded all of the leucogranites; the dike is mineralogically similar to biotite monzogranite-quartz monzodiorite (Ybg), which was dated at 1,040±9 and 1,028±9 Ma. The intrusive relations and structures indicate multiple episodes of magmatic injection and deformation over an extended interval of time.

Foliated, Garnetiferous, Porphyroblastic Monzogranite

Foliated, garnetiferous, porphyroblastic monzogranite (Ypg; 1,172±8 Ma) occurs in the eastern part of the map area, east of the Robertson River Igneous Suite (Zrr). The reference locality is immediately east of the Flint Hill 7.5-minute quadrangle boundary in the Jeffersonton 7.5-minute quadrangle (Aleinikoff and others, 2000; Davis and others, 2002). This unit has diagnostic ovoid porphyroblasts of orange to pink microcline that are 1 to 3 cm in diameter in a matrix of finer grained plagioclase with clots of blue quartz, biotite, and garnet (fig. 2*B*). Foliation is defined by flattened porphyroblasts.

Orthopyroxene Granodiorite Gneiss

Orthopyroxene granodiorite gneiss (Yod; 1,165±7 Ma) is restricted to the north end of the Shenandoah National Park. The reference locality is a roadcut along Skyline Drive on the south side of Carson Mountain between Lands Run Gap and Compton Gap (table 1, sample 33; Chester Gap 7.5-minute quadrangle) (fig. 2*C*). The gneiss is composed of 35 to 40 percent alkali feldspar (microcline or orthoclase), 30 to 40 percent plagioclase, and 20 percent quartz, 3 to 5 percent amphibole, and 3 to 5 percent orthopyroxene (Tollo, Aleinikoff, Borduas, and Hackley, 2004; Tollo, Bailey, Borduas, and Aleinikoff, 2004; Tollo and others, 2006). The compositional layers and strong foliation are defined by alternating centimeter-scale, leucocratic layers rich in quartz and feldspar and dark layers rich in pyroxene and amphibole. Some of the leucosomes are transposed intrafolial folds. Interlayered and crosscutting dikes of light-gray to white leucogranite are likely Old Rag Granite.

Orthopyroxene Syenogranite and Monzogranite Gneiss

Orthopyroxene syenogranite and monzogranite gneiss (Yon, 1,164±8 Ma) is a strongly foliated, compositionally layered rock containing enclaves of garnetiferous quartz-feldspar leucosome. The reference locality is Skinners Ridge, opposite the Buck Hollow Overlook, about 2 km southeast of Thornton Gap (table 1, sample. 32; Thornton Gap 7.5-minute quadrangle) (fig. 2*D*). This unit is composed of 28 to 49 percent alkali-feldspar microperthite (chiefly microcline), 14 to 30 percent plagioclase, 28 to 35 percent quartz, 4 to 6 percent orthopyroxene, less than 1 to 7 percent garnet, less than 1 to 3 percent biotite, and rare clinopyroxene (Tollo, Aleinikoff, Borduas, and Hackley, 2004; Tollo, Bailey, Borduas, and Aleinikoff, 2004). Gneissic layering and foliation are defined by the planar alignment of minerals within interlayered quartzofeldspathic and orthopyroxene- and biotite-rich domains that are 1 to 70 cm thick. The unit occurs as two 6-km-long xenoliths within the megacrystic orthopyroxene syenogranite-monzogranite gneiss (Yos; 1,159±14 Ma). At the contacts between the two gneisses, the younger gneiss is finer grained as the result of contact-related cooling.

Orthopyroxene Quartz Diorite Gneiss

Orthopyroxene quartz diorite gneiss (Yoq, 1,161±10 Ma) occurs as a xenolith that is 4 km long and 1 km wide within garnetiferous metasyenogranite (about 1,063 Ma) of the Old Rag Granite. The reference locality is along the east side of Virginia State Route 649, about 1.5 km north of Boyd's Mill (table 1, sample 31; Chester Gap 7.5-minute quadrangle). The gneiss is composed of 20 to 25 percent orthoclase, 60 to 65 percent plagioclase, 12 to 15 percent quartz, 5 to 7 percent orthopyroxene, and 2 to 4 percent biotite (Tollo, Aleinikoff, Borduas, and Hackley, 2004; Tollo, Bailey, Borduas, and Aleinikoff, 2004). The low-silica composition (55 to 57 weight percent) and trace-element concentrations of high-field-strength elements suggest sources with a significant volcanic-arc component (Tollo, Aleinikoff, Borduas, and Hackley, 2004; Tollo, Bailey, Borduas, and Aleinikoff, 2004; Tollo and others, 2006). Excavated blocks of this rock along the east bank of Gooney Run show compositional layers defined by alternating domains of (1) quartz and feldspar, and (2) garnet, biotite, and orthopyroxene.

Granodiorite Gneiss

Granodiorite gneiss (Ygd; 1,161±9 Ma) is leucocratic. The reference locality is on the southwest side of Turkey Mountain north of Fletcher Mill (table 1, sample 30; Washington 7.5-minute quadrangle) (Tollo and others, 2006). The granodiorite is composed primarily of plagioclase, microcline, quartz, and biotite. Compositional layers of centimeter-thick leucocratic dikes and sills are parallel to foliation. Northwest-dipping gneissic foliation of Mesoproterozoic age is cut by southeast-dipping spaced cleavage of Paleozoic age (fig. 3*A*).

Megacrystic Orthopyroxene Syenogranite-Monzogranite Gneiss

The compound unit, megacrystic orthopyroxene syenogranite-monzogranite gneiss (Yos; 1,166±14 and 1,159±14 Ma), occurs in the highlands from near Stony Man (Old Rag Mountain 7.5-minute quadrangle) north to near Pignut Mountain (Thornton Gap 7.5-minute quadrangle). The 1,166±14 Ma rock occurs as a xenolith that is 2.5 km long and 1 km wide within megacrystic biotite monzogranite (Ybm; about 1,057 Ma), on Slaughter Mountain within the Woodville 7.5-minute quadrangle (table 1, sample 34). The reference locality of the 1,159±14 Ma rock is the south portal of the tunnel at Marys Rock along Skyline Drive south of Thornton Gap (fig. 3*B*; table 1, sample 29; Tollo, Aleinikoff, Borduas, and Hackley, 2004; Tollo, Bailey, Borduas, and Aleinikoff, 2004; Tollo and others, 2006). The rock is composed of 22 to 57 percent alkali-feldspar microperthite (chiefly microcline), 10 to 37 percent plagioclase, and 11 to 49 percent quartz; it also contains as much as 14 percent orthopyroxene, 11 percent amphibole, and rare clinopyroxene (Tollo, Aleinikoff, Borduas, and Hackley, 2004; Tollo, Bailey, Borduas, and Aleinikoff, 2004). The gneissic layering is defined by alternating quartzofeldspathic layers and domains of orthopyroxene and (or) amphibole, ranging from about 1 cm to greater than 13 cm thick. A strong foliation is parallel to the compositional layering. Weathered gneiss forms light-gray, spheroidal boulders. Fresh rock at the portal is dark and contains diagnostic subhedral to euhedral, monocrystalline alkali-feldspar megacrysts as much as 13 cm long. The gneiss was intruded by the Old Rag Granite (about 1,063–1,060 Ma) and occurs as a kilometer-scale xenolith within it.

Orthopyroxene Monzogranite-Quartz Monzodiorite Gneiss

The compound unit, orthopyroxene monzogranite-quartz monzodiorite gneiss (Yomg; 1,158±13 Ma), occurs in the southern part of the map area within the Crozet, Browns Cove, Free Union, Swift Run Gap, and Stanardsville 7.5-minute quadrangles. The reference locality is on the south side of Virginia State Route 810 north of Nortonsville along the Greene and Albemarle County line (table 1, sample 28; Free Union 7.5-minute quadrangle). Other good exposures occur along the north side of Virginia State Route 674 north of the Moormans River in the Browns Cove 7.5-minute quadrangle. The gneiss is composed of perthite, quartz, plagioclase, orthopyroxene, clinopyroxene, and biotite. The gneiss has a strong foliation that is parallel to millimeter- to centimeter-wide compositional layers defined by alternating quartzofeldspathic and pyroxene-rich domains, which weather as finely spaced ribs. Leucocratic layers as much as 5 cm thick are transposed intrafolial folds.

Lineated Leucogranite Gneiss

Lineated leucogranite gneiss (Yll, 1,150±23 Ma) occurs in the southern part of the Shenandoah National Park area in the Browns Cove, Free Union, Swift Run Gap, and Stanardsville 7.5-minute quadrangles. The reference locality is on the south side of Roundtop, along the Skyline Drive, about 0.5 km east of Powell Gap (table 1, sample 27). The leucogranite has a chemical composition that ranges from alkali-feldspar granite to monzogranite. The rock is composed of 40 percent perthite, 40 percent quartz, and 15 to 20 percent plagioclase (Bailey and others, 2006). This gneiss also contains as much as 15 percent biotite and is characterized by distinctive lineations that give the rock a striped texture.

Flint Hill Gneiss

The Flint Hill Gneiss (Yfh; 1,144±8 Ma) (Lukert and others, 1977) is a strongly deformed syenogranite with layers of leucogranite and blue quartz (fig. 4*A*). The sample locality for U-Pb analysis is along the roadcut in a county road adjacent to Hittles Mill Stream, west of U.S. Route 522, and northwest of Flint Hill (table 1, sample 26; Chester Gap 7.5-minute quadrangle). The reference locality is a collection of roadcuts along U.S. Route 522, about 0.9 km south of Flint Hill. The gneiss is composed of 35 percent microcline, 20 percent plagioclase, and 20 percent quartz, 5 percent chlorite, and 2 percent biotite (Tollo, Aleinikoff, Borduas, and Hackley, 2004; Tollo, Bailey, Borduas, and Aleinikoff, 2004). Gneissic layering is defined by alternating quartzofeldspathic domains and biotite-rich domains that are typically 1 to 10 cm thick. Foliation is defined by the planar alignment of biotite that is parallel to layers of leucogranite and blue quartz and commonly is mylonitic. The Mesoproterozoic compositional layering and mylonitic foliation are contorted and kinked by Paleozoic cleavage (Clarke, 1984). The Flint Hill Gneiss contains diagnostic grains of blue quartz and veins of blue quartz parallel to the gneissosity. Dikes of leucocratic granite locally crosscut it (Clarke, 1984).

Group 2

Orthopyroxene Granite Monzogranite

Orthopyroxene granite-monzogranite (Yog, 1,120±12 and 1,111±16 Ma) underlies the rugged mountains around the town of Washington in the Washington and Chester Gap 7.5-minute quadrangles and underlies the northwest slopes of Stony Man and Skyland in the Old Rag Mountain and Big Meadows 7.5-minute quadrangles. The reference locality of the rock dated at about 1,120 Ma is on the east side of The Peak, northwest of the county road between Washington and Flint Hill (table 1, sample 25). The sample locality of the rock dated at about 1,111 Ma is on the northwest slope of Stony Man and Skyland (table 1, sample 24) (Tollo, Aleinikoff, Borduas, and Hackley, 2004; Tollo, Bailey, Borduas, and Aleinikoff, 2004; Tollo and others, 2006). These rocks are composed of 30 to 40 percent microcline, 30 to 40 percent plagioclase, 15 to 20 percent quartz, 5 to 7 percent orthopyroxene, 5 percent biotite, as much as 2 percent amphibole, and rare garnet (Tollo, Aleinikoff, Borduas, and Hackley, 2004; Tollo, Bailey, Borduas, and Aleinikoff, 2004). Foliation is defined by discontinuous, centimeter-thick domains of orthopyroxene, amphibole, biotite, and garnet. Dikes of coarse-grained, nonfoliated, alkali feldspar- and blue quartz-bearing leucogranite and coarse-grained, nonfoliated, garnet-bearing leucogranite, as much as 1.5 m wide, locally intruded these rocks parallel to the foliation. Xenoliths up to 2 km long and 2 km wide occur within the orthopyroxene monzogranite-quartz monzodiorite unit (Yomg), which is dated at about 1,050 Ma.

Group 3

Porphyroclastic Metagranitoid

Porphyroclastic metagranitoid (Yml, 1,078±9 Ma) occurs along the east bank of the Conway River about 1.5 km north of Fletcher (table 1, sample 23; Fletcher 7.5-minute quadrangle) (fig. 4*B*). This unit occurs as a xenolith about 0.5 km wide and 2.2 km long within the orthopyroxene monzogranite-quartz monzodiorite unit (Yom, about 1,050 Ma) adjacent to an outlier of metabasalt of the Catoctin Formation; smaller xenoliths are abundant. Intrusion of the orthopyroxene monzogranite-quartz monzodiorite and extrusion of the metabasalt resulted in hydrothermal alteration of the metagranitoid, with epidote and potassium feldspar converting it to unakite. The very coarse grained metagranitoid is composed of 80 percent alkali-feldspar-dominant mesoperthite (chiefly microcline), 20 percent gray quartz, and less than 1 percent biotite (Tollo, Aleinikoff, Borduas, and Hackley, 2004; Tollo, Bailey, Borduas, and Aleinikoff, 2004). Foliation is defined by 2 domains, interlayered quartz and feldspar, and quartz-rich domains, ranging from less than 3 cm to greater than 13 cm, and tabular megacrysts of monocrystalline, subhedral to euhedral microcline as much as 10 cm long.

Old Rag Granite

The Old Rag Granite (Yor) (Lukert, 1982) consists of garnetiferous syenogranite (1,063±8 Ma; fig. 4*C*) and garnetiferous leucogranite (1,060±5 Ma; fig. 4*D*). The unit extends from Whiteoak Canyon in the Old Rag Mountain 7.5-minute quadrangle north to near Front Royal. The garnetiferous syenogranite (1,063±8 Ma) underlies the cove at Browntown (Chester Gap 7.5-minute quadrangle), where the reference locality for this variation is along the lower slope of Buck Mountain above the north bank of Gooney Run (table 1, sample 22; Chester Gap 7.5-minute quadrangle). At the reference locality, the unit ranges from alkali-feldspar granite to syenogranite and is composed of 35 to 40 percent orthoclase, 15 to 25 percent plagioclase, 20 to 25 percent gray to white quartz, 2 to 10 percent garnet, 2 to 5 percent biotite, and 2 to 5 percent pseudomorphs after orthopyroxene (Tollo, Aleinikoff, Borduas, and Hackley, 2004; Tollo, Bailey, Borduas, and Aleinikoff, 2004). The light-gray syenogranite contains diagnostic clots of dark-red garnet that locally are as much as 3 cm in diameter (fig. 4*C*). This unit was intruded by dikes of medium-grained, equigranular alkali-feldspar granite, syenogranite, and pegmatite.

The garnetiferous leucogranite variation (1,060±5 Ma) extends from Whiteoak Canyon in the Old Rag Mountain 7.5-minute quadrangle north to near Little Hogback Mountain in the Washington 7.5-minute quadrangle. The reference locality for the garnetiferous leucogranite is the crest of Old Rag Mountain (table 1, sample 20; Hackley, 1999; Tollo and others, 2006). The rocks include coarse-grained varieties composed of variable amounts of alkali feldspar, plagioclase, blue quartz, and accessory garnet, biotite, and orthopyroxene (Hackley, 1999, 2000, 2006). The Old Rag Granite intruded Group 1 orthopyroxene-bearing gneisses and there are map-scale xenoliths of the gneiss within it. The garnetiferous leucogranite was intruded by megacrystic biotite monzogranite (Ybm) and orthopyroxene monzogranite-quartz monzodiorite (Yom).

Crozet Granite

The Crozet Granite (Ycg, 1,060±7 Ma) (Nelson, 1962) is a very coarse grained syenogranite that underlies a broad area from the Crozet 7.5-minute quadrangle north to the Swift Run Gap 7.5-minute quadrangle (fig. 5*A*). The reference locality is along the Moormans River immediately downstream of the dam at the Charlottesville Reservoir in Sugar Hollow (table 1, sample 21; Browns Cove 7.5-minute quadrangle). Megacrysts of euhedral feldspar as much as 10 cm long are randomly oriented and locally define a flow foliation. Undeformed dikes of pegmatite as much as 0.5 m thick intruded the granite. The Crozet Granite may be the same rock known as the Vesuvius Megaporphyry, which was described by Werner (1966).

Megacrystic Biotite Monzogranite

Megacrystic biotite monzogranite (Ybm, 1,057±8 and 1,057±7 Ma) underlies a broad area from Etlan in the Old Rag Mountain 7.5-minute quadrangle north to Sperryville in the Washington 7.5-minute quadrangle. The sample localities are on Long Mountain (table 1, sample 18) and along the Covington River west of Rock Mills (table 1, sample 19), both of which are in the Washington 7.5-minute quadrangle. The reference locality is sample 19.

Megacrystic Quartz Monzonite

Megacrystic quartz monzonite (Ypb; 1,049±6 Ma) occurs in the extreme northeast part of the map area in the Linden and Flint Hill 7.5-minute quadrangles. The reference locality is in the adjacent Upperville 7.5-minute quadrangle (Nelson, 1997; Aleinikoff and others, 2000, 2008), which in turn is within the Washington West 30′×60′ quadrangle (Davis and others, 2002). The coarse-grained quartz monzonite intruded the Flint Hill Gneiss (dated at about 1,144 Ma), which was in turn intruded by the Neoproterozoic Cobbler Mountain Alkali Feldspar Quartz Syenite (Zrc; 722±3 Ma) of the Robertson River Igneous Suite. The dominant foliation is marked by augen of potassium feldspar (fig. 5*B*) and interpreted to be Paleozoic in age.

Orthopyroxene Monzogranite-Quartz Monzodiorite

The compound unit, orthopyroxene monzogranite-quartz monzodiorite (Yom; 1,050±8, 1,049±9, and 1,044±6 Ma), is a dark, massive, nonfoliated pluton in the center of the map area (fig. 5*C*). This unit occurs from near Swift Run Gap in the Swift Run Gap 7.5-minute quadrangle to north of Skyland in the Big Meadows 7.5-minute quadrangle. Similar rocks occur as dikes throughout the region. The reference locality is on a dirt road on the east side of Bluff Mountain within the Rapidan Wildlife Management Area, about 3 km northwest of Graves Mill (table 1, sample 17; Fletcher 7.5-minute quadrangle); a sample from the type locality yielded an age of 1,050±8 Ma. At the reference locality, the rock is composed of 9 to 30 percent alkali-feldspar microperthite (chiefly microcline), 30 to 49 percent plagioclase, 14 to 26 percent quartz, 10 to 17 percent orthopyroxene, as much as 7 percent amphibole, and rare clinopyroxene (Tollo, Aleinikoff, Borduas, and Hackley, 2004; Tollo, Bailey, Borduas, and Aleinikoff, 2004). Weak foliation is defined by the planar alignment of the ferromagnesian minerals. Subhedral to euhedral, monocrystalline alkali-feldspar megacrysts as much as 10 cm long on weathered surfaces have a distinctive orange rind. This rock is locally cut by dikes of coarse-grained leucogranite. The location of the sample that yielded the 1,049±9 Ma date is along the Skyline Drive southwest of Swift Run Gap(Swift Run Gap 7.5-minute quadrangle (table 1, sample 16), and the locality of the 1,044±6 Ma rock is on Jenkins Mountain in the Washington 7.5-minute quadrangle (table 1, sample 15).

Biotite Monzogranite-Quartz Monzodiorite

The compound unit, biotite monzogranite-quartz monzodiorite (Ybg; 1,040±9, 1,032±10, and 1,028±9 Ma), is the youngest known rock in the Blue Ridge of Virginia. The reference locality is the 1,028±9 Ma rock (table 1, sample 12) on the west side of the driveway of a parsonage on the north side of Virginia Route 670 in Criglersville (Madison 7.5-minute quadrangle; Bailey and others, 2003) (fig. 5*D*). This rock is very dark due to as much as 25 percent biotite and blue quartz. The dominant mineral assemblage is alkali feldspar, plagioclase, quartz, and biotite (Tollo, Aleinikoff, Borduas, and Hackley, 2004; Tollo, Bailey, Borduas, and Aleinikoff, 2004; Bailey and others, 2006). The rock does not have a conspicuous Mesoproterozoic foliation, but greenschist-facies mineral assemblages define a pervasive foliation of Paleozoic age. The unit contains light-gray xenoliths of foliated leucogranite, which are interpreted to be derived from leucogranite gneiss dated at about 1,183 Ma. The location of the rocks dated at 1,040±9 and 1,032±10 Ma is a single outcrop on Poortown Mountain (table 1, 2 samples at location 14; Washington 7.5-minute quadrangle). These rocks may be equivalent to similar biotite granitoids called the Lovingston Formation (Jonas, 1928; Virginia Division of Mineral Resources, 1993) south of Charlottesville.

Neoproterozoic rocks

Introduction

The oldest recognized Neoproterozoic rocks are metasedimentary paragneisses deposited unconformably above Mesoproterozoic gneisses after about 960 Ma (Southworth and others, 2008). Plutonic rocks of the Robertson River Igneous Suite intruded Mesoproterozoic gneisses between 730 and 700 Ma (Tollo and Aleinikoff, 1996). Volcanic rocks associated with the Robertson River Igneous Suite were locally deposited at about 714 and about 700 Ma (in the Chester Gap and Castleton 7.5-minute quadrangles, respectively) and were intercalated with clastic sedimentary rocks (Tollo and Hutson, 1996). Unconformably overlying the Mesoproterozoic gneisses and Robertson River Igneous Suite and underlying the Neoproterozoic Catoctin Formation are clastic sedimentary rocks of the Mechum River and Swift Run Formations. These rocks occupy the same general stratigraphic position. Rocks of the Mechum River Formation are confined to the medial area of the Blue Ridge anticlinorium (Bailey, Peters, and others, 2007), where they form a linear belt more than 80 km long along the eastern margin of the map. The Swift Run Formation is mostly confined to the west limb of the anticlinorium and in several outliers within the Mesoproterozoic rocks around Yanceys Mill and Boonesville. The Catoctin Formation, the youngest Neoproterozoic unit, consists of basaltic volcanic rocks that locally unconformably overlie Mesoproterozoic gneiss and the Swift Run Formation. Associated with these volcanic flows are metadiabase feeder dikes that intruded the Mesoproterozoic and older Neoproterozoic rocks.

Garnet-Graphite Paragneiss

Distinctive rusty-weathered paragneiss (Zp) occurs as isolated, poorly exposed bodies (between about 0.25 to 5 km long) that are surrounded by orthogneisses and granitoids ranging in age from about 1.18 to 1.03 billion years (giga-annum, or Ga) (fig. 6). Other than a small pegmatite, these rocks lack crosscutting relations observed in the orthogneisses and outcrop-scale xenoliths have not been recognized within the granitoids. Burton and Southworth (1993) and Southworth, Brezinski, and others (2007) interpreted these layered metamorphic rocks as inliers of the older crust that were intruded by younger orthogneisses. The graphitic rocks consist of alternating quartz-plagioclase-potassium feldspar, garnet-biotite, and quartzite layers, which locally contain as much as 25 percent modal garnet (up to 1 cm in diameter). The mineralogy and texture suggest that the protoliths were impure sandstone or graywacke.

Detrital zircons separated from quartzite layers within the paragneisses are round, frosted, and pitted and have a wide range of colors and truncated internal oscillatory zones. These zircons were dated by the U-Pb isotopic technique using the USGS-Stanford University SHRIMP-RG. Results from three samples from three different localities indicate that the rocks contain detrital zircons that are younger than the crystallization and metamorphic ages of the enclosing orthogneisses. Paragneiss along Buck Mountain Creek (table 1, sample 11; Free Union 7.5-minute quadrangle) yielded six detrital zircons that range in age from 1,011 to 997 Ma. Paragneiss exposed along Lickinghole Creek (table 1, sample 10; Crozet 7.5-minute quadrangle) yielded three detrital zircons that range in age from 1,011 to 959 Ma.

Paragneiss along Garth Run (table 1, sample 9; Fletcher 7.5-minute quadrangle) yielded seven detrital zircons that range in age from 1,007 to 812 Ma. The minimum ages of the detrital zircons suggest that the sediments were deposited after 997, 959, and 812 Ma, respectively. Therefore, the sedimentary protoliths of the paragneisses were unconformably deposited on the Mesoproterozoic plutonic rocks when the plutonic rocks were exposed at the surface.

Field relations suggest that two of the paragneiss outcrops are overlain by Neoproterozoic phyllite and graywacke of the Swift Run Formation. The reference locality of these rocks are on a bluff along the bank of Buck Mountain Creek (Free Union 7.5-minute quadrangle). The paragneisses are likely erosional remnants of the missing rock record between the emplacement of the biotite monzogranite-quartz monzodiorite (Ybg) dated at 1,028 Ma and the Robertson River Igneous Suite dated at 730 Ma . The deep-crustal Mesoproterozoic rocks were uplifted, and zircons and sediment derived from them were locally deposited in shallow basins. The strata were then buried, metamorphosed, and uplifted to the surface before deposition of younger Neoproterozoic sediments between about 700 and 575 Ma.

Robertson River Igneous Suite

The Robertson River Igneous Suite constitutes a batholith that extends continuously for 110 km from the Charlottesville area to Markham. Exposures are found in the northeast part of the map area north of Madison. The rocks were named by Allen (1963) after exposures along the Robinson River, but the misuse of the name “Robertson” has prevailed. Field mapping, chemical analysis, and U-Pb thermal ionization mass spectrometry (TIMS) isotope analyses were used to differentiate the batholith into nine units of peralkaline to metaluminous granites and syenites, including several local areas of volcanic rocks (Tollo and Lowe, 1994; Tollo and Aleinikoff, 1996; Tollo and Hutson, 1996). The linear belt of massive igneous rocks was likely emplaced along a fracture zone during crustal extension (Bailey and Tollo, 1998; Burton and Southworth, 2004) and is analogous to a large dike.

The plutonic rock units within the map area include the Rivanna Granite (Zrr; 735±4 Ma), the Arrington Mountain Alkali Feldspar Granite (Zram; 730±4 Ma), the Laurel Mills Granite (Zrl; 729±1 Ma), the Cobbler Mountain Alkali Feldspar Quartz Syenite (Zrc; 722±3 Ma), the Hitt Mountain Alkali Feldspar Syenite (Zrh; 706±2 Ma), the Battle Mountain Alkali Feldspar Granite (Zrbg; 705±2 Ma), and the Amissville Alkali Feldspar Granite (Zra; 700 Ma). Most of these rocks are best exposed at their type localities within the map area: Arrington Mountain (Brightwood 7.5-minute quadrangle), along Mill Run just south of Laurel Mills (Massies Corner 7.5-minute quadrangle), Hitt Mountain (Woodville 7.5-minute quadrangle), Battle Mountain (Massies Corner 7.5-minute quadrangle), and the north side of U.S. Route 211 northwest of Amissville (Massies Corner 7.5-minute quadrangle). The Cobbler Mountain Alkali Feldspar Quartz Syenite (Zrc) is poorly exposed in the northeast part of the map area northeast of Cresthill (Flint Hill 7.5-minute quadrangle), but its type locality is on the north side of Interstate Route 66 on Little Cobbler Mountain, just east of the map area.

Spatially and temporally associated with some of the plutonic rocks are several areas of volcanic rocks (some intercalated with sedimentary rocks) and felsic dikes. East of Skyline Drive, between Jenkins Gap and Hogwallow Flat, quartz trachyte (Zrqt; about 719 to 714 Ma; table 1, samples 5 and 4; Chester Gap 7.5-minute quadrangle) (fig. 7*A*, *B*) overlies Mesoproterozoic orthopyroxene granodiorite gneiss (Yod) and orthopyroxene monzogranite-quartz monzodiorite (Yom). The quartz trachyte was intruded by metadiabase dikes and was overlain by Neoproterozoic metabasalt of the Catoctin Formation. The quartz trachyte rocks are considered to be the extrusive equivalent of the Cobbler Mountain Alkali Feldspar Quartz Syenite (about 722 Ma TIMS age), which occurs about 35 km to the east.

The felsite component of the Battle Mountain Alkali Feldspar Granite (Zrbf; about 702 Ma) is a volcanic rock located on the west side of Battle Mountain and along U.S. Route 522 near Boston (Woodville and Castleton 7.5-minute quadrangles). The rhyolite and conglomerate component of the Battle Mountain Alkali Feldspar Granite (Zrbr) is a rhyolitic rock that was intercalated with clasts of rhyolite, cobbles of Mesoproterozoic granitoids, and sedimentary rocks near Battle Mountain (Tollo and Hutson, 1996). These volcanic and sedimentary rocks are isolated from and are not contiguous with rocks of the Mechum River Formation (Bailey, Peters, and others, 2007). Felsic dikes that intruded Mesoproterozoic rocks and plutonic rocks of the Robertson River Intrusive Suite have the same composition as the felsite, so they are called felsic dikes of the Battle Mountain Alkali Feldspar Granite. The patches of volcanic rocks support the interpretation that a significant unconformity was present, which predated the deposition of the Neoproterozoic Swift Run and Catoctin Formations (Southworth, Aleinikoff, and others, 2007).

Regionally, rocks of the Robertson River Igneous Suite (730–700 Ma) are temporally and petrologically similar to other rocks in the Blue Ridge of southern Virginia and western North Carolina (Tollo, Aleinikoff, Bartholomew, and Rankin, 2004). The A-type granitoid plutons of the Crossnore Suite (Rankin, 1993) yielded U-Pb ages of 765 to 740 Ma (Su and others, 1994). Extrusive felsic volcanic rocks associated with the A-type plutons include (1) the metavolcanic rocks found at Ada, east of the map area in the Washington West 30′×60′ quadrangle (about 708 Ma U-Pb SHRIMP age), (2) the Mount Rogers Formation (Rankin, 1993; about 760 Ma TIMS age; J.N. Aleinikoff, unpub. data, 2008), in southwestern Virginia, and (3) the Grandfather Mountain Formation (Fetter and Goldberg, 1995) in northwestern North Carolina.

Mechum River Formation

Sedimentary rocks in the central portion of the Blue Ridge anticlinorium were named the Mechum River Formation (Zmr) (Gooch, 1958; Nelson, 1962; Schwab, 1986; Bailey, Peters, and others, 2007) after exposures along the railroad cuts and banks of the Mechums River at the small town of Mechums River (Crozet 7.5-minute quadrangle), but the misuse of the name “Mechum River” has prevailed. The rocks crop out in a belt 0.5 to 6 km wide that extends northeast about 100 km from near Batesville to near Ben Venue and U.S. Route 211 (Massies Corner 7.5-minute quadrangle) (Bailey, Peters, and others, 2007). The sedimentary rocks were deposited unconformably on Mesoproterozoic rocks. The sediments have been interpreted to be deposited in deep water (Gooch, 1958) by rivers and alluvial fans (Schwab, 1986) and in marine and nonmarine environments (Conley, 1989). Glaciogenic marine deposits in the southwestern part of the belt were overlain by marine turbidites, which in turn were overlain by fluviodeltaic deposits to the northeast (Bailey and Peters, 1998; Bailey, Peters, and others, 2007); the rocks become younger to the northeast.

The rocks of the 500-m-thick formation (Bailey, Peters, and others, 2007) range from boulder conglomerates to mudstones. The top of the formation is not preserved. At a scale of 1:24,000, the Mechum River can be mapped as three distinct rock types: (1) metaconglomerate, (2) meta-arkose and arkosic metawacke, and (3) metasiltstone and phyllite (Bailey, Peters, and others, 2007); they were not differentiated on this map. Metaconglomerate is well exposed on Bessie Bell Mountain (fig. 8*A*; Woodville 7.5-minute quadrangle). Boulder conglomerate and diamictite that occur locally at the base were derived mostly from Mesoproterozoic granitoids, but some boulders were derived from the Neoproterozoic Robertson River Igneous Suite. The conglomerate and diamictite are glaciomarine deposits laid down as proglacial outwash fans. Meta-arkose and arkosic metawacke dominate the formation from near Charlottesville to as far north as U.S. Route 33. The graded and scoured beds represent subaqueous gravity-flow deposits (turbidites) laid down as submarine fans with some fluviodeltaic, nonmarine influence (Bailey, Peters, and others, 2007).

The Mechum River Formation is loosely correlated with sedimentary rocks of the Swift Run Formation and the Lynchburg and Fauquier Groups, but nowhere do the formations spatially overlap. The Mechum River Formation occurs in the middle part of the anticlinorium, whereas Swift Run rocks are found on the west limb of the anticlinorium and the rocks of the Fauquier and Lynchburg Groups (Kasselas, 1993) are on the east limb of the anticlinorium. Mechum River rocks were considered to occupy a Neoproterozoic graben (Gooch, 1958; Schwab, 1986). Recent work suggests that they are in a footwall syncline beneath a Paleozoic thrust fault that placed Mesoproterozoic rocks above them (Bailey, Peters, and others, 2007). Rocks of the Swift Run Formation are mostly fluvial with minor lake sediments (Southworth, Brezinski, and others, 2007). Rocks of the Fauquier and Lynchburg Group are subaqueous to deepwater-marine sediments that accumulated in distinct basins that may have been bounded by faults (Kasselas, 1993). The age of the Mechum River Formation is poorly constrained but is bracketed between about 700 and 575 Ma because it contains clasts derived from the Robertson River Igneous Suite (Tollo and Hutson, 1996; Bailey, Peters, and others, 2007) and was intruded by metadiabase dikes that were probably related to Catoctin Formation volcanism.

Swift Run Formation

Introduction

Quartzose and phyllitic rocks between the Mesoproterozoic gneisses and the Catoctin Formation initially were named the Swift Run tuff by Jonas and Stose (1939; Stose and Stose, 1946). They designated the type locality as exposures on U.S. Highway 33 just east of Swift Run Gap and on the Skyline Drive just north of the gap. The protoliths of the Swift Run tuff were considered to be sedimentary and volcanic (Jonas and Stose, 1939; Stose and Stose, 1946), but the unit was renamed the Swift Run Formation by King (1950) because the volcanic component was considered to be subordinate. The Swift Run rocks also occur as outliers in the basement rocks southeast of Blue Ridge. The clastic rocks generally are thin and discontinuous along strike. Regionally, the Swift Run Formation (Zsr) has been mapped from the type locality to the north around the plunging nose of the Blue Ridge-South Mountain anticlinorium in Maryland (Brezinski and Fauth, 2005), and down the east limb of the anticlinorium about 40 km to near Leesburg (Southworth, Brezinski, and others, 2007). The rocks are terminated against a transverse down-to-the-north normal fault, and two transverse normal faults separate rocks of the Swift Run Formation from rocks of the Fauquier Group (Southworth, Brezinski, and others, 2007). The faults were probably active during deposition (Kline and others, 1991). Rocks of the Swift Run Formation have been mapped as far south as the Tye River (Bloomer and Werner, 1955; Bailey and others, 2002). On the western limb, the southernmost occurrence is in a fault block between U.S. Route 60 and the James River (Bloomer and Werner, 1955), approximately 5 km southeast of the west limb of the Blue Ridge anticlinorium.

Previous Studies

King (1950) recognized gray, pink, and green arkosic quartzite; conglomerate; gray and purple laminated slate; and green, blebby pyroclastic slate in the Shenandoah National Park area. He suggested that the rocks of the Swift Run and Catoctin Formations were deposited concurrently and stratigraphically overlapped; metabasalts were interbedded with the uppermost Swift Run rocks and sedimentary rocks were interbedded with metabasalt of the Catoctin Formation. From Big Meadows north to Elkwallow Gap, Reed (1955, 1969) considered the thin and discontinuous arkose, conglomerate, and graywacke as the lower part of the Catoctin in which the thickest sections of sedimentary rocks originated in depositional basins. He noted that there were fragments of phyllite and metabasalt in the conglomerate and he considered that the laminated slates and phyllites north of Hawksbill were rhythmites and altered volcanic ash (Reed, 1969). Gathright (1976) described the Swift Run Formation as a heterogeneous and discontinuous sequence of metaconglomerate, metasandstone, tuffaceous phyllite, and coarse-grained volcaniclastic rocks. The volcaniclastic rocks were derived from the earlier phases of Catoctin volcanism such that pyroclastic rocks grade upward into the Catoctin (Gathright, 1976). He assigned thin beds of metabasalt interbedded with sedimentary rocks to the Swift Run. The Swift Run Formation rocks are alluvial-fan, flood-plain, and lacustrine sediments (Schwab, 1986) deposited on an upland that had paleotopographic relief.

Stratigraphic Relationships Within Shenandoah National Park

The rocks of the Swift Run Formation are highly variable. Monolithic conglomerate composed of cobbles of Mesoproterozoic granitoids (fig. 8*B*) and rounded cobbles of vein quartz likely originated as colluvium (fig. 8*C*). Locally, the basal rocks are clean, medium-grained, crossbedded orthoquartzite, such as on the west side of Hawksbill (near Hogwallow Flat) and west of Chester Gap. Pinkish-green quartzite, beneath and intercalated with the lowest metabasalt of the Catoctin Formation, consists of quartz, pink potassium feldspar, yellowish-green epidote, and volcanic detritus. Well-bedded quartzite (fig. 9*A*) may have originated as unconsolidated fluvial sand deposits that were buried and hydrothermally altered by basalt flows.

Noteworthy outcrops of Swift Run Formation lithologies are (1) phyllite (fig. 9*B*), conglomeratic phyllite, and metasandstone west of Powell Gap (Forte and others, 2005; A.M. Forte, K.M. Wooton, B.A. Hasty, and C.M. Bailey, College of William and Mary, unpub. data, 2005; Swift Run Gap 7.5-minute quadrangle)), (2) quartzite, meta-arkose, and metaconglomerate from Sugar Hollow northeastward across Pigeon Top Mountain (Olney and others, 2007; J.G. Olney, M.H. Lamoreaux, J.E. Tadlock, O. Nicholls, and C.M. Bailey (College of William and Mary, unpub. data, 2007; Browns Cove 7.5-minute quadrangle), (3) an outlier of quartzite, quartz-sericite schist (fig. 9*C*), and phyllite exposed around Boonesville (Free Union 7.5-minute quadrangle), and (4) coarse-grained meta-arkose, quartzite, and phyllite east of Hogwallow Flat along the fire road near Sprucepine Branch (Gathright, 1976; Chester Gap 7.5-minute quadrangle).

Regionally, the rocks that directly underlie the Catoctin Formation are variable and include Mesoproterozoic gneisses, sedimentary rocks of the Swift Run Formation (west limb of the Blue Ridge anticlinorium), and Fauquier Group and Lynchburg Group (east limb of the anticlinorium; not shown on the map). In at least five locations, there is evidence that the uppermost sedimentary rocks assigned to the Swift Run Formation and the volcanic rocks of the Catoctin Formation are interbedded and gradational. All but one location are within the map area.

1. Northwest of Big Meadows (Big Meadows 7.5-minute quadrangle), sedimentary rocks in the upper Swift Run are overlain by thin basalt flows that are cut by anastomosing dikelets of light-tan fine-grained quartzite. This relation suggests that the original sand was unconsolidated and was injected as liquefied dikes soon after the basalt was laid down.
2. Along the West Branch of Naked Creek in Weaver Hollow (Elkton East 7.5-minute quadrangle), Mesoproterozoic gneiss that is altered to unakite is overlain by a sequence of metabasalt, quartzite, and sediment derived from metabasalt.

3. East of Compton Gap (Chester Gap 7.5-minute quadrangle), the basal conglomerate contains clasts of metabasalt, which is overlain by metabasalt interbedded with calcareous crossbedded quartzite and arkose; the arkose contains clasts of metabasalt.
4. In Sugar Hollow, there are thin layers of metabasalt interbedded with phyllite (assigned to the Swift Run Formation), and metabasalt of the Catoctin Formation is interbedded with meta-arkose and phyllite (Bailey, Kunk, and others, 2007; Browns Cove 7.5-minute quadrangle).
5. One locality in northern Virginia exposes a vesicular metabasalt conglomerate interbedded with quartz-sericite schist of the Swift Run Formation (Southworth, 1995).

Furthermore, in two places near Wintergreen, to the south of the map area, Bartholomew (1977) showed metabasalt beneath sedimentary rocks assigned to the Swift Run Formation. The basal conglomerate contains fragments of tuffaceous phyllite and both units have the same mineralogy (Bartholomew, 1977). Further to the southwest, gradational relations between the Swift Run and Catoctin Formations were described by Bloomer and Werner (1955) along the Oronoco belt, which extends 20 km from U.S. Route 60 southward to the James River. They described conglomerate, graywacke, sandstone that had been affected by epidotization, a 26-m-thick bed of andesite, tuffs, and metabasalt as rocks of the Swift Run Formation. They also described paleosaprolite as much as 30 m thick that separated unaltered Mesoproterozoic gneisses from the basal Swift Run. They suggested that the reddish-brown slate that occurs in the Swift Run, Catoctin, and Unicoi Formations was volcanic tuff that did not have preserved shards of glass. All of the sedimentary rocks were assigned to the Catoctin Formation and they pinch out near the James River. Southwest of the James River, the basal clastic rocks have been called the Unicoi Formation at the base of the Chilhowee Group. They consist of conglomerate and meta-arkose interbedded with metasiltstone, tuffaceous phyllite, and metaconglomerate of scoriaceous metabasalt. A 4.5-m-thick bed of red metabasalt locally is near the base of the Unicoi. These rocks were called Unicoi after similar rocks about 300 km to the southwest around Watauga Lake, east of Elizabethton, Tenn.

Northeast of the map area, the Swift Run Formation consists of a fining-upward sequence of metasandstone, schist, phyllite, and marble. Crossbedded, arkosic metasandstone contains pebbles and cobbles of quartz, phyllite, and ferruginous sandstone. The Swift Run grades upward and laterally into quartz-sericite schist, which in turn grades upward into silver, gray, tan, and purple phyllite that contains sand grains. Podiform bodies of calcitic to dolomitic marble are locally within the quartz-sericite schist and phyllite. The rocks types are not present in the same stratigraphic position and reflect complex facies variations. The basal rocks locally include paleosols, which are fossil soils and (or) grus layers from in-place weathering of Mesoproterozoic gneiss.

The map distribution of the folded rocks of the Swift Run Formation indicates that (1) the heterogeneous rocks represent primary deposits in small, isolated basins on a paleotopographic surface that had considerable relief, and (or) (2) there could have been an erosional event before the Catoctin Formation was deposited, thereby producing a disconformity and destroying evidence that the deposits once were continuous. On the east limb of the Blue Ridge anticlinorium, soft-sediment deformation of a marble bed at the contact between the Fauquier Group and Catoctin Formation suggests that basalt flowed onto the calcium-carbonate-rich mass before it was lithified (Jay Kaufman, University of Maryland, written comm., 2008). Elsewhere on the east limb, marble occurs at the contact between the Swift Run and Catoctin Formations at five localities; locally, marble is found in the lower, middle, and upper parts of the Catoctin Formation (Southworth, Brezinski, and others, 2007). Neoproterozoic metadiabase dikes intruded rocks of the Swift Run Formation in at least six places and rocks of the Fauquier Group in at least eight places (Southworth, Brezinski, and others, 2007), suggesting that the older sediments were already lithified. Also, a metarhyolite dike intruded rocks of the Swift Run

Formation. Therefore, volcanism associated with the Catoctin Formation may have started during deposition of the upper Swift Run, whereas most of it postdates the Catoctin.

The age of the Swift Run Formation is poorly constrained. Swift Run rocks were deposited on Cobbler Mountain Syenite (722 Ma; Tollo and Aleinikoff, 1996) and also were intruded by a metarhyolite dike that has a SHRIMP U-Pb age of 555±4 Ma (Southworth and Aleinikoff, 2007). Deposition of the upper part of the Swift Run locally could have been synchronous with Catoctin volcanism, but the middle and lower parts could be considerably older.

Metadiabase Dikes

Tabular dikes of metadiabase (Zmd) intruded the Mesoproterozoic gneisses, Neoproterozoic Robertson River Igneous Suite, and Mechum River Formation in the Shenandoah National Park area. Elsewhere in northern Virginia, metadiabase dikes also occur in Neoproterozoic paragneisses, the Fauquier Group, and the Swift Run Formation. The dikes were subsequently metamorphosed and foliated during the Paleozoic. Most of the dikes are dark-green metadiabase, but a few are metagabbro (Bailey and others, 2003) and amphibolite (Lukert and Nuckols, 1976). The dikes commonly range in thickness from 0.5 to 5 m. The textures are mostly fine grained and aphanitic, although some are coarse grained and porphyritic. There are good exposures of metadiabase dikes that intruded Mesoproterozoic gneisses in roadcuts along Skyline Drive: (1) northeast of Mount Marshall (fig. 10*A*), (2) at the north portal of the tunnel at Marys Rock (fig. 10*B*), and (3) north of Stony Man. Trails on the top of Old Rag Mountain follow the trend of such dikes (fig. 10*C*; Hackley, 1999, 2000). On the west side of Carson Mountain, metadiabase dikes in basement rocks are very close to the contact of the overlying metabasalt flows of the Catoctin Formation. They most likely were feeder dikes to the volcanic flows of the Catoctin Formation, and the chemistry of the two are identical (Southworth, Burton, and others, 2006). The only reported exposure of “a dike feeding into a flow” is at the base of the Unicoi Formation in the James River gorge (Bloomer and Werner, 1955), southwest of the map area. Nowhere else in the Blue Ridge have dikes been documented to feed or cut the metabasalt flows.

Most of the dikes are probably the same age as the Catoctin Formation. Two metarhyolite dikes have U-Pb ages of 567±4 and 555±4 Ma (J.N. Aleinikoff, unpub. data, 2009.). Hornblende metagabbro dikes and sills at the base of the Mechum River Formation (Bailey and others, 2003) may be older. Metabasalt of the lower part of the Catoctin was intruded by a metarhyolite dike in Maryland (Fauth and Brezinski, 1994) and a metarhyolite flow of the Catoctin was cut by a metadiabase dike in Pennsylvania (Fauth, 1977).

Catoctin Formation

Introduction

The Catoctin Formation (Keith, 1894) was named after the volcanic rocks that underlie Catoctin Mountain on the east limb of the Blue Ridge-South Mountain anticlinorium in Maryland and Virginia. The Catoctin Formation consists mostly of green, gray, and purple, aphanitic, massive to schistose basalt that was metamorphosed during the Paleozoic to metabasalt. The metabasalt (Zcm) consists of albite laths in a fine-grained matrix of chlorite, actinolite, epidote, and pyroxene. The unit contains flow autobreccias, interbeds of pebble conglomerate, sandstone, metarhyolite (Zcr), laminated phyllite, and dark, variegated phyllite and slate. Laminated phyllite and sandstone beds (Zcs) are interpreted as alluvial deposits laid down on the volcanic landscape. Phyllite and slate beds (Zcp) containing spots and blebs of smeared chlorite have been interpreted as vesicular flows, volcanic tuffs, pumice, and volcanic ash. The Catoctin Formation locally unconformably overlies Mesoproterozoic gneiss and may be disconformable or transitional above the Neoproterozoic metasedimentary rocks of the Swift Run Formation. Where the molten lava altered the surface of the basement rocks, the granitic gneisses are bleached and altered to unakite, which consists of pink potassium feldspar, quartz, and yellowish-green epidote. Named after rocks in the Unaka Range in the Blue Ridge of eastern Tennessee and North Carolina, unakite (Bradley, 1874) is the unofficial state stone of Virginia.

Stratigraphy Within Shenandoah National Park

Reed (1955, 1969) studied the Catoctin Formation from Big Meadows northward to Stony Man (Big Meadows and Old Rag Mountain 7.5-minute quadrangles), Gathright (1976) studied the formation at North and South Marshall (Chester Gap 7.5-minute quadrangle), Badger and Sinha (1988) studied it in roadcuts along Interstate Route 64, and Badger (1992) studied the formation from Big Meadows to the Hawksbill-Spitler Hill area (Big Meadows 7.5-minute quadrangle). The Catoctin underlies most of the highlands of the Blue Ridge, including some of the highest peaks. Metabasalt flows typically form horizontal to moderately southeast-dipping homoclines (fig. 11*A*) with bedrock terraces and benches of individual flows that are visible on DEMs (Chirico and Tanner, 2004) and aerial photographs (Gathright, 1976). The distribution of the Catoctin Formation is controlled locally by faults, such as thrust faults along its eastern boundary (Fletcher, Big Meadows, Old Rag Mountain, Thornton Gap, and Chester Gap 7.5-minute quadrangles), and sixteen northwest-striking normal faults (Swift Run Gap 7.5-minute quadrangle). Also, the distribution is controlled by erosional remnants of folds such as along Conway River (Fletcher 7.5-minute quadrangle), and Moormans River (Crozet 7.5-minute quadrangle).

The majority of the formation is metabasalt (fig. 11*B*). The basalt was derived from tholeiitic magmas with upper mantle to midcrustal gabbroic fractionation from a mantle source at depths of 40 to 70 km (Badger and Sinha, 2004). The basal rocks locally are autobreccias and unakitic quartzite interbedded with metabasalt. Thin autobreccia consists of angular clasts of unakite in a matrix of metabasalt (west of Hogwallow Flat, Chester Gap 7.5-minute quadrangle); coarser volcaniclastic rocks occur on the west side of Skyline Drive between Compton Gap and Indian Run Overlook. Unakitic quartzite (which exhibits hues of pink, yellow, and green) contains clasts of metabasalt and is located beneath or within the lowest metabasalt (fig. 11*C*). The quartzites were likely unconsolidated fluvial sands that were hydrothermally altered by the basalt flows. They are found on Grindstone Mountain and in Weaver Hollow (Elkton East 7.5-minute quadrangle), on Bearfence Mountain (Fletcher 7.5-minute quadrangle), on the north side of Hawksbill (Big Meadows 7.5-minute quadrangle), and east of Carson Mountain (Chester Gap 7.5-minute quadrangle). Southwest of the Shenandoah National Park area, Bloomer and Werner (1955) described similar rocks, where epidotizing fluids from the lavas permeated the pore spaces of uncemented sand. Locally interbedded with the metabasalt are phyllite (laminated and spotted, Matthews Arm, Bentonville 7.5-minute quadrangle), and quartzite (Interstate Route 64 roadcuts, Waynesboro East 7.5-minute quadrangle). In several areas, the stratigraphy of the lava flows are well preserved. Between Stony Man and Big Meadows, and at Franklin Cliffs and Crescent Rocks, individual flows are generally 46 to 61 m thick, and one is 82 m thick (Reed, 1955, 1969, 1989; Fletcher and Big Meadows 7.5-minute quadrangles). Columnar-jointed metabasalt commonly marks the tops of flows (fig. 11*D*).

Northwest of Big Meadows are seven flows that have an average thickness of 66 m (Reed, 1955). Twelve to sixteen benches on Mount Marshall and Spitler Hill correspond with individual flows (Gathright, 1976). At least 10 flows on the north side of Big Meadows and 9 flows on Spitler Hill were correlated with those found on Hawksbill (Badger, 1992). In the Shenandoah National Park, the number of flows and their thickness decreases to the southwest (Badger and Sinha, 1988), presumedly away from the source feeder dikes. Volcanic breccias mark the base of the flows (fig. 12*A*). These breccias consist of fragmental basalt and phyllite cemented with quartz, epidote, and jasper. The phyllite localized strain and later developed into shear zones (Reed, 1969), as is seen on Stony Man. The lower and upper portions of the flows are commonly marked by amygdaloidal basalt (Reed, 1955; Gathright, 1976; Badger, 1992). The amygdules are relict air and gas cavities that were filled with fluids that precipitated white quartz, light-green epidote, dark-green chlorite, and pink calcite during greenschist-facies metamorphism (fig. 12*B*). Coarse-grained, white and pink plagioclase-feldspar phenocrysts locally form porphyritic beds in some flows (fig. 12*C*) (Reed, 1969). Discontinuous coarse-grained sandstone, quartzite, and phyllite beds also locally occur between flows (Gathright, 1976). In the extreme northeast part of the Catoctin Formation in the map area, a bed of light-gray metarhyolite tuff with white feldspar phenocrysts (Zcr; fig. 12*C*) is within metabasalt in the lower part of the formation near its base (Lukert and Nuckols, 1976).

The rocks at the top of the Catoctin Formation range from metabasalt, pyroclastic breccia (Byrds Nest No. 4, northwest of Beahm's Gap Overlook, Thornton Gap 7.5-minute quadrangle), and red- and purple-spotted phyllite and slate (fig. 13*A*) (Knob Mountain, Thornton Gap 7.5-minute quadrangle) to dark, variegated phyllite (fig. 13*B*) (Interstate Route 64, Waynesboro East 7.5-minute quadrangle). The spotted phyllite and slate were interpreted as amygdaloidal lava (Furcron, 1934), pyroclastic rocks (King, 1950), ash-fall tuffs (Gathright, 1976; Bartholomew, 1977), and felsic tuffs (Badger and Sinha, 1988). They also have been interpreted to be metamorphosed soil and saprolite derived from the subaerial in-place weathering of basalt (Nickelsen, 1956; Reed, 1969). Along Overall Run and near Byrds Nest No. 4, maroon, amygdaloidal metabasalt transitions upward into purple-spotted slate, which could be its weathered byproduct. The spots are likely deformed vesicles. The phyllite and slate are well exposed in the northern part of the park over a distance of about 20 km. From the Shenandoah National Park Headquarters northward to Byrds Nest No. 4 and westward to Overall Run is perhaps the thickest section of red phyllite and slate (25 m). The phyllite and slate contain diagnostic light-green- and white-smeared vesicles and altered phenocrysts consisting of dark-green chlorite and white sericite. These rocks are best exposed on the west side of Knob Mountain, southeast of Rileyville, where they are overlain by light-brown laminated siltstone of the Weverton Formation (Thornton Gap 7.5-minute quadrangle).

Some of the slates contain angular fragments of scoriaceous metabasalt (fig. 13*C*), thin beds of amygdaloidal lava with small white amygdules, volcanic breccias (fig. 13*D*) with jasper (King, 1950), and detrital quartz pebbles interbedded with rounded metabasalt clasts (fig. 13*E*). Gathright (1976) considered some of the slates to be devitrified volcanic ash because he noted glass shards and pumice lapilli with curlycue patterns in thin sections. These features are characteristic textures of ash-flow tuffs and ignimbrites. Phyllite along Interstate Route 64 (Gathright and others, 1977; Waynesboro East 7.5-minute quadrangle) and at the park headquarters (Thornton Gap 7.5-minute quadrangle) was sampled and found to contain only detrital zircons, suggesting that it was epiclastic sediment derived from the weathering of volcanic rocks. The spotted phyllite and slate at the top of the formation are likely a combination of ash-fall tuffs, weathered amygdaloidal basalt, and recycled saprolite bound by unconformities. The base of the phyllite and slate may be an nonconformity (King, 1950), but the upper contact between the phyllite and slate and the overlying rocks of the Weverton Formation likely is a disconformity (Keith, 1894; Furcron, 1934; Stose and Stose, 1946; King, 1950; Reed, 1955; Nickelsen, 1956; Rodgers, 1970; Gathright, 1976; Southworth and Aleinikoff, 2007; Southworth, Aleinikoff, and others, 2007).

Depositional Setting and Thickness

During Neoproterozoic extension associated with continental rifting, normal faults and fractures in the older basement gneiss served as conduits for metadiabase dikes. Where the dikes broke the surface, basaltic lava flows as much as 82 m thick locally accumulated. The thickest section is in excess of 700 m (Reed, 1955; Badger, 1992). Basaltic lava flowed into and filled the valleys and covered the adjacent hills. Erosion has exposed Mesoproterozoic gneiss in the cores of at least ten such paleohills. Seven paleohills are situated along Honey Run (Stanley 7.5-minute quadrangle), Harris Cove, Weaver Hollow, near Jollett (Elkton East 7.5-minute quadrangle), along Basin Hollow and the Rose River (Big Meadows 7.5-minute quadrangle), and in Whiteoak Canyon (Old Rag Mountain 7.5-minute quadrangle). The southernmost (Jollett and Weaver Hollow) and northernmost (Rose River and Basin Hollow) paleohills show Catoctin Formation overlying Mesoproterozoic gneiss. In the westernmost paleohills (northwest part of Harris Cove and Honey Run), the Catoctin Formation is absent and the Weverton Formation unconformably overlies the Mesoproterozoic gneiss. In the northern part of park, two paleohills are located at Fourway (Thornton Gap 7.5-minute quadrangle) and at Big Devils Stairs (Chester Gap 7.5-minute quadrangle). As the lava flowed over uncompacted sediment, dikes of fine-grained quartz sand were injected into the solidified basalt (Reed, 1955). Polygonal columnar joints formed as the lava cooled at the surface. Lava flows blocked streams and incorporated local sediment, such as the crossbedded quartzite within basalt near the base (south of Chester Gap, on Hawksbill, near Weaver Hollow, and on Bearfence Mountain). Pillow lava indicates that basalt locally flowed into ponds or impounded streams, such as at Dark Hollow Falls, east of Big Meadows (Badger, 1992).

The thickness of the Catoctin Formation ranges considerably because the volcanic flows overlie a landscape of hills and valleys with at least 300 m of topographic relief (Reed, 1955; Gathright, 1976). From south to north, the thickness of the formation is about 600 m along Interstate Route 64 (Badger and Sinha, 1988), about 700 m near Pasture Fence Mountain (Browns Cove 7.5-minute quadrangle), about 460 m at Big Meadows, and about 300 m at Mount Marshall (Chester Gap 7.5-minute quadrangle); south of Stanley, the unit is absent. Individual flows range from 46 to 82 m thick (Reed, 1969). Interbeds of conglomeratic sandstone and phyllite are as much as 61 m thick (Gathright, 1976), and the phyllite and slate at the top of the formation is as much as 60 m thick.

Age

Five samples of metabasalt and clinopyroxene minerals collected from the base to the top of the Catoctin Formation section exposed along Interstate Route 64 yielded an Rb-Sr isochron age of 570±36 Ma (Badger and Sinha, 1988). Beds of metarhyolite tuff in the lower (Lukert and Nuckols, 1976), middle (Southworth, 1994), and upper (Southworth, 1995; Southworth and others, 2006) parts of the Catoctin Formation were sampled for U-Pb analysis over a distance of 50 km from near Interstate Route 66 (Linden 7.5-minute quadrangle) to near the Potomac River. Zircons from the middle bed yielded a SHRIMP U-Pb zircon age of 568±4 Ma (J.N. Aleinikoff, unpub. data, 2009). In addition, porphyritic metarhyolite at the top of the formation in Pennsylvania yielded a SHRIMP U-Pb age of 559±5 Ma (Southworth and Aleinikoff, 2007). Two metarhyolite dikes in northern Virginia yielded SHRIMP U-Pb zircon ages of 567±4Ma and 555±4 Ma (J.N. Aleinikoff, USGS, unpub. data, 2009). Collectively, these data suggest that the Catoctin is between 568 and 555 Ma and volcanism lasted at least 13 m.y.

Regional Relations

The Catoctin Formation is confined to the Blue Ridge-South Mountain anticlinorium, where it extends about 350 km from southern Pennsylvania south to near the James River (Spencer, 2000). Metabasalt has not been reported from deep oil-and-gas drill core in the Valley and Ridge or

Appalachian Plateaus provinces to the west. South of the James River in Virginia are several metabasalt beds and epiclastic deposits of basalt in the Unicoi Formation (Henika, 2004). From southwest Virginia to northeastern Tennessee, at least three flows of amygdaloidal metabasalt are interbedded with the clastic rocks of the Unicoi (King and Ferguson, 1960; Rankin, 1993). Whether or not these volcanic rocks in the Unicoi are distal flows of or were derived from erosion of the Catoctin is unknown. There is considerable variation in the thickness and proportion of lithologies within the Catoctin, both along and across strike. Regionally, the rocks thin out to the southwest near the James River and thicken to the southeast near Warrenton. In the Shenandoah National Park, the rocks thin to the west near Stanley. Subaerial metabasalt flows are dominant in the park region. Subaqueous flows consisting of hyaloclasites, breccias, and pillows (Espenshade, 1986; Kline and others, 1990) are restricted to the east limb of the anticlinorium. Marble locally occurs in the formation in northern Virginia (Southworth, Brezinski, and others, 2007). There are three beds of metarhyolite tuff in northern Virginia. The intercalation of metabasalt and phyllite increases northward into Maryland (Fauth, 1977). Metarhyolite is the dominant rock type in the Catoctin Formation in Pennsylvania (Fauth, 1977). From Maryland southward to at least Wintergreen, over a distance of more than 250 km, spotted phyllite and slate often mark the top of the Catoctin Formation. These rocks were assigned to the Loudoun Formation even though they were considered to be volcanogenic (King, 1950; Nickelsen, 1956; Southworth, Brezinski, and others, 2007). Similar ovoids occur in cleaved porphyritic metarhyolite of the Mount Rogers Formation in southern Virginia, Catoctin Formation metarhyolite in Pennsylvania, and volcanic slate of the Sams Creek Formation in Maryland (Southworth, Brezinski, and others, 2007). Similar spotted phyllite beds are not seen in sedimentary phyllites and slates of the Blue Ridge and Piedmont of northern Virginia and southern Maryland.

Metabasalt assigned to the Sams Creek Formation in the western Piedmont of Maryland (Southworth, Brezinski, and others, 2007) may be the equivalent of the Catoctin Formation basalts that flowed offshore into marine waters. Correlative metavolcanic rocks in the northern Appalachians include the Pinney Hollow Formation in Vermont and the Tibbit Hill Formation in southern Quebec. The Pinney Hollow consists of metasedimentary and bimodal metavolcanic rocks; a felsic metavolcanic rock yielded a TIMS U-Pb age of 571±5 Ma (Walsh and Aleinikoff, 1999). Metafelsite interbedded with metabasalt in the Tibbit Hill Formation yielded a U-Pb zircon age of 554±4 Ma (Kumarapeli and others, 1989).

Paleozoic Sedimentary Rocks

Lower Cambrian Chilhowee Group

Introduction

The Lower Cambrian Chilhowee Group consists of the Weverton (Ꞓcw), Harpers (Ꞓch), and Antietam Formations (Ꞓca). These siliciclastic rocks are mostly restricted to the western slope of Blue Ridge, but locally they occur along Skyline Drive near the crest in the southern and northern parts of the map area. They are thrust over Cambrian and Ordovician rocks of the Great Valley section of the Valley and Ridge province in the southwestern part of the map area, from Marksville northward to near Bentonville, and from Front Royal northward to the map boundary. These rocks crop out in several thrust sheets between Bentonville and Front Royal.

From southwest to northeast in the map area, the Chilhowee Group was studied by King (1950), Lukert and Nuckols (1976), Rader and Biggs (1976), Gathright and others (1977, 1978b,c), and Rader and Conley (1995). The stratigraphic divisions by King (1950) were revised by Gathright (1976) and further revised in this study.

Rocks of the Chilhowee Group represent fluvial to shallow-marine transgressive and regressive depositional environments. They exhibit the depositional transition from the continental rift to the development of the passive continental margin, which is marked by the overlying carbonate rocks of the Tomstown Formation. The Chilhowee Group was deposited on what was the coastal plain and shallow-water shelf of the ancient continent Laurentia. The shallow-shelf depositional environments include shoreline beaches, tidal channels, and bays. The sediments were not homogeneous sheetlike deposits, but rather heterogeneous accumulations of lithologies that change both along and across strike. The coarse-grained, siliciclastic rocks are resistant to erosion and tend to underlie numerous linear ridges that extend for several kilometers. The rocks at the base and lower part of the Weverton are variable. Locally, the basal rocks are conglomerates that consist mostly of vein-quartz clasts, but elsewhere the basal rocks are metasiltstones. Many of the metasiltstones show no primary bedding due to intense bioturbation, but some bedded metasiltstones commonly have thick laminae. Similarly, the contacts between the Weverton, Harpers, and Antietam Formations vary along strike with the predominant lithologies. Over most of Shenandoah National Park, the Chilhowee Group consists of upward-coarsening sequences that reflect the transgression and regression of shallow-water depths of the Iapetus Ocean. Almost all conglomerates, sandstones, and quartzites contain features indicative of wave sorting and bioturbation. Fluvial deposits are restricted to a thin interval along the basal contact that is commonly missing due to postdepositional erosion. The variability of the rocks along the basal contact is attributed to differential onlap of the marine deposits on an irregular surface developed on the Catoctin Formation, which was modified by fluvial sedimentation. Some of the variability in the thickness of the formations, as well as their characteristics, was probably due to their location on the continent within the transgression-regression belts.

Weverton Formation

The Weverton Sandstone (Keith, 1894) was revised as the Weverton Formation (Ꞓcw) by King (1950) because of the variety of clastic rocks that are found within it in the Elkton region. Nickelsen (1956) applied three informal members in northern Virginia. Gathright (1976) and Gathright and others (1977) restricted the Weverton Formation to the rocks that King (1950) assigned to the lower member of the Weverton Formation. Gathright and others (1977) suggested that sericite-quartz and chlorite-sericite-quartz phyllite were the most abundant rock types. Basal rocks include purple, maroon, and gray, laminated sandstone (fig. 14*A*), phyllite, 1- to 8-m-thick beds of pebbly quartzite (fig. 14*B* and *C*), laminated metasiltstone (fig. 14*D*), and crossbedded quartzite (fig. 14*E*) (Gathright, 1976). Coarse-grained, sandy and pebbly metasandstone interbedded with silvery-green, quartzose phyllite and reddish-purple, coarse-grained, thick-bedded, ferruginous metasandstone occur in the southwestern area of the park (Gathright and others, 1977). Light-tan to brown, pebbly metasandstone and maroon, ferruginous metasandstone locally mark the upper contact with the predominantly finer grained rocks of the overlying Harpers Formation (Gathright and others, 1977). Brezinski (1992) formally named three ridge-forming members of the Weverton based on exposures in Maryland. The division into members cannot be adopted in Shenandoah National Park because the rocks are not the same as those designated in Maryland and the succession of ridges is not at the same stratigraphic level along strike.

Within the northern part of park, the Weverton consists mostly of upward-coarsening sequences of heavily bioturbated metasiltstone and phyllite with intervals of wavy, thick laminations. These rocks grade into blocky, fine-grained quartzite that is commonly bioturbated with flat and wavy laminations capped by crossbedded, pebbly metasandstone and metaconglomerate. These upward-coarsening sequences vary from about 6 to 18 m thick and several of these may be included in a larger sequence that is 30 to 46 m thick, which is comparable to Brezinski's (1992) sequence of members. Locally along the basal contact, metaconglomerate beds contain vein-quartz pebbles, metabasalt clasts, and red phyllite derived from the underlying Catoctin Formation. The metaconglomerate beds are commonly less than 0.5 m thick or are included in a 3- to 4-m thick, upward-fining succession with trough crossbedding in progressively thinner sets. These rocks are typically capped by a 1- to 3-m-thick tabular bed of quartz-pebble metaconglomerate that has tabular to trough crossbedding with wavy metasandstone interbeds. In contrast, rocks in the same stratigraphic position in northern Virginia and Maryland consist of light, massive, vitreous quartzite interbedded with thin beds of metasiltstone (Brezinski, 1992 Southworth, Brezinski, and others, 2007).

Harpers Formation

The Harpers Shale (Keith, 1894) was revised as the Harpers Formation (Ꞓch) (King, 1950) because of its varying lithology. Within the Shenandoah National Park region, it consists of abundant beds of sandstone, quartzite, and metasiltstone. Rocks assigned to the middle and upper members of the Weverton Formation by King (1950) were reassigned to the Harpers Formation by Gathright (1976) and Gathright and others (1977). The Harpers Formation in the park consists of predominantly greenish- to bluish-gray quartz-chlorite-sericite phyllite and metasiltstone (fig. 15*A*) interbedded with thin, gray metasandstone, quartzite, and meta-arkose.

The majority of the formation consists of interbedded layers of quartzite, metasandstone, and metasiltstone; metasandstone dominates the unit in the area south of Elkton and metasiltstone dominates to the north (King, 1950). The lower part locally consists of green and brown metasiltstone interbedded with fine-grained metasandstone, but dark quartzite occurs at Blackrock and the surrounding region in the southern section of the park. Light quartzite in the upper part contains burrows of the trace fossil *Skolithos linearis* (King, 1950). In the southern part of the park, distinctive marker beds of maroon and dark-blue to black, ferruginous metasandstone (Ꞓchs) were mapped by Gathright and others (1977).

In the northern part of the park, the Harpers Formation is a poorly exposed interval between the ridge-forming units of the Weverton and Antietam Formations. The interval is characterized by upward-coarsening sequences of bioturbated phyllite with intervals of wavy, thick laminations that grade into blocky beds of metasiltstone and fine-grained sandstone that contain abundant burrows. The lower part contains local beds of pebbly metasandstone and metaconglomerate; the upper part contains vitreous quartzite and ferruginous arkosic metasandstone. In contrast to the Weverton and Antietam Formations, metasiltstone and phyllite are included in more than half of each upward-coarsening sequence, each of which vary in thickness from 6 and 24 m. Metasandstone and metaconglomerate beds are generally thin or absent. The Harpers Formation appears to consist of two large-scale, upward-coarsening sequences about 60 m thick. The top of the lower sequence consists of quartzite with interbeds of metasandstone and thin dolomite. The top of the upper sequence is the base of the metasandstone of the Antietam Formation. The trace fossil *Skolithos linearis* appears to be limited to clean quartzite beds in the upper part of the Harpers Formation.

Antietam Formation

The Antietam Sandstone (Keith, 1894) and the Antietam Quartzite (King, 1950) were revised as the Antietam Formation (Ꞓca) (Brezinski, 1992). In the Shenandoah National Park, the Antietam consists of tan to white, thin-bedded metasandstone and quartzite (fig. 15*B*) interbedded with laminated metasiltstone (Gathright and others, 1977). White, massive, crossbedded quartzite (fig. 15*C*) that contains burrows of the trace fossil *Skolithos linearis* (fig. 15*D*) forms ledges and underlies ridges in the park area. The upper part contains brown friable calcareous sandstone that is less resistant and poorly exposed. In the northern part of the park, the quartzite at the base is the cap to an upward-coarsening sequence in the Harpers Formation; the formation includes at least one other upward-coarsening sequence. The quartzite typically contains planar, low-angle crossbeds and wavy interbeds with thin phyllite partings. Beds dominated by *Skolithos linearis* burrows are at the top of the quartzite caps. The base of the higher upward-coarsening sequence is a poorly exposed, 10-m-thick interval containing thin interbeds of phyllite and metasandstone.

Regional Relations

The siliciclastic rocks of the Chilhowee Group occur all along the 885-km-long Blue Ridge from Pennsylvania southward to Georgia. The lithologies and formational names, however, are different both along and across strike. The formations of the group were named by Keith (1894) in northern Virginia and Maryland and then subsequently in two different areas of eastern Tennessee (Keith, 1895). The formational names have been used interchangeably in Virginia and specifically in the Shenandoah National Park region (Gathright, 1976). The Chilhowee Group is a section of interbedded sandstones and shales which was named by Safford (1856) for Chilhowee Mountain, Tenn., adjacent to the Great Smoky Mountains National Park (King and others, 1958; Southworth and others, 2005). In northern Virginia and southern Maryland, the Chilhowee Group consists of the Loudoun, Weverton, Harpers, and Antietam Formations (Keith, 1894; Southworth, Brezinski, and others, 2007). There is no type locality designated for the Loudoun Formation, and the use of the name has long been controversial (Southworth and Brezinski, 1996). The type locality of the Weverton Formation is a collection of exposures on Weverton Cliffs at the southern end of South Mountain along the Potomac River in Maryland. The type locality of the Harpers Formation is a collection of exposures on the cliffs along the Potomac River at Harpers Ferry, W. Va. A collection of exposures along tributaries of Antietam Creek in Maryland is the type locality of the Antietam Formation.

In the Shenandoah National Park, slate and phyllite that locally occur between the metabasalt at the top of the Catoctin Formation and the siliciclastic rocks at the base of the Weverton Formation are herein assigned to the Catoctin Formation; therefore the Loudoun Formation is not recognized in this area. Rocks of the Loudoun Formation occur only from southern Pennsylvania to northern Virginia (Southworth, Brezinski, and others, 2007). The contact between the rocks of the Catoctin Formation and the overlying rocks of the Chilhowee Group has long been recognized as a disconformity (Keith, 1894; Furcron, 1934; King, 1950; Stose and Stose, 1946; Reed, 1955; Nickelsen, 1956; Rodgers, 1970; Gathright, 1976; Southworth and Aleinikoff, 2007; Southworth, Aleinkoff, and others, 2007). Red and blue slate and variegated phyllite that are at the top of the Catoctin Formation are considered to be oxidized, weathered, and locally transported vesicular metabasalt and fossil soil (Reed, 1955; Nickelsen, 1956) and, therefore, are not assigned to the Chilhowee Group. Clasts of metabasalt and red jasper derived from the Catoctin Formation occur locally within the basal conglomeratic rocks of the Weverton Formation, which indicates that, there was a period of subaerial exposure and erosion before deposition of the Weverton.

The name Weverton Formation is applied from southern Pennsylvania southward through the Shenandoah National Park region. South of the map area in eastern Tennessee, coarse conglomeratic rocks at the base of the Chilhowee are assigned to the Unicoi Formation (Rankin, 1993; Virginia Division of Mineral Resources, 1993), and further to the southwest, finer grained strata are called the Cochran Formation (Keith, 1895; King and others, 1958). The name Harpers Formation has been used from southern Pennsylvania south to near Roanoke, Va., but to the southwest into Tennessee, similar fine-grained rocks are called the Hampton Formation (King and others, 1958). Rocks correlative to the

Harpers and Hampton Formations in east-central Tennessee were named the Nichols Shale (Keith, 1895). The name Antietam Formation has been used from southern Pennsylvania southward to near Roanoke, equivalent rocks to the southwest are called the Erwin Formation (Keith, 1895), and those even further to the southwest are called the Nebo Quartzite, Murray Shale, Hesse Quartzite, and Helenmode Formation (Keith, 1895; King and Ferguson, 1960). Of all the rocks in the Chilhowee Group, only the *Skolithos linearis*-bearing, clean, fine-grained sandstone and quartzite (beach sands) of the Antietam (or equivalent Erwin or Hesse) Formation is lithologically consistent across the region.

Paleozoic Sedimentary Rocks of the Great Valley and Page Valley sections of the Valley and Ridge Province

Great Valley and Page Valley

Introduction

The bedrock units of the Great Valley and Page Valley sections of the Valley and Ridge province were compiled from Rader and Gathright (2001a,b). The following descriptions come from the region immediately to the northeast, where the rocks are well exposed (Southworth, Brezinski, and others, 2007). Although the Great Valley and Page Valley are mostly underlain by calcium-carbonate-rich rocks that exhibit well-developed karst topography, the fine-grained, siliciclastic Waynesboro and Martinsburg Formations also are present.

Tomstown Formation

The carbonate rocks of the Lower Cambrian Tomstown Formation (Ꞓt) (Tomstown Dolomite of Stose (1906)) were subdivided into the ascending Bolivar Heights, Fort Duncan, Benevola, and Dargan Members (Brezinski, 1992) to the north; however, these members are not subdivided on the geologic map (plate 1). These rocks are the first evidence that a passive continental margin developed on the rifted continent. The vertical sequence of rocks represents a change from a shallow, carbonate shelf to a deepwater shelf and then to a carbonate bank (Brezinski, 1992). The Bolivar Heights Member is a thin-bedded limestone containing wispy dolomitic burrows that increase in abundance upsection and become bioturbated. Locally, at the base of the formation in Maryland, there is a 15-m-thick interval of mylonitic marble that defines a detachment fault above the underlying Antietam Formation (Brezinski and others, 1996). The Fort Duncan Member is a dark, thick-bedded, burrow-mottled dolostone whose base was an erosional surface (unconformity) on the top of the Bolivar Heights Member. The Benevola Member is a very thick bedded to massive, sugary dolostone with faint crossbedding. The Dargan Member consists of a lower bioturbated dolostone alternating with intervals of laminated dolostone. In the upper part, bioturbated and oolitic dolostone is interbedded with laminated limestone and silty dolostone.

Waynesboro Formation

The Lower Cambrian Waynesboro Formation (Ꞓwa) (Stose, 1906) was subdivided in Maryland into the Red Run, Cavetown, and Chewsville Members (Brezinski, 1992); however, these members are not subdivided on the geologic map (pl. 1). These members have distinctive physiographic expressions that facilitate field mapping; sandstone in the lower part of the Red Run Member and upper part of the Chewsville Member underlies low hills with the intervening swale underlain by less resistant carbonate rock of the Cavetown Member. The Red Run Member consists of interbedded sandstone and laminated, ribbony, sandy dolostone. The Cavetown Member consists of a thick-bedded, massive limestone and bioturbated dolostone; bioturbated dolomitic limestone; dolostone; thin, calcareous sandstone; and shale, which in turn is overlain by thick-bedded, bioturbated dolomite, and laminated ribbony dolomite at the very top. The Chewsville Member consists of dark siltstone, sandstone, and shale. The siltstone commonly exhibits ripple marks and mudcracks and the sandstone is crossbedded and contains *Skolithos linearis* burrows.

Elbrook Limestone

The Upper and Middle Cambrian Elbrook Limestone (Ꞓe) (Stose, 1906) consists of cyclic intervals of limestone, shale, and shaly dolostone in the lower part; thin-bedded to bioturbated limestone in the middle part; and a thick sequence of medium-bedded algal limestone, dolostone, and dolomitic shale in the upper part.

Conococheague Limestone

The Upper Cambrian and Lower Ordovician Conococheague Limestone (OꞒc) consists of interbedded limestone, dolostone, and sandstone arranged in cycles. The lower 100 m consists of coarse-grained, calcareous to dolomitic sandstone; fine-grained limestone with intraformational conglomerate; and fine-grained dolostone that forms low ridges. Above these rocks are cycles of intraformational conglomerates, algal bioherms, ribbon rock, and oolites.

Stonehenge Limestone

The Lower Ordovician Stonehenge Limestone (Os) consists of silty, laminated limestone overlain by thick-bedded, fossiliferous limestone. The upper part consists of algal bioherms, intraformational conglomerates, and bioclastic beds that alternate with thin dolostone beds.

Beekmantown Group

The Middle and Lower Ordovician Beekmantown Group (Ob) consists of the Stonehenge Limestone and its Stoufferstown Member, Rockdale Run Formation, and Pinesburg Station Dolomite, which are not differentiated on the geologic map (plate 1). The basal part of the Stonehenge Limestone (the Stoufferstown Member) consists of silty, laminated limestone, which is overlain by thick-bedded, fossiliferous limestone; algal bioherms, intraformational conglomerates, and bioclastic beds alternate with thin dolostone beds. The Rockdale Run Formation consists of cyclically bedded limestone and dolostone. The limestone is thin to medium bedded, fossiliferous, and contains intraformational conglomerates, algal bioherms, bioclastic zones, mottled burrows, and chert nodules. The dolostone is laminated and contains abundant mudcracks. The Middle Ordovician Pinesburg Station Dolomite consists of medium- to thick-bedded dolostone, and laminated dolostone that contains chert nodules. Crosshatched joints weather to form a diagnostic “butcher-block” pattern on the surface. Thin intervals of fine-grained limestone occur in the lower part. Near the top of the formation, irregularly bedded, brecciated dolostone indicates paleokarst.

Edinburg Formation, Lincolnshire Limestone, and New Market Limestone, Undivided

An unconformity separates rocks of the Beekmantown Group from the overlying Middle Ordovician rocks (Oen). The unconformity is slightly angular and marks the change from deposition along a passive margin to an active margin or convergent continental plate boundary. The New Market

Limestone is a very distinctive medium- and light-gray, micritic, fenestrated limestone. The Lincolnshire Limestone is very dark gray, medium- to coarse-grained, medium-bedded limestone that contains distinctively bedded black chert nodules and beds of bioclastic limestone. The Edinburg Formation consists of dark-gray, interbedded limestone and calcareous shale with characteristic knobby weathering features. The Edinburg contains yellowish-brown volcanic ash or bentonite beds; the beds may be correlative to those in the Chopawamsic Formation, which occurs about 95 to 125 km to the southeast. The metavolcanic rocks of the Chopawamsic Formation yielded isotopic ages of 470±1.4 Ma (Coler and others, 2000) and 451±4 Ma (Horton and others, 1998; J.N. Aleinikoff, unpub. data, 2007).

Martinsburg Formation

The Upper and Middle Ordovician Martinsburg Formation (Om) consists of shale, siltstone, and sandstone with argillaceous limestone at the base. The lower part consists of calcareous shale of the Stickley Run Member of the Martinsburg Formation (Epstein and others, 1995), which transitions upward into turbidite deposits that exhibit typical Bouma cycles with load clasts. The turbidites indicate the foundering of the margin and its evolution into a basin. The margin was sinking at the same time that sea level was rising; therefore, shallow-water carbonate sedimentation gave way to deepwater clastic sediments.

Devonian and Silurian Rocks That Underlie Massanutten Mountain

The rocks that overlie the Martinsburg Formation are resistant to erosion and underlie Massanutten Mountain in the Massanutten synclinorium. The Silurian Massanutten Sandstone (Sm) caps the ridges and consists of light-gray, fine- to coarse-grained, locally conglomeratic, cross-laminated sandstone. These rocks are overlain by a sequence of Silurian and Devonian rocks (McKenzie Limestone through Mahantango Formation) that are mapped undivided (DSu) on the geologic map (plate 1). The Upper and Middle Silurian McKenzie Formation is predominantly gray, calcareous shale. The Upper Silurian Bloomsburg Formation that overlies it consists of red mudstone interbedded with red, ferruginous sandstone and shale, which grades upward into gray limestone and greenish-gray, calcareous siltstone and mudstone of the Upper Silurian Wills Creek Formation. The Upper Silurian Tonoloway Limestone overlies the Wills Creek and consists of gray, laminated limestone with mudcracks. Gray, fossiliferous limestone and gray, laminated limestone containing black, nodular chert and white, blocky chert of the Lower Devonian and Upper Silurian Keyser Limestone grades upward to light-gray, laminated to thick-bedded limestone; black, nodular chert; and white, blocky chert of the Lower Devonian Helderberg Group. Light-gray, fine- to coarse-grained, cross-laminated, calcareous, and fossiliferous sandstone, locally conglomeratic, of the Lower Devonian Ridgeley Sandstone represents beach sand that was deposited during a regression; as the water deepened during a transgression, shale and sandstone were deposited as turbidites. The Middle and Lower Devonian Marcellus Shale consists of dark-gray to black, fissile shale with interbeds of gray, silty limestone and calcareous shale. The Middle and Lower Devonian Tioga Ash Bed consists of gray shale and siltstone; brown, calcareous ash containing biotite; and black, fissile shale. The ash marks the influx of volcanic island-arc material associated with the Acadian orogeny in the northern Appalachians. The Middle and Lower Devonian Needmore Shale consists of greenish-gray, fossiliferous shale and calcareous mudstone with black shale at the base. The Needmore is overlain by gray mudstone, sandstone, and fossiliferous shale of the Middle Devonian Mahantango Formation, which is the youngest formation preserved in the core of the Massanutten synclinorium in Fort Valley.

Jurassic Dikes

Linear, discontinuous, Early Jurassic diabase dikes (Jd) have both olivine-normative and low-titanium, quartz-normative tholeiitic compositions. The dikes have variable trends and were emplaced along dilated fractures in the Blue Ridge and Valley and Ridge provinces. The rocks are readily recognized in the field as massive, hard, subrounded boulders of black rock that have a rusty stain (fig. 16). A diabase dike exposed along the Potomac River in Maryland yielded an $^{40}Ar/^{39}Ar$ age on amphibole of 200 Ma (Kunk and others, 1992). One northeast-trending Late Jurassic peridotite dike (Jpd) that is 182 m long and 46 m wide intruded the Martinsburg Formation west of Front Royal. The poorly exposed, weathered, and altered rock is composed of chlorite, phlogopite, biotite, ankerite, and talc derived from the breakdown of olivine and pyroxene (Young and Bailey, 1955). Several northwest-trending, near-vertical Late Jurassic alkalic dikes (Jad) with compositions of nepheline syenite, teschenite, and teschenite-picrite intruded the Martinsburg Formation west of Grottoes (Johnson and others, 1971). Biotite and hornblende from a nepheline syenite and teschenite dike to the west of the map area yielded K-Ar ages ranging from 157±8 to 148±7 Ma (Zartman and others, 1967; Johnson and others, 1971). The northwest- and northeast-trending Early to Late Jurassic dikes are related to continental extension associated with the opening of the Atlantic Ocean (Southworth and others, 1993).

Structure

Mesoproterozoic

Ten of the older units are gneisses (1,183–1,144 Ma) that crop out in northwest-striking belts in which parallel foliations in the rocks developed under amphibolite- to granulite-facies metamorphic conditions. Mineral assemblages in the foliations include quartz, plagioclase, hornblende, clinopyroxene, and orthopyroxene (Burton and Southworth, 2004; Tollo and others, 2006). The northwest-trending foliations postdate emplacement at 1,175 Ma of leucogranite dikes, which were later folded. Strong foliation in six orthopyroxene-bearing gneisses (1,177–1,158 Ma) is defined by aligned ferromagnesian minerals that are oriented parallel to leucocratic layers; these layers are transposed, intrafolial folds (see fig. 2*C*). The northwest orientation of the rocks and foliations is not the product of Paleozoic deformation because the rocks and foliations were transected by (1) northeast-trending Neoproterozoic metadiabase dikes, (2) pervasive cleavage that developed under greenschist-facies metamorphic conditions in the Paleozoic, and (3) faults and high-strain zones (Southworth, 2005). Three bodies of orthopyroxene-bearing gneisses containing north-northeast-trending Mesoproterozoic foliation are xenoliths within younger unfoliated metagranitoids. The structural orientations of the metagranitoids (Yod, 1,165±7 Ma; Yoq, 1,161±10 Ma) are interpreted to have been rotated during emplacement of the younger host.

Rocks as young as 1,144 Ma show evidence of strong penetrative deformation. The Flint Hill Gneiss (1,144 Ma) is commonly migmatitic (Lukert and others, 1977) and contains mylonitic foliations which are probably Mesoproterozoic in age because the foliations exhibit the folding and kinking with spaced cleavage that is associated with Paleozoic deformation (Clarke, 1984) (fig. 17*A*). Metagranitoids as old as 1,120 Ma display variably developed Mesoproterozoic foliations defined by augen and porphyroclasts of feldspar, but the rocks retained their primary igneous textures. These foliations are subparallel to the orientation of Paleozoic cleavage and commonly are difficult to differentiate in outcrop (Burton and Southworth, 2004). A map-scale xenolith of foliated porphyroclastic metagranitoid (1,078 Ma) within a massive, orthopyroxene-bearing metagranitoid body (about 1,050 Ma) ,indicates that augen-forming foliation locally occurred between about 1,078 and 1,050 Ma (Tollo, Aleinikoff, Borduas, and Hackley, 2004), but not under granulite-facies conditions. Younger metagranitoids (1,063–1,028 Ma) crosscut the foliated older units and most lack an obvious Mesoproterozoic foliation. Two units of similar age (Yor and Ycg, 1,060 Ma) are both massive and weakly foliated. A meter-scale xenolith of foliated biotite monzogranite-quartz monzodiorite (1,032±10 Ma; table 1, sample 14) within unfoliated biotite monzogranite-quartz monzodiorite indicates that some foliations developed at a later time.

The chronology of events defined by Mesoproterozoic igneous rocks correlates with events documented elsewhere in eastern North America. Group 1 rocks crystallized and were structurally deformed at granulite-facies conditions during the late phase of the Shawinigan orogeny, as established in the Grenville province of Canada (Rivers, 2008). Crystallization of Groups 2 and 3, less intense deformation at amphibolite-facies conditions (on the basis of foliated xenoliths dated at about 1,078 and about 1,035 Ma), and the formation of zircon overgrowths (1,020–960 Ma) temporally coincide with the early part of the Ottawan orogenic phase of the Grenvillian orogeny, also established in the Grenville province (Rivers, 2008). The dominant tectonic event affecting northern Blue Ridge rocks appears to be the Shawinigan orogeny, not the Ottawan phase of the Grenvillian orogeny. Because Group 1 rocks occur as large xenoliths within Group 3 rocks, map-scale folds and faults were not recognized. The generalized tectonic map on plate 1 shows that within Group 1 there are abundant foliations that trend westward, and some internal units are parallel to them.

Neoproterozoic

Neoproterozoic extensional tectonics resulted in the emplacement of the anorogenic granitic and volcanic rocks of the Robertson River Igneous Suite. These rocks were emplaced from 730 to 700 Ma (Tollo and Aleinikoff, 1996) in northerward-trending, dike-like sheets during crustal extension (Bailey and Tollo, 1998). Near Madison, the steeply northwest dipping White Oak Run high-strain zone contains mylonitic foliation with kinematic indicators that record Neoproterozoic extension under amphibolite-facies conditions (Bailey and Simpson, 1993; Knight and Bailey, 1999). Regionally, the Robertson River rocks separate distinct groups of Mesoproterozoic rocks (Burton and Southworth, 2004). East of the belt of Neoproterozoic rocks in northern Virginia, the rocks are predominantly Group 1, whereas west of the belt, Group 3 rocks dominate. The Neoproterozoic fault and batholith likely reactivated an existing Mesoproterozoic fault and (or) the boundary of an intrusive rock body.

The Mechum River Formation has long been interpreted to represent the sediments in a Neoproterozoic graben (Nelson, 1962; Mitra and Lukert, 1982). Bailey, Peters, and others (2007) suggest, however, that the rocks were deposited in a west-dipping half graben. They postulate that the listric, southeast-dipping western border fault exists west of the present outcrop belt, but the fault has not been recognized in the field. No northwest-dipping beds were recognized because Paleozoic deformation modified the original attitudes of the beds. Southeast of Castleton, felsic metavolcanic and conglomeratic rocks (Zrbr) associated with the felsite component of the Battle Mountain Alkali Feldspar Granite (Zrbf) were interpreted to be preserved in a Neoproterozoic graben (Tollo and Hutson, 1996; Bailey, Peters, and others, 2007). Locally in the Shenandoah National Park, the contact between basement and cover rocks has abrupt vertical changes that are best attributed to high-angle normal faults. Such small, vertical offsets of the rocks of the Swift Run and Catoctin Formations occur northwest of Big Meadows. Brecciated orthopyroxene monzogranite-quartz monzodiorite (Yom) (fig. 17*B*) in the dome at the head of Weaver Hollow (Elkton East 7.5-minute quadrangle) may be due to Neoproterozoic extension because the contact between the dome and the cover rocks is not cut. Three faults located north of Bucks Elbow Mountain (Browns Cove 7.5-minute quadrangle) were interpreted to be Neoproterozoic normal faults that were reactivated during the Paleozoic (Bailey, Peters, and others, 2007).

Numerous Neoproterozoic metadiabase dikes intruding the Mesoproterozoic rocks provide the best evidence of crustal extension associated with the Neoproterozoic rifting event. The highest density of dikes in the map area is within the Flint Hill 7.5-minute quadrangle, where about 520 dikes with north-to-northeast trends were mapped by Lukert and Nuckols (1976). Generally, there appear to be fewer metadiabase dikes within the map area than northeast of the map area, where the distribution and density increases to about 50 percent of volume along the Potomac River (Southworth, Burton, and others, 2006; Southworth, Brezinski, and others, 2007). In that area, the dikes were transposed into the axial-planar cleavage of the Blue Ridge-South Mountain anticlinorium, so virtually all of them strike northeast and are inclined to the southeast (Southworth, Burton, and others, 2006). In contrast, the dikes within the map area commonly strike and dip in variable directions and often lack a Paleozoic foliation (see fig. 10).

Paleozoic

Folds

Paleozoic deformation produced the regional-scale, Blue Ridge-South Mountain anticlinorium and Massanutten synclinorium. This 275-km-long fold was transported westward along the North Mountain thrust fault during the Alleghanian orogeny (Evans, 1988). $^{40}Ar/^{39}Ar$ and fission-track data suggest that deformation, metamorphism, and uplift associated with the Alleghanian orogeny occurred from the Devonian to the Early Permian (Kunk and Burton, 1999; Naeser and others, 2004; Wooton and others, 2005; Southworth, Drake, and others, 2006). The Blue Ridge-South Mountain anticlinorium is a broad, northeast-trending, asymmetrical, west-verging, and gently north plunging F_1 fold complex with associated axial-planar cleavage (S_1); the Shenandoah National Park is situated on the anticlinorium's west limb. In the highlands of the park, the rocks are nearly horizontal. Locally, a younger, minor folding phase (F_2) and associated crenulation cleavage (S_2) developed near high-strain zones. Where this phase has occurred, the rocks are disharmonically folded. In the Blue Ridge, the layered rocks of the Chilhowee Group have second- and third-order folds that are not recognized in the Mechum River, Swift Run, and Catoctin Formations. The irregularly shaped outliers of the Swift Run near Boonesville (Free Union 7.5-minute quadrangle), north of White Hall (Browns Cove 7.5-minute quadrangle), and near Fletcher are found in the troughs of synforms. The rocks of the Mechum River Formation occupy a footwall syncline; an Alleghanian thrust fault placed Mesoproterozoic and Neoproterozoic plutonic rocks above them (Bailey, Peters, and others, 2007). To the west of Blue Ridge, rocks as young as the Middle Devonian Mahantango Formation (392–386 Ma; Harris and others, 1994) are folded in the Massanutten synclinorium. These rocks are in tight, upright, second-, third-, and fourth-order disharmonic folds that verge up the limbs of the higher order syncline. Axial-planar cleavage in the Great Valley is most evident in the shale of the Martinsburg Formation, but locally the cleavage is developed in the carbonate rocks.

Cleavage

A regional, penetrative, S_1 cleavage (Cloos, 1951; Mitra and Elliott, 1980) overprints the Mesoproterozoic foliation and bedding in the Neoproterozoic to Devonian rocks. The northeast-striking and southeast-dipping slaty cleavage (fig. 18) is axial planar to the Blue Ridge-South Mountain anticlinorium and is coeval with its formation. Spaced cleavage that is axial planar to the Massanutten synclinorium was probably coeval. In the Blue Ridge, in response to late-stage north-south contraction of the rocks, the S_1 cleavage locally was folded and warped so that it now strikes northwestward. The Swift Run Formation at the basement-cover contact shows this late-stage warping. Cleavage in the metabasalt of the Catoctin Formation transects columnar joints and gives the joints a tilted aspect. Local high-strain zones produced a strong pervasive cleavage in some of the metabasalt, rendering the rock a foliated metabasalt. Spaced, discontinuous, S_2 pressure-solution cleavage locally is axial planar to F_2 folds.

Faults

Thrust Faults

Gathright (1976, p. 41–42) noted the following: "Faults occur locally in all sections of the park and have had a strong influence on the location, shape, and development of the Blue Ridge. The presence of a fault is generally marked by linear low areas that are formed on the broken rocks or by the foot of an escarpment where resistant and nonresistant rocks are separated by the fault. Faults are commonly obscured by talus, alluvial gravel deposits, or deep soil and saprolite developed on the intensely fractured rocks along the fault trace. Bedrock exposures of fault surfaces are rare, but generally contain evidence of intense deformation such as brecciated or mylonitized rock."

In addition to these landforms, Paleozoic thrust faults were mapped on the basis of stratigraphic truncation and omission, mylonitic foliation, and localized deformation fabrics. Many faults are actually components of fault zones that likely had a complex history of reactivation. Several areas and the features they exhibit are as follows: (1) the frontal boundary between the Blue Ridge and the Great Valley, (2) high-strain zones in the Blue Ridge basement rocks, and (3) a thrust fault on the east side of the belt of Mechum River Formation in the eastern part of the map area (center of the anticlinorium).

Frontal Blue Ridge Thrust Fault System

Rocks of the Blue Ridge overlie younger rocks in Page Valley along a family of thrust faults that constitute a frontal Blue Ridge fault system. From Pennsylvania southward to Georgia, a similar fault system is recognized; however, it is not continuous and the faults are not as complex as in the Shenandoah National Park area, where these faults cut the Blue Ridge into four discreet structural blocks that have unique geometries. Unlike the many faults that parallel the northeast-striking structural grain, these frontal thrust faults trend obliquely across the structural grain and have dips that vary from shallow (Front Royal fault) to steep (Stanley fault). Individual faults also dip variably from shallow to steep, and some show reverse motion along strike. Most of these faults were active late in the Alleghanian orogeny based on the observations that they truncate folds and are typically display brittle characteristics (King, 1950; Wickham, 1972; Gathright, 1976; Rader and Biggs, 1976). Several of the faults can be traced across the Blue Ridge highlands into high-strain zones that formed under ductile conditions deeper in the crust, suggesting that the erosion level is superimposed on fault blocks that plunge to the southwest; that is, deeper ductile zones occur to the northeast and brittle zones occur to the southwest. The overall pattern and geometry of these faults suggests a complex history of fault reactivation. Older faults were emplaced, folded, and cut by younger faults.

The frontal Blue Ridge thrust fault system consists of an unnamed southwesternmost thrust fault, the Elkton fault, the Stanley fault, and the Front Royal fault. The southwesternmost thrust fault places Antietam Formation over Waynesboro Formation. The Elkton fault tracks northeastward across the Blue Ridge highlands into a high-strain zone near Sperryville (Gathright, 1976). The Stanley fault trends northeastward across the Blue Ridge, placing Neoproterozoic Catoctin Formation over Paleozoic rocks and places Mesoproterozoic rocks over the Catoctin Formation. As compiled from Rader and Gathright

(2001b), the Stanley fault, the Front Royal fault, and an unnamed fault in the Page Valley frame a tectonic window in the Page Valley. Where the Stanley fault crosses Thornton Gap, the fault is steep and may have the geometry of a tear fault with minor dextral strike-slip displacement. The Stanley fault may be folded with the Massanutten synclinorium because it dips westward beneath the anticlinorium. Zircon and apatite fission-track data across the Stanley fault at Thornton Gap suggest that the fault broke rapidly at about 305 Ma (Naeser and others, 2004). Two splays of the Stanley fault cross Thornton Gap (Reed, 1955), where a structural discontinuity in Mesoproterozoic foliation in the megacrystic orthopyroxene syenogranite-monzogranite occurs. One splay likely merges to the northeast into the high-strain zone north of Sperryville. A thrust fault that placed Mesoproterozoic rocks over the Catoctin Formation north of Thornton Gap was cut by two faults that bound a horst between Little Hogback Mountain and South Marshall. The Stanley fault cuts the frontal Blue Ridge fault near Marksville, but the Front Royal fault may cut it south of Rileysville.

The Front Royal fault is part of a family of faults that places Blue Ridge rocks over rocks of the Great Valley. This fault system places Mesoproterozoic to Early Cambrian rocks over rocks as young as the Middle and Lower Ordovician Beekmantown Group. The Front Royal fault truncates older thrust faults in the Great Valley and Blue Ridge and has the geometry of a strike-slip or tear fault where it crosses strike at Manassas Gap. The Front Royal fault most likely was active late during the Alleghanian orogeny. The fault truncates the older thrust faults that placed Chilhowee Group rocks over Tomstown and Waynesboro Formations.

Thrust Faults in the Core of the Blue Ridge-South Mountain Anticlinorium

The rocks in the core of the Blue Ridge-South Mountain anticlinorium are cut by a series of anastomosing high-strain zones and some discreet thrust faults. The 100-km-long belt of Neoproterozoic Mechum River Formation is bounded on the southeast by a steep, southeast-dipping, brittle thrust fault (Bailey, Peters, and others, 2007). The orientation of the linear belt and fault, as well as steep dips of bedding and foliations, are interpreted to be the result of rotation of the rocks, fault, and foliations above a tectonic ramp in the subsurface as the North Mountain thrust sheet was emplaced (Bailey and Simpson, 1993; Bailey, Peters, and others, 2007). Several outliers of Swift Run Formation around Crozet are similar footwall synforms (A. Gattuso, A. Hatmann, F. Knuth, C. Lemon, and C.M. Bailey, College of William and Mary, unpub. data, 2008).

High-Strain Zones

Mesoproterozoic rocks have zones of mylonitic foliation and phyllonite that superficially resemble fine-grained, metasedimentary rocks. These high-strain zones (Bailey and others, 1994, 2002, 2006) are the result of faulting under ductile conditions. The mylonitic foliations contain minerals typical of lower-greenschist-facies metamorphic conditions and they likely were contemporaneous with the development of S_1 cleavage. Within the basement rocks, there are about eight major anastomosing high-strain zones, which form a family of discontinuous, high-strain zones that range from 0.5 to 2 km wide and extend for 20 to 100 km. The southwesternmost high-strain zone is the northeastern continuation of the Rockfish Valley fault zone (Bartholomew, 1977; Gathright and others, 1977). The brittle Elkton fault becomes a ductile high-strain zone near Sperryville, as erosion has exposed a deeper crustal level. All high-strain zones contain asymmetric kinematic indicators that consistently record a top-to-the-northwest sense of shear, which is indicative of thrust faults and contractional deformation (Bailey and others, 2006). One exception is the White Oak Run high-strain zone near Madison, which formed during extension in the Neoproterozoic (Bailey and Simpson, 1993).

Quantitative analysis reveals that the high-strain zones experienced simple shear dominated by flattening strains, volume loss, and triclinic deformation (Bailey and others, 1994). They formed under greenschist-facies conditions at about 350°C to 400°C (Bailey and others, 1994) with a displacement of 1 to 3 km (Berquist and Bailey, 2000). Some high-strain zones occur adjacent to folds in cover rocks, suggesting that the basement accommodated folding by mylonitization.

The best exposure of a high-strain zone in this region is a 125-m-thick strand of the 1- to 2-km wide Quaker Run high-strain zone along Garth Run (Bailey and others, 2006; Fletcher 7.5-minute quadrangle) (fig. 19). This 200-m-long outcrop was exposed during a flood in June 1995. The high-strain zone was localized along the contact between the leucogranite suite and orthopyroxene monzogranite-quartz monzodiorite. The rocks were converted to mylonites and protomylonites with porphyroclasts. Finely layered mylonitic leucogranite and metadiabase occur as tabular to lenticular bodies that are 0.2 to 2 m thick and that are both subparallel to and slightly discordant to the foliation, which strikes from N. 15° W. to N. 10° E., and dips 30° to 50° to the east. A mineral elongation lineation rakes downdip or obliquely at a high angle to the east-northeast. Sheath folds have axes that plunge moderately to the east, parallel to the mineral elongation lineation. Boudins of coarse-grained leucogranite bodies occur within porphyroclastic mylonites. The boudins record elongation both parallel and normal to the east-northeast-raking elongation lineation. Slightly discordant, tabular leucogranite dikes that exhibit boudinage are locally folded. At a few locations, the mylonitic foliation is kinked into narrow (2 to 5 cm) bands. Asymmetric structures such as porphyroclasts, shear bands, and boudins are consistent with triclinic deformation symmetry. Shear strain integrates to a total displacement of about 500 m across the zone. Brittle normal faults with displacements of 0.2 to 2 m cut the mylonitic foliation.

Normal Faults

The bedrock is cut by about 30 mapped northwest-striking high-angle normal faults and cross faults, some of which bound horsts and grabens. Examples are located from Hogback Mountain northward to South Marshall (Bentonville and Chester Gap 7.5-minute quadrangles), near Big Meadows, (Big Meadows 7.5-minute quadrangle), in Harris Cove (Elkton East 7.5-minute quadrangle), and south of Swift Run Gap (Bailey, 2006). Many of these high-angle normal faults are interpreted to be Mesozoic (Hasty and Bailey, 2005; Bailey and others, 2006) because they truncate and displace folded rocks and thrust faults. Several of the northwest-striking faults, however, were truncated by thrust faults, so they must be Paleozoic or older. Examples are the Harris Cove fault (King, 1950; Elkton East 7.5-minute quadrangle), the fault near White Oak Canyon (Old Rag Mountain 7.5-minute quadrangle),, and two cross faults west of Wakefield Manor (Chester Gap 7.5-minute quadrangle).

Metamorphism

Mesoproterozoic

Metamorphism associated with the Grenvillian orogeny is complex and poorly understood. The oldest orthogneisses were foliated and metamorphosed before the emplacement of Group 3 rocks, which occurred from 1,078 to 1,028 Ma. Peak metamorphic conditions near the amphibolite- to granulite-facies transition are indicated by plagioclase, hornblende, quartz, orthopyroxene, and clinopyroxene assemblages, which display granoblastic texture and triple-junction grain boundaries. In addition, blue quartz veins (fig. 20*A*, *B*) containing rutile (a titanium oxide) were emplaced during metamorphism. SHRIMP U-Pb analyses of zircon from rocks with igneous crystallization ages of 1,183±11 to 1,029±9 Ma yielded zircon overgrowth (recrystallization) ages that range in age from 1,180 to 979±11 Ma (Tollo and others, 2006), and three SHRIMP monazite ages that range in age from 1,046±10 to 1,010±4 Ma.

Two $^{40}Ar/^{39}Ar$ hornblende cooling ages (using a closure temperature of about 550° C) of 1,000 and 920 Ma (Kunk and others, 1993) provide the minimum age for granulite-facies regional metamorphism. Some of the zircon overgrowth and monazite ages can be related to heat associated with the crystallization ages of the granitoids. The relation of the timing of plutonism (which lasted from 1,183 to 1,028 Ma), to the timing of metamorphism, and the development of the foliations and the deformation events remains unresolved.

Neoproterozoic

The Rigolet event of the Grenvillian orogeny in Canada (Rivers, 2008) spanned the Mesoproterozoic-Neoproterozoic boundary from 1,010 to 980 Ma. Five orthogneisses in the map area yielded eight zircon overgrowth ages that indicate recrystallization occurred during the Neoproterozoic: 1,040 to 960, 999±9, 997±19, 990±23, 980, 979±11, 960, and 940 Ma (Tollo and others, 2006). SHRIMP analysis of detrital zircons in a quartzite bed in garnet-graphite paragneiss (Zp) near Crossroads (Free Union 7.5-minute quadrangle) suggests that the sedimentary protolith was deposited after 997 Ma. Six detrital zircon grains have been dated from 1,011 to 997 Ma and predepositional metamorphic overgrowths on nine detrital zircon grains are as young as 1,010 Ma (Southworth and others, 2008). These rocks were metamorphosed to amphibolite-facies conditions (garnet, plagioclase, biotite, and quartz) after the 997 Ma deposition. Metamorphosed and foliated garnet-graphite paragneiss along Garth Run (Fletcher 7.5-minute quadrangle) suggests that deposition, metamorphism, and deformation occurred after 812 Ma. These data suggest that there was metamorphism after the Grenvillian orogeny (as currently defined).

Wherever the lavas of the Catoctin Formation flowed over the basement rocks and unconsolidated sands of the Swift Run Formation, the basement rocks were hydrothermally altered by contact metamorphism to form unakite (fig. 20*C*; Wadman and others, 1998; Tollo, Aleinikoff, Borduas, and Hackley, 2004). The granitoids were bleached to a light-gray color, plagioclase was replaced by green epidote-group minerals, and alkali feldspar was replaced by a pink, hematite-bearing assemblage. The effect of the metamorphism diminishes away from fractures and the contact with the overlying metabasalt. These features are well exposed along Skyline Drive, north of Stony Man.

Paleozoic

Paleozoic deformation was accompanied by recrystallization at lower-greenschist-facies conditions. The formation of S_1 cleavage was accompanied by the growth of white, fine-grained mica, chlorite, epidote, actinolite, and locally potassium feldspar. Retrograde mineral assemblages in the Mesoproterozoic gneisses are white mica, biotite, and chlorite that replaced feldspar, brown biotite, and garnet. Neoproterozoic metadiabase dikes and metabasalt contain actinolite, chlorite, and epidote. The rocks in the eastern part of the Page Valley were subjected to the same Paleozoic deformation as the rocks of the westernmost Blue Ridge province, but the metamorphic conditions were less intense and only minor amounts of chlorite are evident. The carbonate rocks in the Page Valley are not metamorphosed, except for some jasperoid deposits in the Conococheague Formation beneath the Front Royal fault (Front Royal 7.5-minute quadrangle).

The precise timing of Paleozoic metamorphism and deformation is poorly constrained but generally is attributed to the late Paleozoic Alleghanian orogeny (Mitra and Elliott, 1980). $^{40}Ar/^{39}Ar$ and fission-track data suggest that deformation, metamorphism, and uplift occurred from the Devonian to the Permian (Kunk and Burton, 1999; Naeser and others, 2004; Wooton and others, 2005; Southworth and others, 2006; Bailey, Kunk, and others, 2007). In northern Virginia, phlogopite in Neoproterozoic marble cooled between 409 and 403 Ma, muscovite cooled between 350 and 300 Ma and grew between

330 and 310 Ma (Kunk and Burton, 1999). Zircon fission-track ages across the Stanley fault are 390 Ma in the footwall and 305 Ma in the hanging wall, which indicates that thrust motion occurred at 305 Ma. Apatite fission-track data indicate the rapid cooling of the hanging-wall rocks during faulting at this time. The rocks of the Blue Ridge and Great Valley were transported westward along the North Mountain thrust fault during the Alleghanian orogeny (Evans, 1988) and regional fission-track ages suggest that the cooling of the rocks in the region between 300 and about 280 Ma (Permian) was associated with emplacement of the regional thrust sheets (Naeser and others, 2004; Southworth, Drake, and others, 2006).

Summary of Geologic History and Tectonics

The bedrock of the Shenandoah National Park region reflects several Wilson cycles (Wilson, 1966) of opening and closing ocean basins. The Grenvillian orogeny resulted in the amalgamation of the ancient supercontinent Rodinia. Rocks of the Blue Ridge province record evidence of this mountain-building episode that lasted about 154 m.y. After about 30 m.y. of magmatism (1,183–1,144 Ma) there was high-grade, strong deformation that produced northwest-trending structures, transposition foliation, and local mylonitic foliation and migmatization. Magmatism and variable deformation continued for about another 30 m.y., followed by an additional 50 m.y. of magmatism with minor deformation. Deformation was locally penetrative but not regional, as indicated by rocks that have variably developed foliations and (or) are massive and nonfoliated. The elongate, north-trending Robertson River batholith (730–700 Ma) intruded a boundary between distinct Mesoproterozoic rocks. Mesoproterozoic orthogneisses and metagranitoids of Group 1 (1,172–1,144 Ma) and Group 2 (1,143–1,122 Ma) occur east of the batholith. Map patterns and structures suggest that Group 2 rocks intruded after deformation, at around 1,144 Ma. West of the batholith, mostly Group 3 anorogenic metagranitoids (1,063–1,028 Ma) intruded Group 1 rocks (1,183–1,144 Ma) and Group 2 rocks (1,120–1,111 Ma) and contain abundant xenoliths of the older groups (Tollo, Aleinikoff, Borduas, and Hackley, 2004; Tollo, Bailey, Borduas, and Aleinikoff, 2004; Tollo and others, 2006). Furthermore, the Group 1 orthogneisses and Group 2 metagranitoids east and west of the batholith are unique to each region. The northwest-trending orientation of structures in the Group 1 rocks reflect the Mesoproterozoic contractional tectonic grain that was directed northeast to southwest and which predated emplacement of Groups 2 and 3. The Robertson River batholith likely intruded into a pre-existing zone of weakness during Neoproterozoic extension (Bailey and Tollo, 1998) and thus marks a fundamental structural boundary within Mesoproterozoic basement rocks of the northern Blue Ridge.

Uplift and erosion of the deep-crust granitoids resulted in sediments locally deposited in the early Neoproterozoic. These sedimentary rocks were buried, metamorphosed, and exhumed before Rodinia began to break apart. From 730 to 700 Ma, plutonic and metavolcanic rocks of the Robertson River Igneous Suite were emplaced during extension. This extension did not result in a successful rift, and sediments of the Mechum River and Swift Run Formations were deposited. From about 568 to 555 Ma, extensional tectonics resumed with volcanism and dike swarms resulting in the separation of the ancient continental plates of Laurentia and Gondwana and the formation of the Iapetus Ocean. The siliciclastic rocks of the Blue Ridge were deposited by early continental and coastal river systems as the Iapetus transgressed and regressed. As the Iapetus Ocean developed, the rocks of the Great Valley were deposited in various marine environments and at varying depths. Evidence of the Ordovician Taconian orogeny was marked by the change in deposition from carbonate rocks to clastic rocks of the Ordovician Martinsburg Formation. The clastic sediments were derived from the uplifted and eroding rocks outboard (east) of the Blue Ridge. Sedimentation and subsidence continued through the Paleozoic.

Metamorphism and deformation of the deeply buried Blue Ridge rocks probably began in the Devonian, when the rocks that now underlie Massanutten Mountain were being deposited in deep water to the west. Metamorphism and deformation continued until the rocks were transported west to their current geographic position along the North Mountain thrust fault (Evans, 1988) during the late Paleozoic Alleghanian orogeny. The resulting Massanutten synclinorium and Blue Ridge-South Mountain anticlinorium are west-verging folds above this thrust fault. Prior to transport, the rocks were situated in the subsurface near the location of Great Falls Park in Virginia, to the north. This orogeny resulted from the collision of Laurentia (North America) and Gondwana (Eurasia) that closed the Iapetus Ocean. The mountainous area created by the converging fold-and-thrust belt collapsed about 60 m.y. later as the continent began to rift again during the Triassic and Jurassic. Diabase, alkalic, and peridotite dikes intruded the bedrock during continued extension.

Drainage systems that had flowed to the west into the Appalachian basin for millions of years were captured and reversed flow toward the lowlands of the Atlantic Ocean. The proto-Potomac River breached the Blue Ridge from the east in the vicinity of the Harpers Ferry water gap during the mid-Oligocene to early Miocene (28–20 Ma; Naeser and others, 2006). Debris shed from the eroding highlands formed thick deposits of unconsolidated sediments along the mountain flanks and over the Piedmont to the east, thus forming the Atlantic Coastal Plain. Modern rivers and their tributaries continue to erode the bedrock in the highlands and carry sediment down the Potomac and James Rivers to the Chesapeake Bay. Sediment collected in streams that drain small basins within the Shenandoah National Park region suggests that the mountains are eroding at a slow geologic rate of 4 to 12 meters per million years (m/m.y.), a rate at which it would take several years to erode material the thickness of a human hair (Duxbury and others, 2007). The rocks and slopes are relatively stable with average erosion rates of 25 to 30 m/m.y. Data from four samples of fluvial sand yield erosion rates that are different depending on the underlying bedrock. Rates are 4.3 m/m.y. for areas underlain by metabasalt, 5.6 m/m.y. for areas underlain by quartzite, 11.9 m/m.y. for areas underlain by siliciclastic rocks, and 13.8 m/m.y. for areas underlain by granite (Duxbury and others, 2007). These data suggest that (1) different lithologies control basin-scale erosion rates, (2) grain size has little or no effect on *10*Be concentration, (3) cosmogenically determined erosion rates in the park area are similar to or lower than those in the southern Blue Ridge (Matmon, Bierman, Larsen, Southworth, Pavich, and Caffee, 2003; Matmon, Bierman, Larsen, Southworth, Pavich, and others, 2003), (4) short-term cosmogenic erosion rates (10^4 yrs) are consistent with long-term rates of 11 to 18 m/m.y. estimated using U/Th/He techniques across the Blue Ridge escarpment to the south (Spotila and others, 2004), and (5) the rates are consistent with rates of 20 m/m.y. determined by regional fission-track analysis (Naeser and others, 2004; Duxbury and others, 2007).

Using an average effective uplift and denudation rate of about 25 m/m.y., about 6.5 km of rock was removed since the Alleghanian orogeny ended to form the current surface. Since the capture of westward drainage, about 600 m of rock has been removed, which is close to the present topographic relief of the park. Thus, it is likely that the highlands were twice as high as they are today.

Mines and Prospects

Iron and Manganese

From the late 1800s to about 1960, iron and manganese oxides were locally mined in and around Shenandoah National Park. Limonite and goethite were mined as early as 1836 and manganese ore and manganiferous oxides (such as psilomelane, pyrolusite, and manganite) were mined as early as 1884. To assess the availability of resources for making steel during World War I, Stose and others (1919) performed a comprehensive study of the manganese deposits that occur along the west slope of the Blue Ridge in Virginia; about 28 active or abandoned mines and prospects were described in the map area. King (1950) further studied the manganese deposits in the Elkton area to support steel production for World War II. He described spherical, grapelike, botryoidal nodules of manganese (fig. 20*D*), 1.3 to 20 cm in diameter, that were embedded in yellow, brown, red, and black clay residuum derived from the carbonate rocks of the Tomstown Formation. The manganese contained too much silica and phosphorous to be suitable for steel production. Exploration trenches, open-cut mines, and shafts mark the location of more than 40 abandoned mines in or just west of the park. Many open-cut mines and prospects are filled with water and resemble ponds and (or) water-filled sinkholes. Remnants of the Crimora Mine, the largest and most productive open-cut manganese mine, can be seen from Crimora Lake Overlook on Skyline Drive.

Manganese occurs in several settings, but almost always it is found at the contact between the metasandstone of the Antietam Formation and the carbonate rocks of the Tomstown Formation. Where the metasandstone dips westward into Page Valley, the ore is beneath the alluvial fans. In areas such as the Crimora Mine, the ore was concentrated in synclines above the metasandstone. The origin of these deposits was proposed by Stose and others (1919), King (1950), and Hack (1965). The manganese minerals originally were primary iron-carbonate minerals that lithified in the dolostone of the Tomstown Formation. The iron-carbonate minerals were later dissolved by bicarbonate in circulating groundwater. Before or during this process, the weathering of calcareous shale and carbonate rocks resulted in the production of clay residuum. Surface and groundwater percolated through the strata and manganese minerals were redeposited as oxides and replaced the clay where the manganese-rich waters mixed with oxygen-rich waters. Hack (1965) recognized that the west slope provided a mechanical and chemical trap for the manganese deposits. The siliciclastic gravels formed the mechanical trap that prevented their erosion, and the chemical trap was the dissolution of the underlying carbonate bedrock that effectively lowered the altitude of the deposits with time. He also noted that the manganese deposits were not restricted to a stratigraphic horizon related to an erosion event or landform, as the process had been active throughout geologic time. The deposits are found on slopes, elevated terraces, and in the subsurface about 30 m below modern stream levels and are forming presently in drainages (Hack, 1965).

Manganese in Breccia

Manganese also forms the matrix and cement for breccias consisting of angular metasandstone clasts (fig. 20*D*), as is seen near the Compton Mine along Dry Run (Luray 7.5-minute quadrangle). The manganese-cemented breccias have been interpreted as cataclastic rocks associated with brittle faults (Rader and Biggs, 1976). However, these breccias probably resulted from several processes. Dissolution of carbonate rocks adjacent to inclined metasandstone may have resulted in gravitational collapse (“caving in”) (Stose and others, 1919), thus producing what is referred to as “collapse breccias” (Gathright and others, 1977). Similar breccias cemented with silica and stained with hematite occur in two places: (1) on the slope above undeformed metasandstone along the railroad cut above Interstate Route 66 (Front Royal 7.5-minute quadrangle), and (2) in the flood plain and slope underlain by Antietam Formation, well above the Front Royal fault and along Dry Run southeast of Compton (Bentonville 7.5-minute quadrangle). The metasandstone closest to the faults is not brecciated, thereby indicating that these breccias may have originated as silicified colluvium. In either of these cases, the breccias were cemented by precipitates from groundwater. Although these breccias currently crop out, they likely formed in the subsurface when there was significant overburden and have subsequently been exhumed by erosion.

Copper

Copper was mined at several places in or near the Shenandoah National Park from the 1850s to the 1940s. Small amounts of copper oxides, sulfides, and native copper (fig. 20*F*) were found in the metabasalt of the Catoctin Formation here and throughout the Blue Ridge of Virginia (Brophy, 1960; Allen, 1963; Allen, 1967; Rader and Biggs, 1975; Lukert and Nuckols, 1976; Bartholomew, 1977; Rader and Conley, 1995). Native copper, covelite, malachite, azurite, cuprite, chalcopyrite, chalcocite, and bornite were found in irregular masses of altered and sheared metabasalt that exhibit epidotization, as well as in quartz veins along joints and cleavage. Copper mineralization in the lower part of the Catoctin may have been the result of primary accumulations of base metals concentrated in fluids during metamorphism and deformation (Allen, 1963). From north to south, mines and prospects are located (1) on Dickey Ridge (Chester Gap 7.5-minute quadrangle), (2) on Stony Man (Old Rag Mountain 7.5-minute quadrangle), (3) between Dark Hollow Falls and Rose River Falls, (4) northwest of Ida (Big Meadows 7.5-minute quadrangle), (5) west of Fletcher, and (6) on Hightop (Swift Run Gap 7.5-minute quadrangle).

The Stony Man prospect is situated at about 1,158 m on the east side of the crest of Stony Man. The copper ore was processed at Furnace Springs to the south near Skyland. The Rose River prospect is north of the Rose River Loop Trail. Concrete foundations, several adits, and as many as eight pits that are several meters across are aligned west-northwest. Small prospects are at an elevation of 674 m on the Matthews Arm Trail (Bentonville 7.5-minute quadrangle) and along Virginia Route 662 east of Rileyville.

Several copper prospects are located in the Mesoproterozoic gneissic rocks (1) southeast of Compton (Thornton Gap 7.5-minute quadrangle), (2) southeast of the Pinnacle Overlook (Old Rag Mountain 7.5-minute quadrangle), (3) on the northwest side of Catlett Mountain along the Catlett Mountain Trail (Old Rag Mountain 7.5-minute quadrangle), and (4) west of Dickey Ridge (Chester Gap 7.5-minute quadrangle). On Dickey Ridge, a lens-shaped mineralized zone containing copper, gold, and silver is located in fractured Mesoproterozoic rocks above the Front Royal fault (Chester Gap 7.5-minute quadrangle; Rader and Biggs, 1975).

Unconsolidated Surficial Materials and Landforms

Introduction

Deposits of unconsolidated surficial materials and the resulting landforms document the most recent part of the geologic history of the Shenandoah National Park region. The surficial materials contribute to the different types of soil, associated flora and fauna, and land use. Some of the landforms and deposits reflect millions of years of processes under varied climatic conditions. The Blue Ridge highlands were once more laterally extensive and higher in altitude; the same processes that shaped them will eventually erode them. The surficial units portrayed on the map resulted from at least four main agents and processes: (1) running water (alluvium, terrace deposits, alluvial-fan deposits, and alluvial-plain deposits), (2) chemical weathering (sinkholes and residuum), (3) gravity and high rainfall events on slopes (debris-flow and debris-fan deposits), and (4) gravity and freeze-thaw processes on slopes (stratified slope deposits, colluvium, and block-field deposits). With the exception of bedrock outcrops, most of the land is covered with surficial deposits, including residuum derived from the weathering of the underlying bedrock, and transported material that is different than the underlying bedrock. In general, (1) colluvium is concentrated in the highlands, (2) debris-fan deposits are concentrated in coves and hollows on the upper to lower slopes, (3) alluvial-fan and alluvial-plain deposits are located on the lower slopes and valleys on the west and east sides of Blue Ridge, respectively, and (4) terrace deposits are located along major rivers in the valleys. Each type of deposit and resulting landform occurs as a function of the underlying bedrock; for example, debris-fan deposits are abundant in areas underlain by gneisses and metabasalt. Block-field deposits are most abundant in the areas underlain by metasandstone and quartzite. The carbonate bedrock of the Great Valley is mostly covered by extensive alluvial-fan deposits that consist of gravel derived from metasandstone and the South Fork of the Shenandoah River has incised the alluvial fans to form terraces and expose bedrock. In addition, sinkholes have developed in the carbonate bedrock, including the bedrock that is buried by the alluvial fans. In contrast, the lowlands east of Blue Ridge are characterized by broad alluvial plains and some terraces that are mantled with gravel.

Fluvial Deposits and Landforms

Introduction

There are four distinct types of fluvial deposits and their associated landforms in the Shenandoah National Park region: (l) alluvial fans west of Blue Ridge, (2) alluvial plains east of Blue Ridge, (3) terraces along major rivers, especially the South Fork of the Shenandoah River in the Great Valley, and (4) alluvium that is ubiquitous along all streams.

Alluvial-Fan Deposits

Alluvial-fan deposits (Nf) (Gathright, 1976; Kochel and Johnson, 1984; Bell, 1986; Simmons, 1988; Kochel, 1990; Duffy, 1991; Kite, 1992; Whittecar and Duffy, 2000; Mills, 2000a,b) are the dominant deposits and landforms west of the Blue Ridge on the east side of the Page Valley. They have also been called valley-floor terraces (Stose and others, 1919), Pleistocene gravels and associated fanglomerates on terrace surfaces (King, 1950), alluvium in piedmont aprons (Hack, 1965), and fluvial-fan deposits on the west flank (Morgan and others, 2004). Along the western foot of the Blue Ridge from Waynesboro north to Luray, thick coalescing alluvial fans cover the bedrock for distances as much as 13 km to the South Fork of the Shenandoah River. Between the towns of Shenandoah and Alma, and from Rileyville north to Front Royal, the alluvial fans have been incised by streams (fig. 21*A*) and the river and are discontinuous or absent.

The alluvial fans have been well studied because they host manganese and iron ores and are a source of aggregate and abundant groundwater; the fans have been referred to as the Elkton aquifer or western toe aquifer of the Blue Ridge Mountains (Hollyday and others, 1997; Swain and others, 2004; Yager and others, 2008). The fans serve as a sponge for groundwater because surface water infiltrates the gravel and chemically weathers the underlying carbonate bedrock. These fans host the only aquifer of unconsolidated sediments west of the Atlantic Coastal Plain province in the Appalachian region. Sinkholes and water-filled depressions in these deposits show that dissolution of carbonate bedrock in the subsurface is active. The original bedrock surface was likely irregular with karst developed within it; therefore, the thickness of the gravel deposits is highly variable. Drill-hole data suggest thicknesses of 36, 51, 111, and 152 m (Gathright and others, 1977). The deposits overlie clay residuum derived from the Tomstown Formation that is as much as 61 m thick (King, 1950).

The alluvial fans have distinct topographic levels, which can vary as much as 61 m in elevation difference (Elkton East 7.5-minute quadrangle; King, 1950). Modern streams and rivers are currently incising the fans. The alluvial fans are a complex composite of deposits that have accumulated over an extended period of time. An exposure of an alluvial fan along Meadow Run near Crimora (Sherwood and others, 1987; Whittecar, 1992) consists of at least 13 discrete units. The basal units are saprolitized, enriched in clay, and highly oxidized. Two overlying units were deformed as the result of sinkhole collapse. Cobbles and gravels are overlain by fine-grained and sandy overbank deposits which were incised by three channels filled with coarse cobbles. Rounded boulders and cobbles in the higher level alluvial fans rarely exceed 0.5 m diameter and are similar to those moved by modern streams that have incised the fans.

The regional distribution of the alluvial fans on the west flank suggests that they are best preserved from Vesuvius north to Luray. They have been eroded by the St. Marys River south of Vesuvius and removed by the Shenandoah River and Potomac tributaries north of Luray. North of the Potomac River, they are preserved in Maryland and Pennsylvania (Grote, 2006). Fission-track data from detrital zircons from the subsurface in the Coastal Plain east of Fredericksburg suggest that the Potomac River drainage breached the Blue Ridge during the Miocene (Naeser and others, 2006). East of the Blue Ridge, terrace deposits in the valleys of the Potomac and James Rivers and preserved Coastal Plain deposits near Tysons Corner in northern Virginia and near Bon Air south of Richmond consist exclusively of cobbles of *Skolithos linearis*-bearing metasandstone and quartzite; these deposits have long been interpreted as Miocene (Weems and Edwards, 2007). The alluvial fans on the west flank of the Blue Ridge were likely eroded, transported, and redeposited as terraces and Coastal Plain sediment when the eastward-draining rivers breached the Blue Ridge; therefore, they may predate the Miocene.

Alluvial-Plain Deposits

Alluvial plain deposits (Np) are manifested as broad flat areas east of the Blue Ridge highlands. Similar deposits or landforms in this area were called terrace and alluvial-fan deposits (Gathright and others, 1977), and high- and low-strath fluvial terraces (Eaton and others, 2001; 2003a,b; Morgan and others, 2004). The alluvial plain deposits are a maximum of 9 m thick, which is thinner than the alluvial-fan deposits west of Blue Ridge that range in thickness from 36 to 152 m based on drill-hole measurements (Gathright and others, 1977). These broad alluvial plains are found where there are no modern rivers. From north to south, these areas are (1) from the Rappahannock River to Sperryville, (2) north of Revercombs Corner, (3) around Peola Mills, (4) between Stanardsville and Dyke, and (5) from Browns Cove to Afton. These low areas probably were part of a larger alluvial plain before incision of modern rivers and dissection lead to their current extents. The terraces show little relation to the present drainage patterns and are thought to be relict landforms resulting from earlier drainage systems. The alluvial plains are as much as 21 m above the modern flood plain (Swift Run 7.5-minute quadrangle) and locally they have only a thin veneer of weathered alluvium that overlies saprolite. Rounded cobbles of vein quartz, granite, and metabasalt locally are on the surface, but most have been stripped of material. Large flat areas between Washington and Sperryville (Washington 7.5-minute quadrangle) and north of Flint Hill may be alluvial plains stripped of material, but they are not shown on the map.

Terrace Deposits

Terrace deposits (Nt) are shown as flat benches that resulted when bedrock was cut into by rivers. Terraces east of Blue Ridge are commonly as much as 35 m above modern river levels (fig. 21*B*) (Thornton River, Washington 7.5-minute quadrangle); terraces west of Blue Ridge are commonly 46 m above modern river levels (South Fork of the Shenandoah River, Bentonville 7.5-minute quadrangle). Much of the unconsolidated material has been eroded, leaving just boulders and cobbles, but some terraces preserve silt, sand, and gravel. Most of the cobbles consist of resistant quartzite, metasandstone, and vein quartz; locally terrace deposits contain chert in the Great Valley and metabasalt east of Blue Ridge.

Terraces are best developed along the South Fork of the Shenandoah River in the Great Valley, where they locally were incised into alluvial fans and shale of the Martinsburg Formation. These terraces formed as the meandering river incised the present flood plain, either as broad paired terraces on both sides of the river, or as sloping terraces restricted to one side. For example, a large abandoned meander of the South Fork of the Shenandoah River is located south of U.S. Route 211, east of New Market Gap on Massanutten Mountain (Hamburg 7.5-minute quadrangle). East of Blue Ridge, terraces are preserved (from north to south) along the Thornton, Hazel, Rapidan, and South Rivers, and the South Fork of the Rivanna River. Lower terraces are inset as much as 12 m above the modern flood plain, but they are not depicted on the map.

The fluvial terraces at the confluence of Kinsey Run and the Rapidan River (Madison 7.5-minute quadrangle) occur at about 15 m above the drainage and were mappable for about 160 km downstream to the Fall Zone at Fredericksburg (Dunford-Jackson, 1978). Because the deposits along the Rapidan River have weathering characteristics (Eaton and others, 2001) similar to deposits of the Fall Zone and Coastal Plain (Dunford-Jackson, 1978), the age of the deposits in this area was suggested by Howard and others (1993) to be late Pliocene and early Pleistocene. For comparison, the only data about such features were derived from studies along the New River, southwest of the map area. There, cosmogenic ^{10}Be exposure dating of alluvial terraces along the New River in the Great Valley of southwest Virginia suggests that fill-cut and strath terraces at elevations 50, 20, and 10 m above the modern river yield model exposure ages of 950,000 to 600,000 years before present (yr B.P.), about 600,000 yr B.P., and 130,000 yr B.P., respectively (Ward and others, 2005). These exposure ages yield a long-term average incision rate of 43 m/m.y., which is comparable to rates measured elsewhere in the Appalachians. In eastern Tennessee, nine discontinuous terraces as much as 30 m above the Little Tennessee River yielded ^{14}C ages of about 31,000 yr B.P. for the highest terrace and 15,000 to 7,000 yr B.P. for the lower terrace (Delcourt, 1980). Delcourt (1980) suggested that the terraces formed as material destabilized from upland surfaces during the transition from cold-phase maxima to interglacial or interstadial conditions during the late Pleistocene. Frost-bounded debris provided large sediment loads, the remnants of which are preserved in the terraces.

Alluvium

Well to poorly sorted alluvium (Qa) is composed of interlayered and stratified, unconsolidated clay, silt, sand, gravel, and cobbles as much as 6 m thick. Alluvium is the basic unit composing the flood plains of all the rivers and tributaries. Channels are commonly incised into bedrock and alluvium is exposed along the banks. The thickness of the alluvium is highly variable and is a function of bedrock, topography, and land-use practices. Locally, thick deposits of alluvium resulted from erosion due to deforestation and agricultural practices of the 1800s, as well as sediment built up behind mill dams. Broad alluvial valleys east of Blue Ridge are only inundated by modern floods. The particle size of alluvium is a function of the parent bedrock and transport distance. Silica-rich bedrock such as quartzite and leucogranite produces abundant boulders and cobbles, whereas finer grained metabasalt and shale produces mostly silt and clay. Alluvium is ubiquitous on the map but the volume is not great. A characteristic feature of the creeks, streams, and rivers that drain the Blue Ridge highlands is the abundance of large boulders of metabasalt and gneiss that are as much as 10 m in diameter. These boulders are lag deposits from debris fans that were incised and that had fine material removed.

Slope Deposits

Introduction

There are five general types of slope deposits: (1) broad, gently sloping debris-fan deposits on the lower slopes; (2) stratified slope deposits; (3) colluvium in hollows on slopes below bedrock escarpments, on side slopes above valley bottoms, and in depressions on the upper slopes of the highlands; (4) block-field deposits of quartzite and metasandstone locally developed on the ridges underlain by the Chilhowee Group; and (5) recent debris-flow deposits that resulted from intense rainfall events.

Colluvium, debris-fan deposits, terrace deposits, and alluvium reflect a continuum of downslope processes. They are differentiated from one another by their altitude, position on slopes, particle size, and distance of transport. Colluvium occurs high on the slopes as talus, block fields, and boulder streams derived from the weathering of resistant rocks such as metasandstone, quartzite, and orthopyroxene-bearing gneisses. On lower slopes, colluvium was transported by creep, by solifluction, and as debris flows during high rainfall events to form debris fans. Debris fans were incised by modern streams that eroded the fine material, leaving coarse boulders and terraces and locally exposing bedrock.

Debris-Fan Deposits

Fan- and irregularly shaped, sheetlike accumulations of rock debris are the dominant Cenozoic deposits in the unglaciated highlands of the Appalachians (Mills, 2000a,b) and they are prominent in the Shenandoah National Park area. Debris-fan deposits (Nd) consist of very poorly sorted boulders and cobbles in a fine-grained matrix of sand, silt, and clay. They are characterized by a bouldery surface (fig. 22). Natural exposures from floods and debris flows show that, in the subsurface, the boulders and cobbles are supported by fine-grained material, and the deposits therefore are classified as a matrix-supported diamicton.

The size, shape, and topographic setting of the debris fans are distinct throughout the area and are a function of the dominant source rock. For example, debris fans derived from orthopyroxene-bearing gneisses near Browntown (Chester Gap 7.5-minute quadrangle) differ from those derived from the Old Rag Granite, which in turn differ from debris fans derived from Catoctin Formation near Big Meadows or debris fans derived from metasandstone of the Chilhowee Group near Paine Run. The debris fans are landforms that accumulated and were modified over an extended period of time. They occupy coves and hollows that are protected from erosion. The resistant boulders form an armorlike layer over the fan surface. The boulders are relict deposits that are undergoing chemical weathering and incision by streams along the borders of the debris fans. The sheet flow of the surface water removed the fine-grained matrix material at the surface and left the boulders behind. The forested areas were cleared by early settlers who, in the process of developing land for crops or pasture, created conical piles of boulders and cobbles or constructed terraces and fence lines using the boulders. Lateral migration of streams and stream capture are the dominant secondary processes (Mills, 2000a,b).

The debris fans were mainly the result of debris flows and floods during glacial and interglacial transitions as warm and cold cycles fluctuated and storms were common (Litwin and others, 2001). Colluvium that accumulated on the upper slopes during cold periods was transported downslope by debris flows that occurred during warm periods of increased precipitation (Mills, 2000b). The fans occupy hollows, coves, and valleys, suggesting that a pre-existing basin had been carved by fluvial erosion. The deposits form extensive, broad, convex-upward fans and aerially extensive sheets that have been modified by erosion. Individual fans consist of multiple surfaces relating to numerous events. The debris fan along Kinsey Run consists of at least four surfaces of different elevations (Fletcher 7.5-

minute quadrangle; Eaton and others, 2003a,b). The thickness of each layer is highly variable. The oldest deposits contain few clasts in a reddish-orange, clay-rich matrix. Clasts of metabasalt have weathering rinds that are 2 to 4 mm thick (Eaton and others, 2001) and granitoid cobbles are saprolitized. The younger debris-fan deposits contain clasts of fresh rock. Some debris-fan deposits have been incised without reaching bedrock and are thicker than 20 m. Boulders in channels on the debris fans subsequently were topographically inverted by gully gravure (a process whereby repeated erosional events cause slopes to retreat) to form a convex-upward pile of boulders (Mills, 1981), some of which are thick as 18 m.

Isotopic Ages of Debris Fans

Debris fans in the Shenandoah National Park area have formed since at least the late Pleistocene. Erosion and incision have been the dominant processes during the Holocene (Whittecar and Duffy, 2000; Eaton and others, 2003b). The debris fans along Kinsey Run that were incised by the high rainfall event in 1995 revealed a stratigraphy of debris flows, stratified slope deposits, colluvium, and alluvium (Fletcher 7.5-minute quadrangle; Eaton and others, 2003a,b). Thirty nine samples of organic material found in debris fan deposits around the park yielded *14*C ages ranging from about 51,000 to 2,000 yr B.P. (Eaton and others, 2003a). Although some of the organic material may have been recycled, the range in ages suggests that debris flows have mostly occurred during the last 25,000 years and that they have recurred an average of every 2,500 years since the onset of the Wisconsin glacial maximum (about 12,000 yr B.P.) (Eaton and others, 2003a,b); however, cosmogenic ^{10}Be and ^{26}Al measured from the top surface of 21 exposed boulders near Kinsey Run (Madison 7.5-minute quadrangle) suggest that the fan surfaces on which the boulders sit are older than 500,000 yr B.P. (Bierman and others, 2002). To the south in North Carolina, debris fans are considerably older; 10 organic samples from 5 debris fans yielded *14*C ages ranging from 25,000 to 1,000 yr B.P., with 18,000 to 16,000 yr B.P. being the most consistent age range (Kochel, 1990). Paleomagnetic reversal of iron oxides from debris fans in western North Carolina suggests that the fans are no younger than 1,000,000 yr B.P. (Mills and Allison, 1995a,b); cosmogenic ^{10}Be and ^{26}Al in quartz support an age of 1.45±0.17 Ma (Mills and Granger, 2002), which is consistent with a cosmogenic age of 1.5±0.3 Ma from the southern advance of the ice sheet along the Ohio River (Mills and Granger, 2002).

Stratified Slope Deposits

There are several noteworthy stratified slope deposits (fig. 23) exposed by the 1995 high rainfall event; similar deposits are likely on many of the slopes in the region. The specific sites that have been studied are shown by a map symbol on plate 1. Stratified slope deposits are rhythmically layered mixtures of clay to cobbles that were deposited parallel to steep slopes where there is no obvious drainage pattern (DeWolf, 1988; Gardner and others, 1991). At 366 m altitude on the east slope along Kinsey Run (fig. 23*A*), a 6.5-m-thick, wedge-shaped, fine-grained, well-stratified deposit yielded radiocarbon ages between 24,570 and 15,800 yr B.P., which is during the late Wisconsin glaciation (Eaton and others, 2003a,b). A gravel pit at about 1,000 m altitude near the top of the mountain east of Big Meadows exposes 3 m of stratified sediment (Smoot, 2004). Along the channel of the Rapidan River at about 700 m altitude, a 5-m-thick section was exposed for 40 m beneath a deposit of boulders in the floor of a hollow (fig. 23*B*) (Big Meadows 7.5-minute quadrangle; Smoot, 2004). The stratigraphy and structure suggest that the deposit accumulated as a combination of solifluction and sheetwash. The deposits at higher altitudes formed by slow creep associated with freeze-thaw cycles. The deposits at lower altitudes were strongly influenced by sheetflow processes (fig. 23*C*) (Smoot, 2004).

Colluvium

Colluvium (Qc) in the Shenandoah National Park area mantles most of the steep slopes and forms aprons (fig. 24*A*) along the west slopes of the ridges underlain by the Chilhowee Group adjacent to the Great Valley. In the highlands, colluvium consists mostly of subangular clasts and boulders of metabasalt and granitoids locally derived from the bedrock (fig. 24*B*). Along the west slopes, the colluvium consists primarily of subangular cobbles of metasandstone and quartzite derived from the outcrop slope of Antietam Formation. The blocky deposits have no matrix at the surface and are not vegetated (Hack, 1965; Gathright, 1976). The colluvium includes mostly talus, block fields, and boulder streams; block fields of quartzite are mapped as a separate unit. Talus is actively forming below steep bedrock cliffs and escarpments. Downslope from the talus are older sheetlike accumulations of colluvium called "block fields." Boulder streams, another type of colluvial deposit, are concentrations of colluvium in valleys that extend downslope. Near Blackrock Summit in the McGaheysville 7.5-minute quadrangle, the long axes of quartzite boulders parallel to the slope indicate that the boulder streams formed as the result of periglacial processes (Felker and others, 2004). Boulders are typically 1 to 3 m long but can be as much as 15 m long. Subsurface drainage and intermittent streams have further modified colluvium. Individual boulders may be covered with moss and (or) lichen (Gathright, 1976), suggesting no transport in the last several decades. Bent and leaning trees locally suggest minor movement by gravity (Hack, 1965). Near escarpments, and especially near waterfalls, blocks of rock have detached from cliffs along bedding, cleavage, and joints, which has resulted in the downslope piles of jumbled blocks of fresh rock.

Massive and resistant orthopyroxene-bearing gneisses weather to large rounded boulders and form colluvium in the highlands. Historical photographs during tunnel construction in 1935 south of Thornton Gap near Marys Rock show matrix-supported diamictite above the portal (fig. 24*C*) where the low stone wall was built by the Civilian Conservation Corps. Massive leucogranite of the Old Rag Granite also weathers to form large, rounded blocks of colluvium on Old Rag Mountain as the result of exfoliation (Gathright, 1976; Hackley, 1999, 2000). Colluvium consisting of metabasalt from the Catoctin Formation is abundant adjacent to ledges in places like South Marshall and North Marshall (Chester Gap 7.5-minute quadrangle) and consists mostly of slab-shaped blocks that developed along foliation planes.

Colluvium formed by gravity, solifluction, freeze-thaw cycles, ice wedging, frost heaving, ice rafting, and tree throw (a process by which an uprooted tree causes soil and rock to be moved) (Hack, 1965; Clark, 1987). Although locally talus is forming at present, the majority of the colluvium is a relict deposit from a Pleistocene periglacial cold climate. Vertically oriented slabs are prominent where there is frost heaving (Washburn, 1985), as seen near the upper reaches of the Rapidan River (Eaton and others, 2001). During cold phases in the Pleistocene, the Blue Ridge highlands were probably above the tree line (Litwin and others, 2003). Different landforms developed on northwest- and southeast-facing slopes because of periglacial conditions as well (Richter, 1973); for example, from near Big Meadows north to Dickey Ridge, colluvium is extensive in the northwest-facing basins and hollows, such as near East Hawksbill Creek (Big Meadows 7.5-minute quadrangle) and Gooney Run (Chester Gap 7.5-minute quadrangle). Erosion oversteepened the bedrock highwall of gneiss and metabasalt and gravity, solifluction, and debris flows transported the colluvium downslope. These hollows accumulate and retain snow and ice into the spring.

Quartzite Block-Field Deposits

Quartzite block-field deposits (Qb) are well developed on the ridges underlain by the Chilhowee Group in the southwestern part of the Shenandoah National Park area (fig. 24*D*, *E*) (McGaheysville 7.5-

minute quadrangle; Hack, 1965; Gathright, 1976; Eaton and others, 2002). They are more extensive on dry, south-facing dip slopes. Because the surface of the block fields was too rocky for use by man, the fields retained their undisturbed, pristine state despite human development in the surrounding areas. The quartzite bedrock mechanically weathered along bedding and fractures to form blocks of uniform shapes and sizes as much as 1.5 m long (Eaton and others, 2002). Using aerial photographs, Hack (1965) identified hundreds of unvegetated block fields that he called "coarse scree" in a 21-square-kilometer (km^2) area. The highly porous, blocky deposits only locally support patches of vegetation. The adjacent forests may have grown on weathered block fields that were partially covered with soil (Hack, 1965).

Debris-Flow Deposits

Debris-flow deposits (Qdf) are the result of catastrophic landslides. The deposits consist of masses of bedrock, soil, and vegetation that move rapidly downslope and usually incorporate more material as they descend. A storm in 1995 caused debris flows and floods in the area between the Robinson and Conway Rivers in Madison County (fig. 25*A*) (Fletcher and Madison 7.5-minute quadrangles) and on the Moormans River in Albemarle County (Browns Cove 7.5-minute quadrangle). As much as 700 mm of rain fell in a 130 km^2 area, which resulted in about 1,000 debris flows in Madison County (Morgan and others, 1999b; Wieczorek and others, 2006); more than 292 mm of rain resulted in about 100 debris flows in the Moormans River area (Morgan and Wieczorek, 1996). Debris flows began as debris slides on upper slopes of approximately 30°±4° in hollows where colluvium had accumulated. The colluvium and regolith detached and slid along the surface of the underlying granitoids and gneisses. They picked up and carried along as much as 90 percent of the load during the descent down an existing channel to the older debris fans on the lower slopes. The scouring and incision of the channels, which were caused by both the debris flows and the subsequent rainfall, were as much as 8 m deep. The high velocities of some debris flows sometimes caused the flows to travel up the sides of the upper channel; the "super elevations" of the deposits left behind indicate that these velocities were as much as 24 meters per second (m/sec) which then decreased to 8 m/sec as the flows neared the valley floor (Wieczorek and others, 2006). The debris flows deposited boulders on the surface of ancient debris fans (fig. 25*B*) and incised channels into them (fig. 25*C*). Scars of the debris flows on the slopes have been slow to revegetate and are still visible 13 years after the event.

In contrast, in the Moormans River area, debris flows occurred on slopes that were less steep (between 20° and 30°), underlain by metabasalt of the Catoctin Formation, and commonly characterized by ground cracks and slumps. The debris flows in this area partially filled the upper end of the Sugar Hollow Reservoir in at least five surges. The bed and banks of the Moormans River were scoured to expose colluvium and stratified slope deposits that are as much as 10 m thick (Morgan and Wieczorek, 1996; Morgan and others, 2000). The stratigraphy exposed in the incised debris fans suggests that debris flows have occurred in the Shenandoah National Park area every 2,500 to 2,000 years (Eaton and others, 2003a). Regionally, at least 51 historical debris-flow events have occurred in the Appalachians between 1844 and 1985 (Clark, 1987) as the result of high rainfall events (Wieczorek and others, 2004). In the past 50 years, storms resulting in debris flows have occurred every 10 to 15 years (summarized in Wieczorek and others, 2004).

Locally, small rock slides, slumps, and landslides along road embankments have been cut into fractured bedrock and (or) unconsolidated soil and debris. Some are found along Skyline Drive but are too small to be portrayed on the map. Several small landslide deposits consisting of blocks of residuum derived from the Tomstown Formation and alluvial-fan deposits along Meadow Run were active during the flooding events in 1937, 1972, 1993, 1997, and 1999. These two types of deposits are the result of an abrupt change in a stream's course during flooding or stream migration along the boundary of the bluff and channel (Chirico, 2004; Wieczorek and others, 2006); the landslides help recycle gravel from the alluvial fans.

Sinkholes

Carbonate rocks that underlie the Great Valley section of the Valley and Ridge province west of the Blue Ridge chemically dissolve to form sinkholes and caverns. Sinkholes were mapped on the basis of (1) areas of closed topographic depressions observed on the ground, (2) depressions indicated by contours on 1:24,000-scale topographic maps, and (3) detection on aerial photographs (Gathright and others, 1977). Only sinkholes that are at least several meters or more in diameter are shown on the geologic map with a special symbol; the maximum diameter of any of the sinkholes in the map area is 305 m across near Luray. Hubbard's (1988) compilation of sinkholes in the Page Valley was based solely on the interpretation of aerial photographs. Although not field checked, his map suggests that sinkholes may be more common than portrayed on this map. Sinkholes are found in all of the rock units in the Great Valley, including the upper part of the Martinsburg Formation. Sinkholes occur also in areas that are covered by alluvial gravels; these formed as the underlying bedrock was chemically dissolved. In the southwestern part of the map area, ponds and historic manganese prospects that are filled with water (Gathright and others, 1977) occur and are difficult to distinguish from water-filled sinkholes. Some of the sinkholes continue into the subsurface and are connected to cave and cavern systems. The commercial caves of Luray Caverns and Skyline Caverns are developed in the rocks of the Beekmantown Group in the map area.

Although karstification is ongoing and sinkholes continue to develop, material deposited within them suggests that they have formed over an extended period of time. Along the west slope of South Mountain (a continuation of Blue Ridge) about 72 km to the north in Pennsylvania, miners of residual iron at Pond Bank from 1864 to 1865 encountered a 5.5-m-thick bed of lignite between 12.5 and 19.5 m below the surface, within residuum that is as much as 122 m thick. The black, carbonaceous clay contained seeds, nuts, and wood fragments and was likely a swamp deposit that formed in a sinkhole (Pierce, 1965). Tschudy (1965) determined that the pollen and spores in the swamp material are Late Cretaceous in age, or about 85 Ma. Pierce (1965) suggested that erosion and dissolution of the underlying bedrock had lowered the original surficial deposit about 427 m, whereas fission-track data (Naeser and others, 2006) suggest that 1,000 m has been removed since deposition of the organic material. Similar deposits have been found in eastern Tennessee (King and Ferguson, 1960). Deposits of kaolinite in eastern Tennessee sinkholes were converted to bauxite by the removal of silica in the kaolin (King and Ferguson, 1960). The surface exposure of the Gray Fossil Site in Tennessee contained bones of the extinct rhinoceros Teleoceras and short-faced bear Plionarctos, that were deposited between 7 and 4.5 Ma (late Miocene to early Pliocene) when a cave collapsed and formed a sinkhole (Kohl, 2000; Wallace and others, 2002; Wallace and Wang, 2004).

Residuum

The physical and chemical weathering of rocks and minerals under different climatic regimes contributed greatly to the landscape evolution of the Shenandoah National Park region. All of the rocks were chemically weathered to form saprolite and (or) residuum that is highly oxidized (red) and leached (yellow). Residuum mantles much of the area and is too extensive to portray on the map. Residuum is currently forming (Hack, 1965), but much of it predates the alluvium, colluvium, alluvial fans, and debris fans which unconformably overlie it. The former landscape of the region may have had a more extensive mantle of residuum and saprolite, which formed during warm and humid climatic conditions. Hillslope residuum was eroded and recycled to form the matrix of debris fans during the Pleistocene.

Local accumulations of subrounded to angular cobbles and boulders of quartz, chert, and jasper mantle the reddish-brown residual soil (terra rossa) in the Great Valley. These silica-rich minerals were derived from veins or from limestone and (or) dolomite beds. They are resistant to erosion and form lag deposits in clay residuum that was derived from the chemical decomposition of the carbonate bedrock. Much of the clay residuum near the boundary between the Blue Ridge and Great Valley also contains limonite and manganese-oxide concretions (Stose and others, 1919; King, 1950). Carbonate rocks of the Beekmantown Group were locally hydrothermally altered and replaced by yellow jasper beneath the Front Royal fault in the Chester Gap 7.5-minute quadrangle and were quarried by native Americans in the Flint Run Archeological District (Bentonville 7.5-minute quadrangle; Gardner, 1977, 1983). The jasper residuum may also be an old deposit that formed in a Late Cretaceous to Eocene climate that favored the development of laterites (which contain bauxite, kaolinite, and lignite) in the Appalachians (Overstreet, 1964; Pierce, 1965; Tschudy, 1965; Hack, 1979; McLaughlin and Darrell, 1972; McLaughlin and McBroom, 1980; Wallace and others, 2002; Wallace and Wang, 2004; and Bearce and Carroll, 2003).

Acknowledgments

This research was funded by the U.S. Geological Survey's National Cooperative Geologic Mapping Program's Federal Mapping Component (FEDMAP) and University Geologic Mapping Component (EDMAP). Significant contributions were made by student research assistants from George Washington University and the College of William and Mary, who were supported by EDMAP. Technical reviews by Arthur P. Schultz and Arthur J. Merschat and additional review by Richard W. Harrison improved this report. We remain indebted to residents and National Park Service managers within the Shenandoah National Park who provided access to outcrops on both private and public lands. Thomas M. Gathright, II, who retired from the Virginia Division of Mineral Resources, provided the foundation for much of the current research with his landmark publication entitled "Geology of the Shenandoah National Park, Virginia" (Gathright, 1976). He continues to provide inspiration, support, and encouragement for studies in the Blue Ridge.

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Description of Map Units

[For thickness of units, see cross sections.]

Unconsolidated Surficial Materials and Landforms

Qa	**Alluvium (Holocene)**—Unconsolidated silt, sand, cobbles, and boulders. Located along streams, flood plains, and alluvial plains. Thickness is as much as 12 m. Alluvium in the highlands is mostly boulders and is transitional with debris fans on lower slopes
Qdf	**Debris-flow deposit (Holocene)**—Debris flow scars and tracks in the Blue Ridge highlands resulted from high rainfall associated with a tropical storm in June 1995. Consists of boulder deposits on chutes in bedrock and residuum
Qb	**Quartzite block-field deposit (Holocene and Pleistocene)—**Clast-supported, open framework of subangular, lichen-covered quartzite blocks and boulders. Commonly located on nonvegetated slopes of ridges underlain by quartzite of the Chilhowee Group. Thickness is variable
Qc	**Colluvium (Holocene and Pleistocene)—**Clast-supported diamicton of subangular to subrounded cobbles and boulders. Grades downslope and is transitional into debris-fan deposits. Thickness ranges from a thin veneer to 30 m
Nt	**Terrace deposit (Neogene)—**Unconsolidated sand, gravel, cobbles, and boulders along major rivers. Deposits range from 0 to 9 m thick on terraces to as much as 9 m above present flood plains. Includes elevated terraces as much as 60 m above present flood plains
Np	**Alluvial-plain deposit (Neogene)—** Broad areas of coalescing terraces along major rivers in the lowlands east of the Blue Ridge highlands. Little to no transported material preserved. Older alluvial plains shown by pattern
Nd	**Debris-fan deposit (Neogene)—**Diamicton consisting of subrounded pebbles, cobbles, and boulders, of local rocks supported in a matrix of unstratified clay, silt, sand, and pebbles. Forms fans and sheets on lower slopes and valleys. Includes terraces of debris fans as much as 36 m above the adjacent debris fans and alluvium. Thickness is as much as 30 m
Nf	**Alluvial-fan deposit (Neogene)—**Unconsolidated sand, pebbles, cobbles, and boulders of quartzite and sandstone, mostly derived from the Harpers and Antietam Formations, Forms coalescing fans and sheets that extend from the lower west slope of Blue Ridge to the South Fork of the Shenandoah River. Thicknesses are highly variable as a function of the weathered state of the underlying bedrock and subsequent alluvial modification. Bedrock residuum is locally exposed beneath a thin veneer of gravel but drill-hole data and mining operations show fan thicknesses may locally exceed 150 m

Valley and Ridge Province

Mesozoic Igneous Rocks

Jpd	**Peridotite dike (Late Jurassic)**—Elliptical peridotite dike 182 m long and 46 m wide intruding the Martinsburg Formation west of Front Royal. Consists of olivine and pyroxene altered to chlorite, phlogopite, biotite, ankerite, and talc (Young and Bailey, 1955)
Jad	**Alkalic dike (Late Jurassic)**—Northwest-trending vertical alkalic dikes consisting of nepheline syenite, teschenite, and teschenite-picrite intruding the Martinsburg Formation west of Grottoes (Johnson and others, 1971)
Jd	**Diabase dike (Early Jurassic)**—Medium- to dark-gray, moderately crystalline and equigranular, massive diabase dikes that weather to subrounded boulders with characteristic orange-brown-surface. Dikes consist of olivine-normative (olivine-plagioclase-pyroxene) tholeiitic diabase and low-titanium, quartz-normative tholeiitic diabase containing centimeter-sized clusters of calcic plagioclase phenocrysts in a fine-grained groundmass of pyroxene and plagioclase

Massanutten Synclinorium

DSu **Devonian and Silurian rocks, undivided**—Includes the formations listed below

Mahantango Formation (Middle Devonian)—Gray mudstone, sandstone, and fossiliferous shale

Needmore Shale (Middle and Lower Devonian)—Greenish-gray, fossiliferous shale and calcareous mudstone; black shale at the base

Tioga Ash Bed (Middle and Lower Devonian)—Gray shale and siltstone; brown-biotite-bearing, calcareous ash; and black, fissile shale

Marcellus Shale, undivided (Middle and Lower Devonian)—Dark-gray to black, fissile shale with interbeds of gray, silty limestone and calcareous shale

Ridgeley Sandstone (Lower Devonian)—Light-gray fine- to coarse-grained, cross-laminated, calcareous, fossiliferous, locally conglomeratic sandstone

Helderberg Group (Lower Devonian)—Light-gray, laminated to thick bedded limestone; black, nodular chert; and white, blocky chert. Upper part is sandy

Keyser Limestone (Lower Devonian and Upper Silurian)—Gray, fossiliferous limestone and gray, laminated limestone with black, nodular chert and white, blocky chert

Tonoloway Limestone (Upper Silurian)—Gray, laminated limestone with mudcracks

Wills Creek Formation (Upper Silurian)—Gray limestone and greenish-gray, calcareous siltstone and mudstone

Bloomsburg Formation (Upper Silurian)—Red mudstone interbedded with red, ferruginous sandstone and shale

McKenzie Formation (Upper and Middle Silurian)—Gray, calcareous shale

Sm **Massanutten Sandstone (Silurian)—**Light-gray, fine- to coarse-grained, cross laminated, locally conglomeratic sandstone

Great Valley and Page Valley

Om **Martinsburg Formation (Upper and Middle Ordovician)—**Light-brown shale, calcareous shale, and siltstone. Contains thin to medium beds of sandstone and metagraywacke in the upper part; gray, argillaceous limestone at the base

Oeln **Edinburg Formation, Lincolnshire Limestone, and New Market Limestone, undivided (Middle Ordovician)—**Gray to black, fossiliferous limestone and black shale; gray, fossiliferous and cherty limestone; and bluish-gray, micritic limestone

Ob **Beekmantown Group, undivided (Middle and Lower Ordovician)—**Light-gray, medium- to thick-bedded dolostone and laminated dolostone containing white and light-gray chert nodules. Weathered surface characterized by “butcher-block” of cross-hatched joints. Irregular bedding at top is due to collapse breccia and paleokarst

Os **Stonehenge Limestone (Lower Ordovician)—**Dark-gray, fine- to medium-grained, thick-bedded, fossiliferous limestone with minor black chert. Contains algal bioherms, intraformational conglomerates, and bioclastic and dolostone beds. Also contains minor light-gray, laminated, silty limestone with thin interbeds of platy limestone and coarse-grained bioclastic limestone; and light- to dark-gray, fine- to medium-grained, thin- to medium-bedded, fossiliferous limestone and crystalline dolostone containing gray chert nodules

OЄc **Conococheague Limestone (Lower Ordovician and Upper Cambrian)—**Light-gray, calcareous to dolomitic sandstone; medium-gray, fine-grained limestone with intraformational conglomerate; and light-gray, fine-grained dolostone interbedded with dark- to light-gray, laminated, algal limestone; dolomitic limestone; and light-brown dolostone and calcareous sandstone

Єe **Elbrook Limestone (Upper and Middle Cambrian)—**Medium-gray, thinly bedded limestone interbedded with white marble; light-brown, laminated dolostone; and thin, calcareous shale and shaly dolostone

Єwa **Waynesboro Formation (Lower Cambrian)—**Consists of three units too thin to map separately, in ascending order: (1) interbedded light-olive-gray shale; light-gray, fine-grained sandstone; and medium- to dark-gray, sandy, dolomitic limestone; (2) interbedded gray, bioturbated dolostone, dolomitic limestone, and laminated limestone, with a few thin, sandy limestone beds near the middle; and (3) interbedded dusky-red shale, mudstone, and argillaceous sandstone; light-gray sandstone; and light-brown, sandy, dolomitic limestone and dolostone

Ꞓt **Tomstown Formation (Lower Cambrian)**—Light- to dark-gray limestone, dolostone, and marble. Includes four subdivisions too thin to map separately, in ascending order: (1) dark-gray, fine-grained, thin-bedded limestone with wispy dolomitic burrows that increase in abundance upsection; may include a basal 15-m-thick interval of gray to white, mylonitic marble; (2) dark-gray, thick-bedded, burrow-mottled dolostone; (3) light-gray, very thick bedded to massive, sugary dolostone with faint cross-bedding with a base that is gradational with the underlying bioturbated dolostone; and (4) a unit consisting of a lower dark-gray, bioturbated dolostone interbedded with intervals of dark-gray, laminated dolostone and dark-gray limestone, and an upper dark-gray, bioturbated, oolitic dolostone interbedded with laminated limestone and silty dolostone

Blue Ridge Province

Chilhowee Group

Ꞓca **Antietam Formation (Lower Cambrian)**—Light-olive- to olive-gray, medium- to coarse-grained, medium-bedded, locally ferruginous, micaceous, silty metasandstone interbedded with very fine grained, silty metasandstone to sandy metasiltstone. Local ferruginous horizons with abundant botryoidal hematite and limonite are located near contact with overlying Tomstown Formation

Ꞓch **Harpers Formation (Lower Cambrian)**—Greenish- to bluish-gray quartz-chlorite-sericite phyllite and metasiltstone interbedded with thin gray metasandstone, quartzite, and meta-arkose; also includes interbedded layers of quartzite, metasandstone, and metasiltstone. Lower part locally consists of green and brown metasiltstone interbedded with fine metasandstone, but dark quartzite occurs at Blackrock. In northern part of the map, characterized by upward-coarsening sequences of bioturbated phyllite with intervals of wavy, thick laminations that grade into blocky beds of metasiltstone and fine sandstone that contain abundant burrows. Top of the lower part consists of quartzite with interbeds of metasandstone and thin dolomite. Local beds of pebbly metasandstone and metaconglomerate are in the lower part and vitreous quartzite and ferruginous arkosic metasandstone are in the upper part. The trace fossil *Skolithos linearis* appears to be limited to clean quartzite beds in the upper part

Ꞓchs Ferruginous metasandstone—Maroon and dark-blue to black, ferruginous metasandstone

Ꞓcw **Weverton Formation (Lower Cambrian)**—Maroon and gray, laminated phyllite that includes beds of pebbly quartzite interbedded with siltstone, and metaconglomerate at the base. Coarse-grained, sandy and pebbly metasandstone interbedded with silvery-green, quartzose phyllite and reddish-purple, coarse-grained, thick-bedded, ferruginous metasandstone. Locally capped by light-tan to brown, pebbly metasandstone and maroon ferruginous metasandstone in the southern part of the Shenandoah National Park. In the northern part of the park, consists of upward-coarsening sequences of heavily bioturbated metasiltstone and phyllite with intervals of wavy, thick laminations; blocky, fine-grained quartzite that is commonly bioturbated with flat and wavy laminations; and a cap of crossbedded, pebbly metasandstone and metaconglomerate

Catoctin Formation (Neoproterozoic)

Zcm Metabasalt—Light-green to dark-greenish-gray to medium-bluish-gray, fine-grained to aphanitic, massive to schistose, amygdaloidal metabasalt composed of actinolite, chlorite, epidote, albite, and rare quartz. Contains lenses and layers of light-green, fine-grained, massive epidosite, which consists of epidote and quartz. Locally contains thin layers of phyllite and metasandstone

Zcp Metavolcanic phyllite—Dark, variegated (gray, blue, dusky-red), mottled to lustrous phyllite containing white to green, elongated vesicles and smeared sericite and chlorite blebs

Zcs Metasandstone and laminated phyllite—Light-greenish-gray, medium- to coarse-grained metasandstone and variegated, finely laminated quartz-muscovite phyllite. Occurs within metabasalt in the upper part of the formation from the extreme southwest part of the map area northeast for about 17 km

Zcr Metarhyolite—Light-gray to pinkish-white, very fine grained metarhyolite and tuff containing feldspar phenocrysts and centimeter-wide clots of white quartz and pale-green epidote. Occurs as a layer about 10 m thick and 5 km long in metabasalt near the base of the formation in the extreme northeast part of the map area (Lukert and Nuckols, 1976)

Zmd **Metadiabase dike (Neoproterozoic)**—Dark-greenish-gray, fine- to medium-grained, massive to schistose metadiabase composed predominantly of chlorite, albite, epidote, and actinolite. Coarse-grained variety has 2- to 8-mm-long actinolite pseudomorphs after clinopyroxene, and aphanitic variety has relict euhedral laths of plagioclase. Similar in composition to metabasalt of the Catoctin Formation; interpreted to be feeder dikes to those metabasalt flows. Includes augite-bearing porphyritic and actinolite-rich amphibolite dikes in the northeastern part of the map area (Lukert and Nuckols, 1976; Lukert and Halladay, 1980)

Zsr **Swift Run Formation (Neoproterozoic)**—Pink to gray, very coarse to medium-grained, crossbedded metasandstone and quartzite containing pebbles and cobbles of quartz phyllite and sandstone; brownish-green chlorite-sericite-feldspar-quartz metagraywacke; and lustrous, silvery quartz-sericite schist. Also includes grayish-reddish-purple phyllite; grayish-green, finely laminated phyllite; dark-greenish-gray to brownish-gray, sandy, sericitic phyllite; and medium-dark-gray slate in fining-upward sequence

Zhg **Hornblende metagabbro dike and (or) sill (Neoproterozoic)**—Greenish-black, medium- to fine-grained, massive to foliated dike or sill consisting of hornblende, plagioclase, epidote, and quartz. An altered ultramafic rock consisting of actinolite, epidote, chlorite, and magnetite is found locally beneath hornblende metagabbro east of Graves Mill (Bailey and others, 2003)

Zmr **Mechum River Formation (Neoproterozoic)**—Consists of four units (Bailey, Peters, and others, 2007) not differentiated on the map: (1) laminated metasiltstone and metamudstone with minor arkosic metawacke; (2) fine- to coarse-grained, graded arkosic metasandstone and metawacke interbedded with minor metasiltstone and metaconglomerate; (3) crossbedded arkosic metasandstone and metaconglomerate; and (4) laminated metasiltstone, metagraywacke, and crossbedded, arkosic metawacke

Robertson River Igneous Suite (Neoproterozoic) (Tollo and Lowe, 1994)

Zra	Amissville Alkali Feldspar Granite—Gray, medium-grained alkali feldspar granite composed of mesoperthite, quartz, and diagnostic quartz phenocrysts. Small, irregular (miarolitic) cavities contain protruding quartz crystals
Zrb	Battle Mountain Alkali Feldspar Granite—Gray, medium-grained alkali-feldspar granite composed of alkali-feldspar-dominant mesoperthite, and quartz; local small, irregular (miarolitic) cavities contain protruding quartz crystals
Zrbr	Rhyolite and conglomerate—Light-gray, aphanitic, extrusive metarhyolite containing fluorite and feldspar phenocrysts; interlayered with metaconglomerate consisting of pebbles of metarhyolite and conglomerate comprised granite boulders
Zrbf	Felsite—Light-gray, aphanitic, extrusive felsic volcanic rock composed of phenocrysts of alkali-feldspar-dominant mesoperthite and quartz; locally displays flow banding, lithophysae (hollow, bubblelike structures composed of concentric shells of finely crystalline minerals), and small, irregular (miarolitic) cavities containing protruding quartz crystals
Zrbd	Felsic dike—Light-gray, aphanitic, felsic dikes; intruded Mesoproterozoic gneisses as well as granites of the Robertson River Igneous Suite
Zrh	Hitt Mountain Alkali Feldspar Syenite—Gray, coarse-grained, inequigranular, locally pegmatitic alkali-feldspar syenite composed of mesoperthite (chiefly microcline), quartz, saussuritized plagioclase, hastingsite, and rare garnet
Zrqt	Quartz trachyte—Dusky-red to gray, aphanitic to fine-grained quartz trachyte composed of mesoperthite and quartz with light-gray mesoperthite phenocrysts; occurs in one locality in the park, southeast of Jenkins Gap, between Mesoproterozoic rocks and Catoctin Formation. Locally layered and fragmented; interpreted to be the extrusive volcanic equivalent of Cobbler Mountain Alkali Feldspar Quartz Syenite
Zrc	Cobbler Mountain Alkali Feldspar Quartz Syenite—Gray, medium-grained, and massive alkali-feldspar quartz syenite composed of conspicuous stubby, euhedral mesoperthite grains that are 2 to 4 mm in diameter and are intergrown with anhedral quartz
Zrw	White Oak Alkali Feldspar Granite—Gray, coarse-grained, inequigranular alkali-feldspar granite composed of mesoperthite (chiefly microcline), quartz, and hastingsite
Zrl	Laurel Mills Granite—Gray, coarse-grained, inequigranular granite composed of alkali-feldspar-dominant mesoperthite, diagnostic pale-blue quartz, hastingsite, biotite, magnetite, and titanite

Zram	Arrington Mountain Alkali Feldspar Granite—Gray, medium-grained, equigranular alkali-feldspar granite composed of mesoperthite (chiefly microcline), quartz, and hastingsite, biotite, fluorite, and rare garnet and muscovite
Zrr	Rivanna Granite—White, medium-grained, and equigranular granite composed of quartz, plagioclase, and alkali-feldspar-dominant mesoperthite (chiefly microcline), biotite, fluorite, and rare muscovite. Local small, irregular (miarolitic) cavities contain protruding quartz and pyrite crystals
Zp	**Garnet-graphite paragneiss (Neoproterozoic)—**Rusty-brown, medium- to fine-grained, compositionally layered graphite-garnet-biotite-plagioclase-quartz gneiss. Includes garnetiferous, quartzofeldspathic, and quartzitic layers that are several millimeters thick. Almandine garnets 0.1 to 1 cm in diameter occur as aggregates; graphite occurs as small disseminated flakes. Exhibits foliation parallel to layering. Locally cut by pegmatite. Commonly retrograded to chlorite schist with Paleozoic schistosity
Ybg	**Biotite monzogranite-quartz monzodiorite (Mesoproterozoic)—**Very dark gray, medium- to coarse grained, inequigranular, nonfoliated to weakly foliated biotite monzogranite and quartz monzodiorite. Biotite content is as much as 25 percent
Yom	**Orthopyroxene monzogranite-quartz monzodiorite (Mesoproterozoic)—**Dark-green to black, medium- to coarse-grained, inequigranular, massive and nonfoliated, orthopyroxene-, amphibole-, and clinopyroxene-bearing orthopyroxene monzogranite and quartz monzodiorite
Ypb	**Megacrystic quartz monzonite (Mesoproterozoic)—**Light- to medium-gray, medium-grained to megacrystic, weakly foliated quartz monzonite containing porphyroblasts of pink microcline
Ybm	**Megacrystic biotite monzogranite (Mesoproterozoic)—**Light- to medium-gray, very coarse grained, inequigranular, weakly to moderately foliated, megacrystic biotite monzogranite. Includes variable amounts of biotite; quartz is typically blue. Intruded by comagmatic dikes of pegmatite and biotite monzogranite
Ycg	**Crozet Granite (Mesoproterozoic)—**Light-gray, very coarse grained, massive, nonfoliated to moderately foliated, biotite- and clinopyroxene-bearing monzogranite. Includes megacrysts of euhedral feldspar and anhedral quartz as much as 10 cm long. Intruded by undeformed pegmatite dikes
Yor	**Old Rag Granite (Mesoproterozoic)—**White- to light-gray, medium- to coarse-grained, inequigranular, massive, nonfoliated to weakly foliated, biotite-, and orthopyroxene-bearing, garnetiferous leucogranite and garnetiferous syenogranite containing gray and blue quartz grains
Yml	**Porphyroclastic metagranitoid (Mesoproterozoic)—**Light-gray, very coarse grained, strongly foliated, biotite-bearing alkali-feldspar granite
Yog	**Orthopyroxene granite-monzogranite (Mesoproterozoic)—**Dark-gray, medium- to coarse-grained, equigranular, weakly to strongly foliated,

orthopyroxene-, amphibole-, biotite-, and garnet-bearing orthopyroxene granite-monzogranite

Yfh **Flint Hill Gneiss (Mesoproterozoic)—**Dark- to medium-gray, medium-grained, inequigranular, strongly foliated, locally migmatitic quartzofeldspathic syenogranite to monzogranite in compositional layers separated by biotite. Includes gray and blue quartz grains and blue quartz veins

Yll **Lineated leucogranite gneiss (Mesoproterozoic)—**Light-gray to tan, medium-to coarse-grained, strongly foliated, lineated leucogranite gneiss. Exhibits diagnostic stripped texture due to lineations of biotite

Yomg **Orthopyroxene monzogranite-quartz monzodiorite gneiss (Mesoproterozoic)—**Greenish-gray to black**,** strongly foliated, orthopyroxene-, clinopyroxene, and biotite-bearing orthopyroxene monzogranite and quartz monzodiorite gneiss. Exhibits rusty-weathered, ribbed surface and transposed leucocratic layers

Yos **Megacrystic orthopyroxene syenogranite-monzogranite gneiss (Mesoproterozoic)—**Dark-gray to dark-greenish-gray, very coarse grained, strongly foliated, orthopyroxene-, amphibole-, and clinopyroxene-bearing megacrystic orthopyroxene syenogranite-monzogranite gneiss; alkali-feldspar megacrysts are as much as 12 cm long

Ygd **Granodiorite gneiss (Mesoproterozoic)—**Light-gray, equigranular, compositionally layered, strongly foliated granodiorite gneiss that includes clots and lineations of biotite

Yoq **Orthopyroxene quartz diorite gneiss (Mesoproterozoic)—**Greenish-gray to black, medium-grained, compositionally layered, strongly foliated, orthopyroxene-, biotite-, and garnet-bearing orthopyroxene quartz-diorite gneiss

Yon **Orthopyroxene syenogranite and monzogranite gneiss (Mesoproterozoic)—**Gray, medium- to coarse-grained, compositionally layered, strongly foliated orthopyroxene-, biotite-, garnet-, and clinopyroxene-bearing orthopyroxene syenogranite and monzogranite gneiss. Exhibits transposed garnetiferous leucocratic layers

Yod **Orthopyroxene granodiorite gneiss (Mesoproterozoic)—**Dark-gray, medium-to coarse-grained, compositionally layered, strongly foliated, amphibole-bearing orthopyroxene granodiorite gneiss. Exhibits transposed leucocratic layers

Ypg **Foliated, garnetiferous, porphyroblastic monzogranite (Mesoproterozoic)—**Medium-gray, variegated, medium- to coarse-grained**,** biotite-bearing, moderately foliated, garnetiferous, porphyroblastic monzogranite. Contains microcline megacrysts or aggregates as much as 3 cm in diameter and distinctive clots of blue quartz

Ylg **Leucogranite gneiss (Mesoproterozoic)—**Light-gray, coarse-grained to megacrystic, weakly to strongly foliated leucogranite gneiss. Composition varies and includes alkali-feldspar granite, syenogranite, and monzogranite. Intruded by medium-gray, medium-grained, equigranular, biotite-bearing, isoclinally folded leucogranite dikes. Both gneiss and dikes are in turn intruded by leucocratic pegmatites

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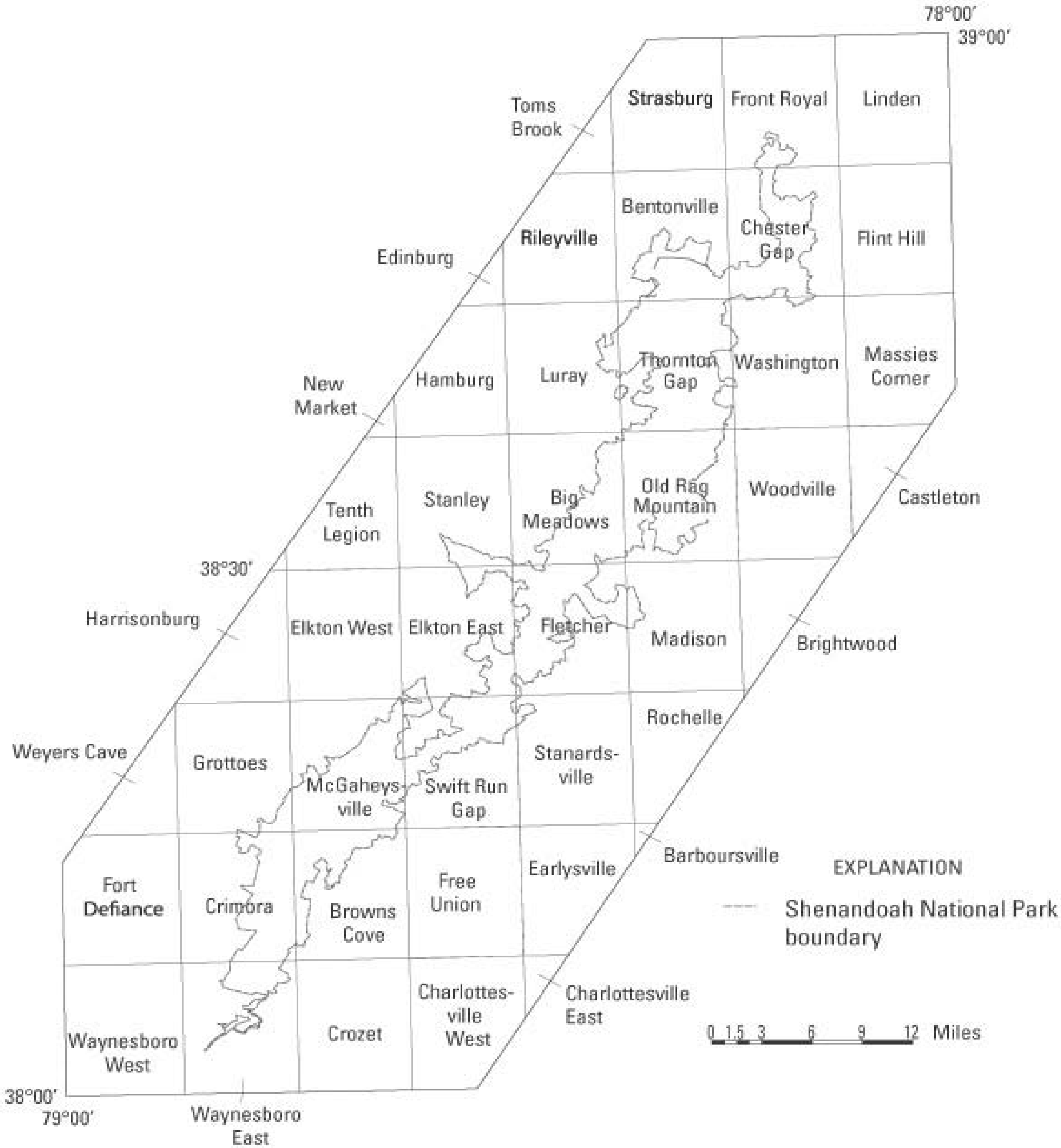

Figure 1. Index map showing the 7.5-minute 1:24,000-scale quadrangles that cover the map area (plate 1) and a list (by quadrangle) of the published reports and unpublished data used in the compilation of the geologic map.

SOURCES OF GEOLOGIC DATA

Barboursville	C.M. Bailey and Scott Southworth (unpub. data, 2008)
Bentonville	Scott Southworth and R.P. Tollo (unpub. data, 2006)
Big Meadows	R.P. Tollo and L.O. Olsen (George Washington University, unpublished EDMAP data, 2004)
Brightwood	Tollo and Lowe (1994); Bailey and others (2007b)
Browns Cove	Olney and others (2007); J.G. Olney, M.H. Lamoreaux, J.E., Tadlock, O. Nicholls, and C.M. Bailey (College of Willaim and Mary, unpublished EDMAP data, 2007); Scott Southworth (unpub. data, 2008)
Castleton	Tollo and Lowe (1994); Bailey and others (2007b)
Charlottesville West	Bailey and others (2007b)
Charlottesville East	Bailey and others (2007b)
Chester Gap	R.P. Tollo and E.A. Borduas (George Washington University, unpublished EDMAP data, 2002); Scott Southworth (unpub. data, 2005)
Crimora	Gathright and others (1978a)
Crozet	Bailey and others (2007b); A. Gattuso, A. Hartmann, F. Knuth, C. Lemon, and C.M. Bailey (College of Willaim and Mary, unpublished EDMAP data, 2008)
Earlysville	Tollo and Lowe (1994); Bailey and others (2007b)
Edinburg	Rader and Gathright (2001a)
Elkton East	King (1950); Scott Southworth (unpub. data, 2005)
Elkton West	King (1950); Rader and Gathright (2001b)
Fletcher	Tollo and others (2004b); Scott Southworth (unpub. data, 2006)

Figure 1 Index map showing the 7.5-minute 1:24,000-scale quadrangles that cover the map area (plate 1) and a list (by quadrangle) of the published reports and unpublished data used in the compilation of the geologic map—Continued.

Flint Hill	Lukert and Nuckols (1976); Tollo and Lowe (1994); R.P. Tollo, A.G. Kentner, C. Parendo, and R.M. Gesserman (George Washington University, unpublished EDMAP data, 2007)
Fort Defiance	Gathright and others (1978b)
Free Union	Bailey and others (2007b); Scott Southworth (unpub. data, 2007)
Front Royal	Rader and Biggs (1975); Scott Southworth and R.P. Tollo (unpub. data, 2005)
Grottoes	Gathright and others (1978a)
Hamburg	Rader and Gathright (2001a)
Harrisonburg	Rader and Gathright (2001b)
Linden	Lukert and Nuckols (1976); Scott Southworth (unpub. data, 2005)
Luray	Rader and Gathright (2001a; Scott Southworth (unpub. data, 2006)
McGaheysville	Rader and Gathright (2001b)
Madison	Bailey and others (2003); Bailey and others (2007b)
Massies Corner	Lukert and Halladay (1980); Tollo and Lowe (1994); R.P. Tollo, A.G. Kentner, C. Parendo, and R.M. Gesserman (George Washington University, unpublished EDMAP data, 2007)
Old Rag Mountain	Hackley (1999); Tollo and others (2004b); Southworth (unpub. data, 2005)
Rileyville	Scott Southworth and R.P. Tollo (unpub. data, 2006)
Rochelle	Tollo and Lowe (1994); Bailey and others (2007)
Stanardsville	Bailey and others (2007); W.C. Burton and C.M. Bailey (unpub. Data, 2008)

Figure 1 Index map showing the 7.5-minute 1:24,000-scale quadrangles that cover the map area (plate 1) and a list (by quadrangle) of the published reports and unpublished data used in the compilation of the geologic map—Continued.

Stanley	King (1950); Scott Southworth (unpub. data, 2008)
Strasburg	Rader and Biggs (1976)
Swift Run Gap	Forte and others (2005); A.M. Forte, K.M. Wooten, B.A. Hasty, and C.M. Bailey (College of William and Mary, unpublished EDMAP data, 2005); Scott Southworth (unpub. data, 2007);
Tenth Legion	Rader and Gathright (2001a)
Toms Brook	Rader and Biggs (1976)
Thornton Gap	Tollo and others (2004b); Scott Southworth (unpub. data, 2005);
Washington	R.P. Tollo, T.M.S. Tensku Hassan, L. Orentzal, and M. Kuhlman (George Washington University, unpublished EDMAP data, 2006)
Waynesboro East	Gathright and others (1977); Scott Southworth (unpub. data, 2007);
Waynesboro West	Gathright and others (1977); Scott Southworth (unpub. data, 2007)
Weyers Cave	Rader and Gathright (2001b)
Woodville	Tollo and Lowe (1994); Bailey and others (2007); R.P. Tollo, T.M.S. Tensku \Hassan, L. Orentzal, and M. Kuhlman (George Washington University, unpublished EDMAP data, 2006

Figure 1 Index map showing the 7.5-minute 1:24,000-scale quadrangles that cover the map area (plate 1) and a list (by quadrangle) of the published reports and unpublished data used in the compilation of the geologic map—Continued.

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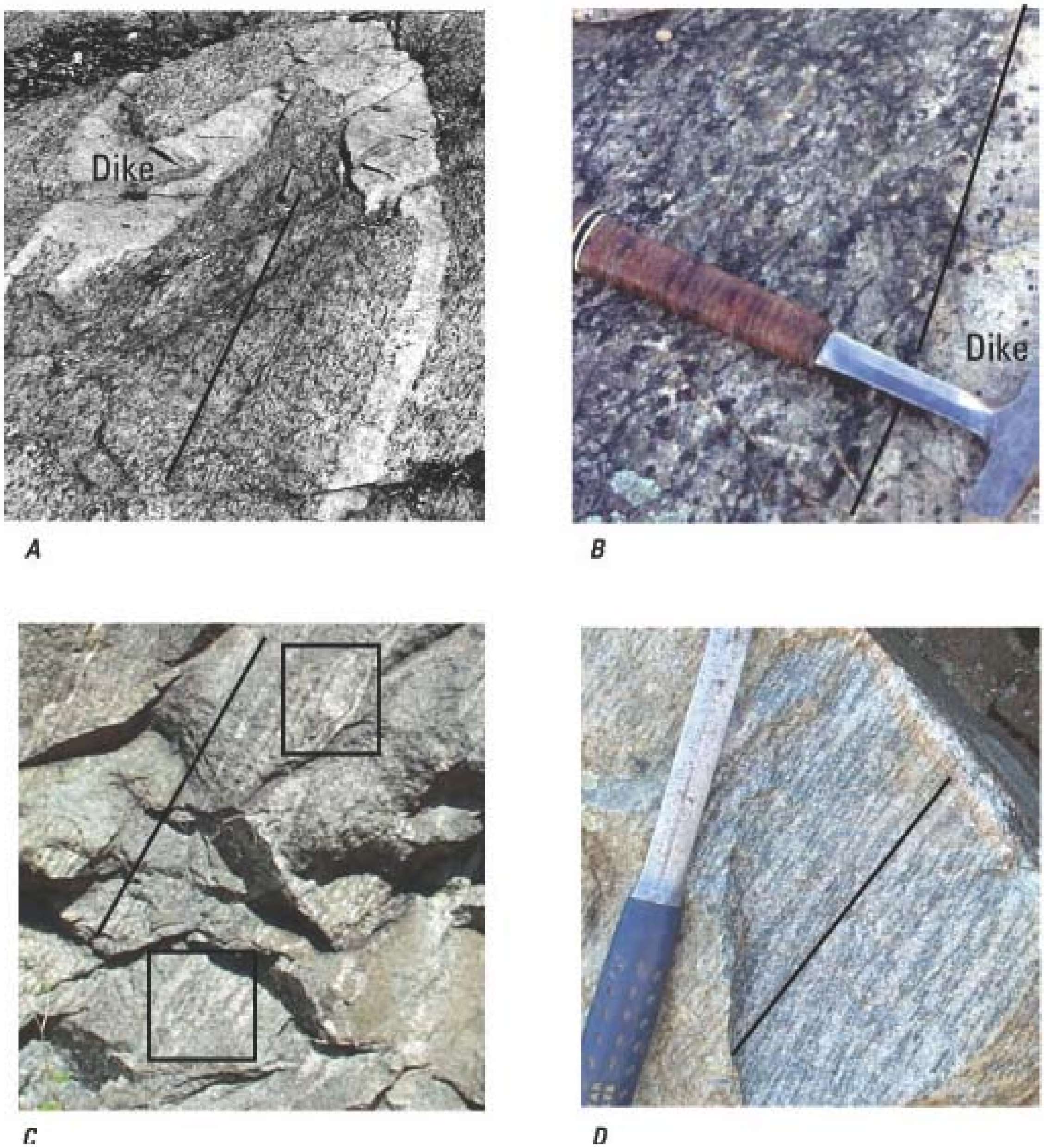

Figure 2. Photographs showing Group 1 foliated orthogneisses. *A*, Northwest-trending foliation (indicated by dark line) in leucogranite gneiss (Ylg; 1,183±11 Ma, table 1, sample 36) is axial planar to folded leucogranite dike (1,175±11 Ma). Pen (for scale) in top center is parallel to foliation (Bailey and others, 2003). *B*, Foliated, garnetiferous, porphyroblastic monzogranite (Ypg; 1,172±8 Ma) cut by garnetiferous monzogranite dike (right of dark line) occurs east of the Rivanna Granite of the Robertson River Igneous Suite (Zrr). Hammer for scale. *C*, Northeast-trending foliation (parallel to dark line) in orthopyroxene granodiorite gneiss (Yod; 1,165±7 Ma, table 1, sample 33) contains transposed, intrafolial folds (within boxes in the center of photograph) Boxes are 1 m square. *D*, Strong foliation (parallel to dark line) in orthopyroxene syenogranite and monzogranite gneiss (Yon; 1,164±8 Ma, table 1, sample 32). Hammer for scale.

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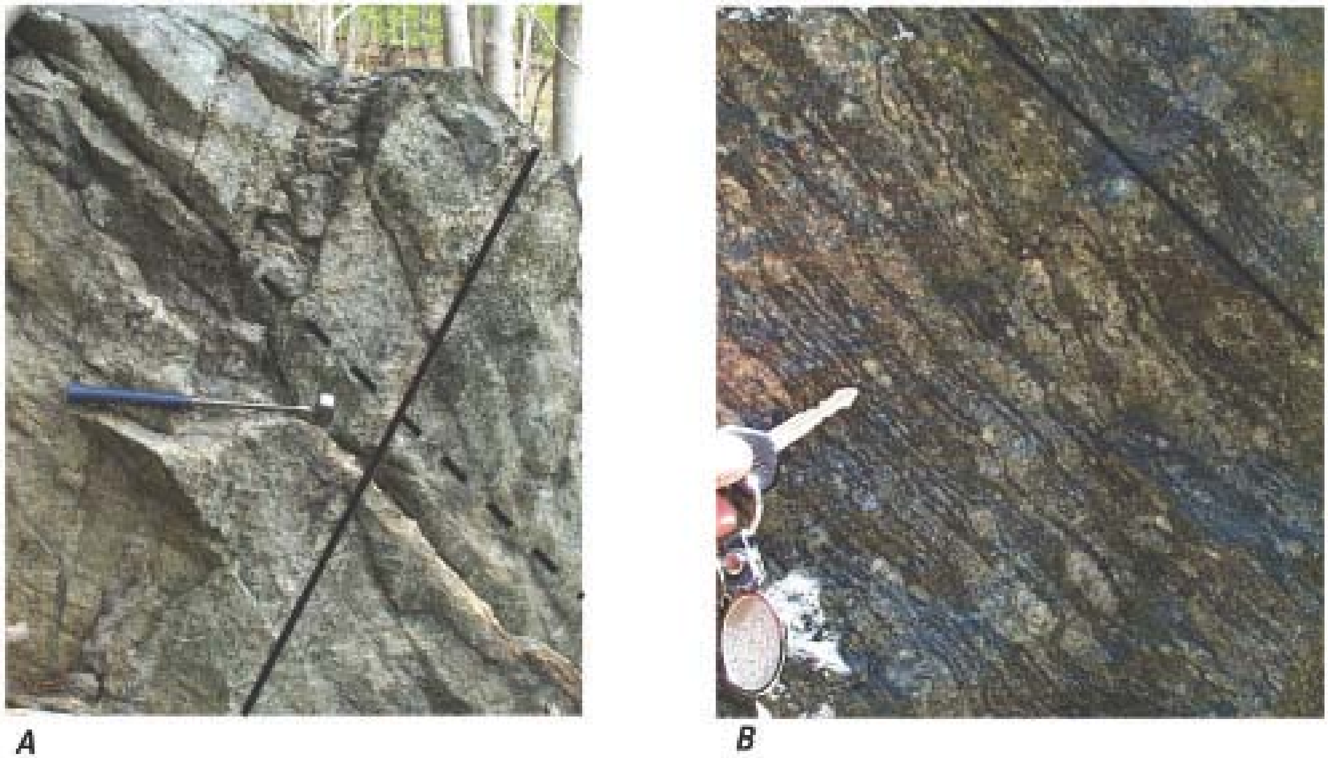

Figure 3. Photographs showing Group 1 orthogneisses. *A*, Granodiorite gneiss (Ygd; 1,161±9 Ma, table 1, sample 30). Northwest-dipping gneissic foliation (lower left to upper right) is cut by southeast-dipping Paleozoic spaced cleavage (upper left to lower right). Hammer for scale. *B*, Strongly foliated megacrystic orthopyroxene syenogranite-monzogranite gneiss (Yos; 1,159±14 Ma, table 1, sample 29) at the south portal of the tunnel at Marys Rock along the Skyline Drive south of Thornton Gap (Tollo, Aleinikoff, Borduas, and Hackley, 2004; Tollo, Bailey, Borduas, and Aleinikoff, 2004; Tollo and others, 2006). Keys for scale.

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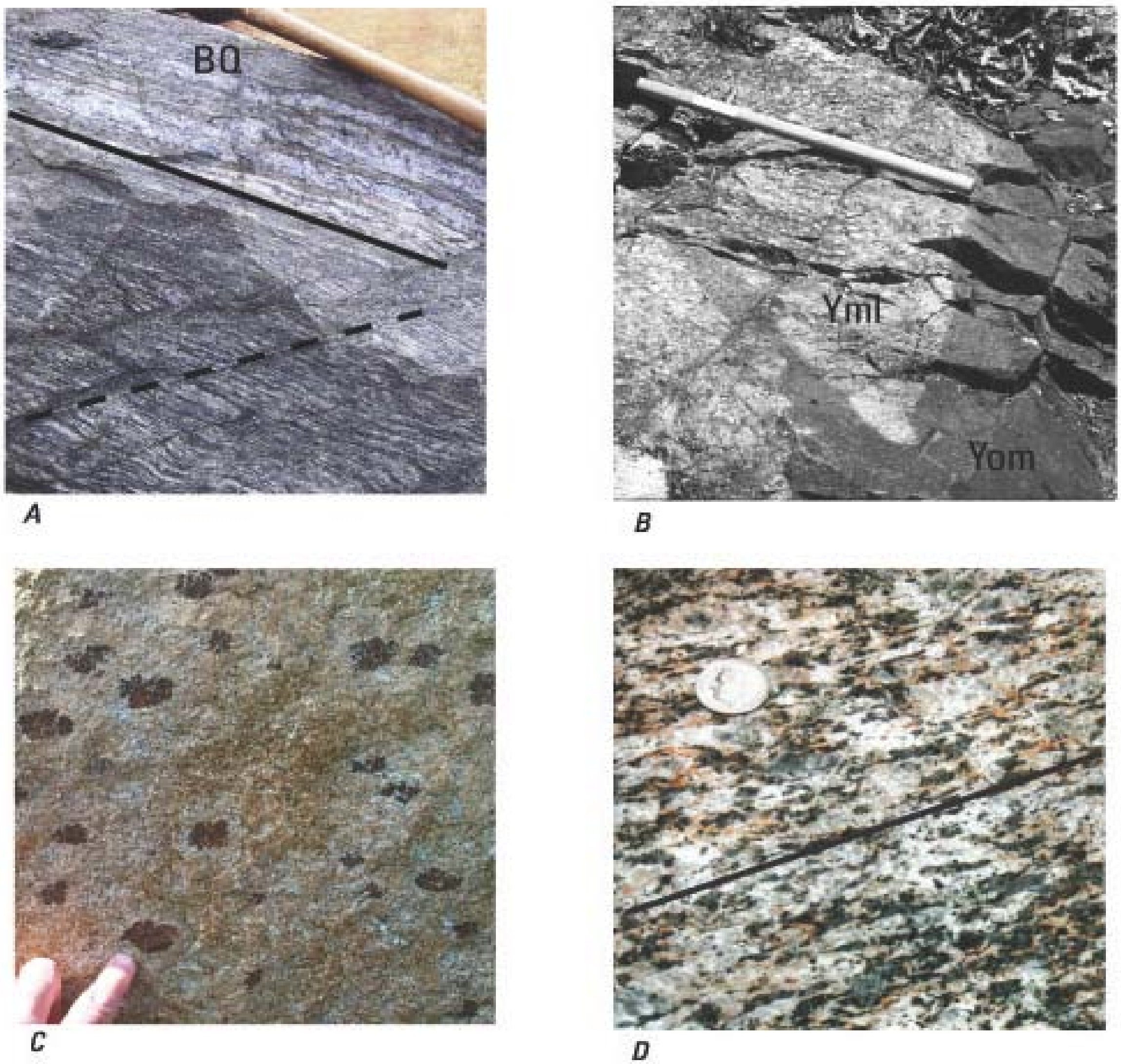

Figure 4. Photographs showing structural foliations in rocks from Groups 1 and 3. *A*, Southeast-dipping foliation in Group 1 Flint Hill Gneiss (Yfh;1,144±8 Ma) (solid line) contains transposed leucosomes and blue quartz veins (BQ). Northwest-dipping Paleozoic cleavage (dashed line inclined to the upper right) kinks the Mesoproterozoic foliation. Hammer for scale. *B*, Xenolith of Group 3 foliated porphyroclastic metagranitoid (Yml; 1,078±9 Ma) in nonfoliated orthopyroxene monzogranite-quartz monzodiorite (Yom; 1,050±8 Ma). Foliation is defined by aligned porphyroclasts that are parallel to the hammer handle. *C*, Group 3 massive garnetiferous syenogranite of the Old Rag Granite (Yor; 1,063±8 Ma, table 1, sample 22). Dark splotches are garnet aggregates. *D*, Group 3 garnetiferous leucogranite of the Old Rag Granite (Yor; 1,060±5 Ma, table 1, sample 20) has a weak foliation (parallel to the dark line). Dime for scale.

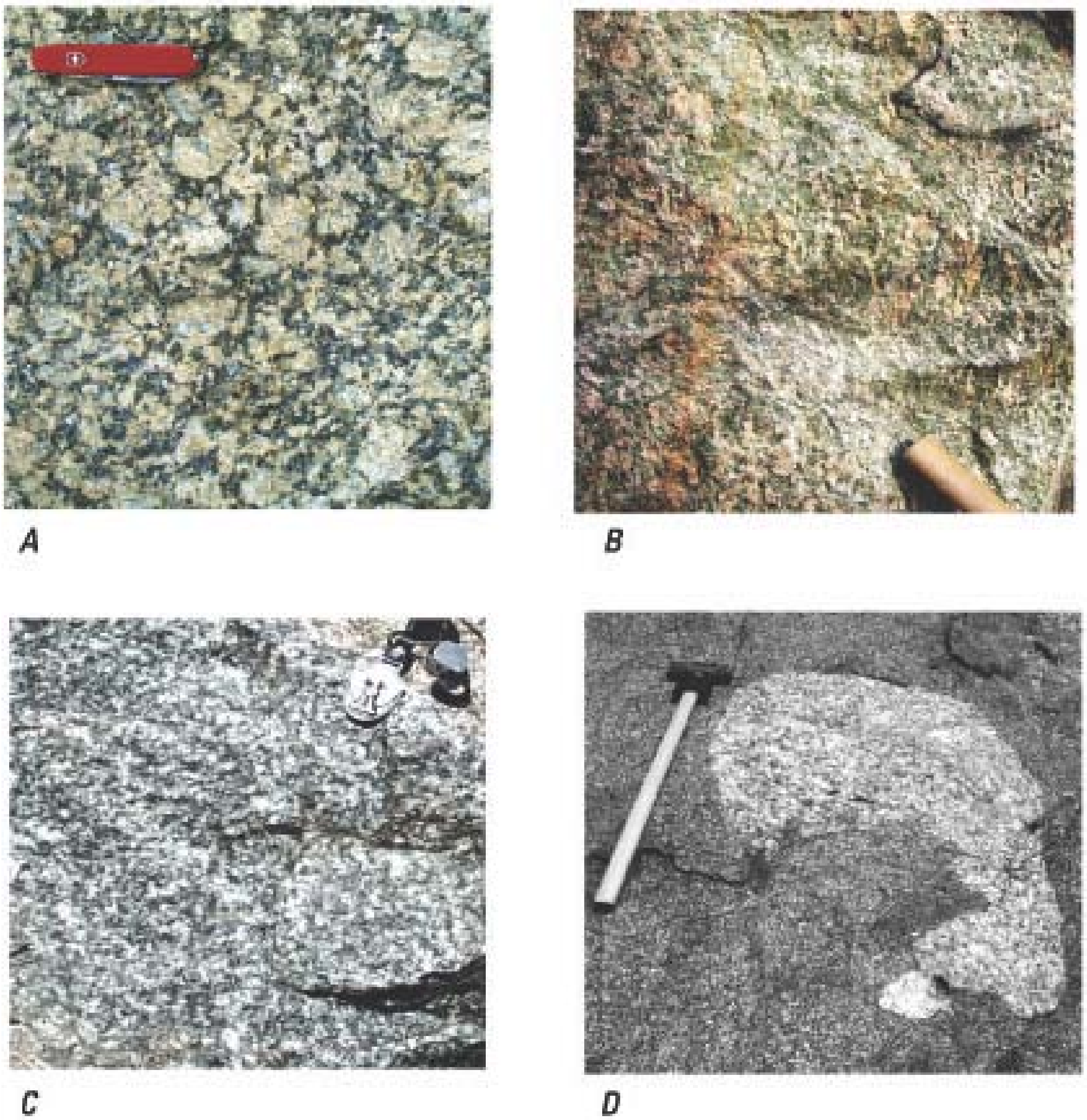

Figure 5. Photographs of Group 3 metagranitoids. *A*, Nonfoliated Crozet Granite (Ycg; 1,060±7 Ma, table 1, sample 21). Pocket knife for scale. *B*, Megacrystic quartz monzonite (Ypb; 1,049±6 Ma) occurs in the extreme northeast part of the map area. The rock is locally altered to unakite, which consists of pink potassium feldspar and green chlorite and epidote. Hammer handle for scale. *C*, Orthopyroxene monzogranite-quartz monzodiorite (Yom; 1,050±8, 1,049±9, and 1,044±6 Ma, table 1, samples 15–17) is a dark, massive, nonfoliated pluton. Hand lens for scale. *D*, Nonfoliated, biotite monzogranite-quartz monzodiorite (Ybg; 1,028±9 Ma) and a xenolith of foliated leucogranite gneiss (Ylg; 1,183±11 Ma). Hammer handle is parallel to the strike of Paleozoic cleavage.

Figure 6. Photograph of Neoproterozoic quartzofeldspathic garnet-graphite paragneiss (Zp) along Buck Mountain Creek. A quartzite bed yielded six detrital zircons that range in age from 1,011 to 997 Ma (table 1, sample 11). Bedding and foliation are parallel and dip to the south (parallel to dark line). Hammer for scale.

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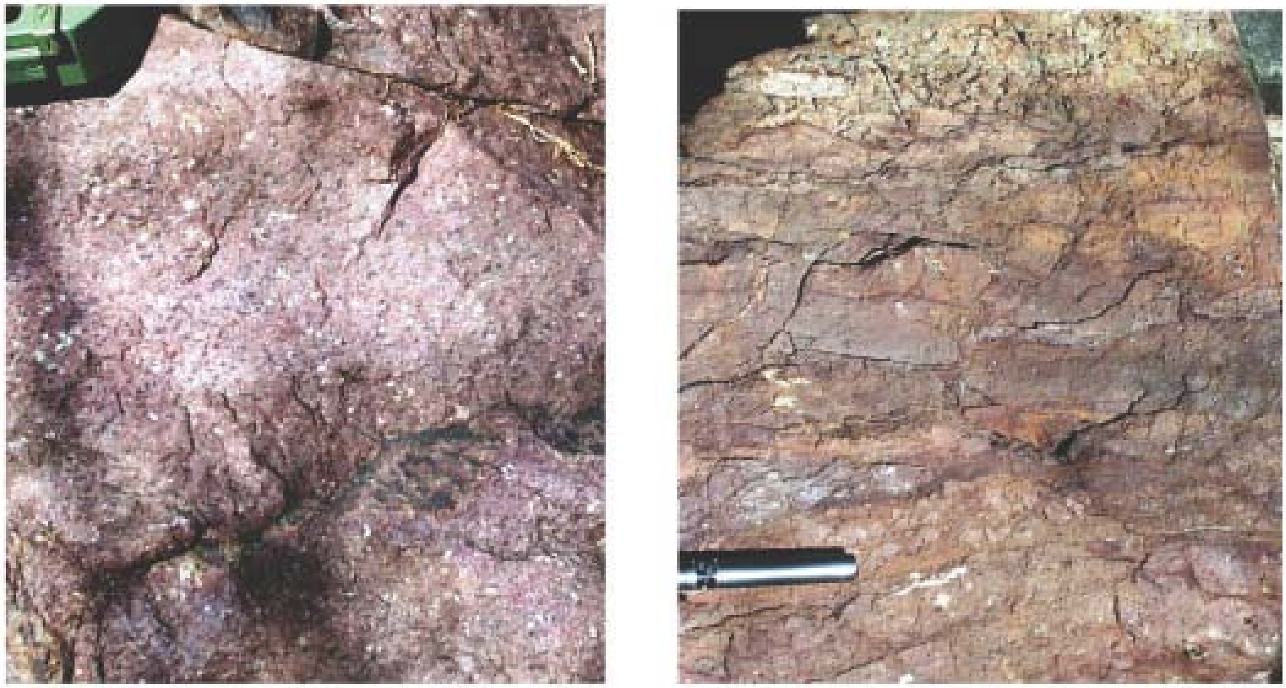


A *B*

Figure 7. Photographs of Neoproterozoic volcanic rocks. *A*, East of the Skyline Drive, between Jenkins Gap and Hogwallow Flat, pink quartz trachyte containing white feldspar phenocrysts (Zrqt; about 719±6 Ma and 714±5 Ma, table 1, samples 5 and 4) overlies Mesoproterozoic orthopyroxene quartz diorite gneiss (Yoq) and orthopyroxene monzogranite-quartz monzodiorite (Yom), and underlies Neoproterozoic Catoctin Formation. Corner of Brunton compass for scale. *B*, Fine-grained, fragmental tuff associated with the quartz trachyte; pen is parallel to layering.

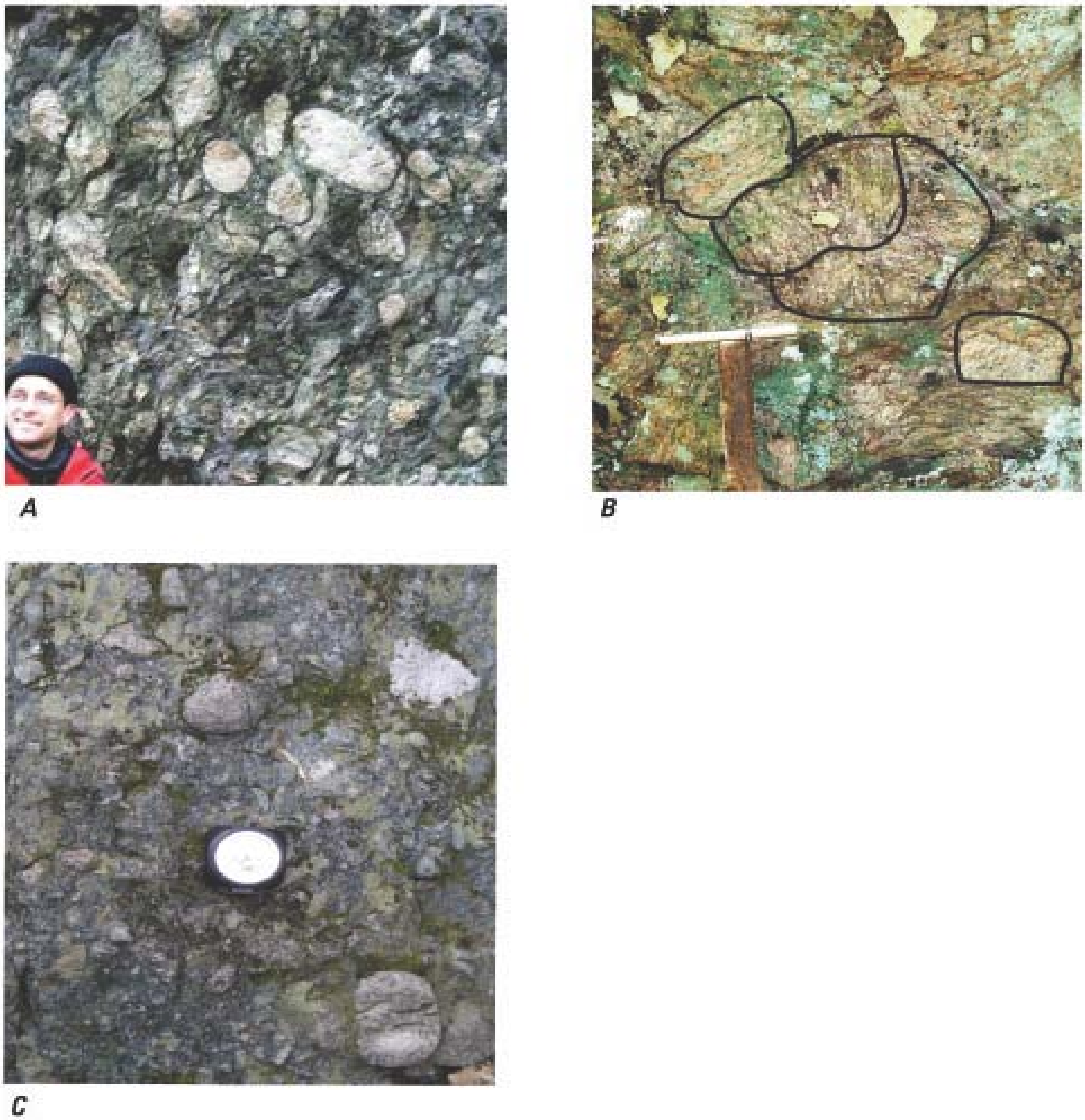
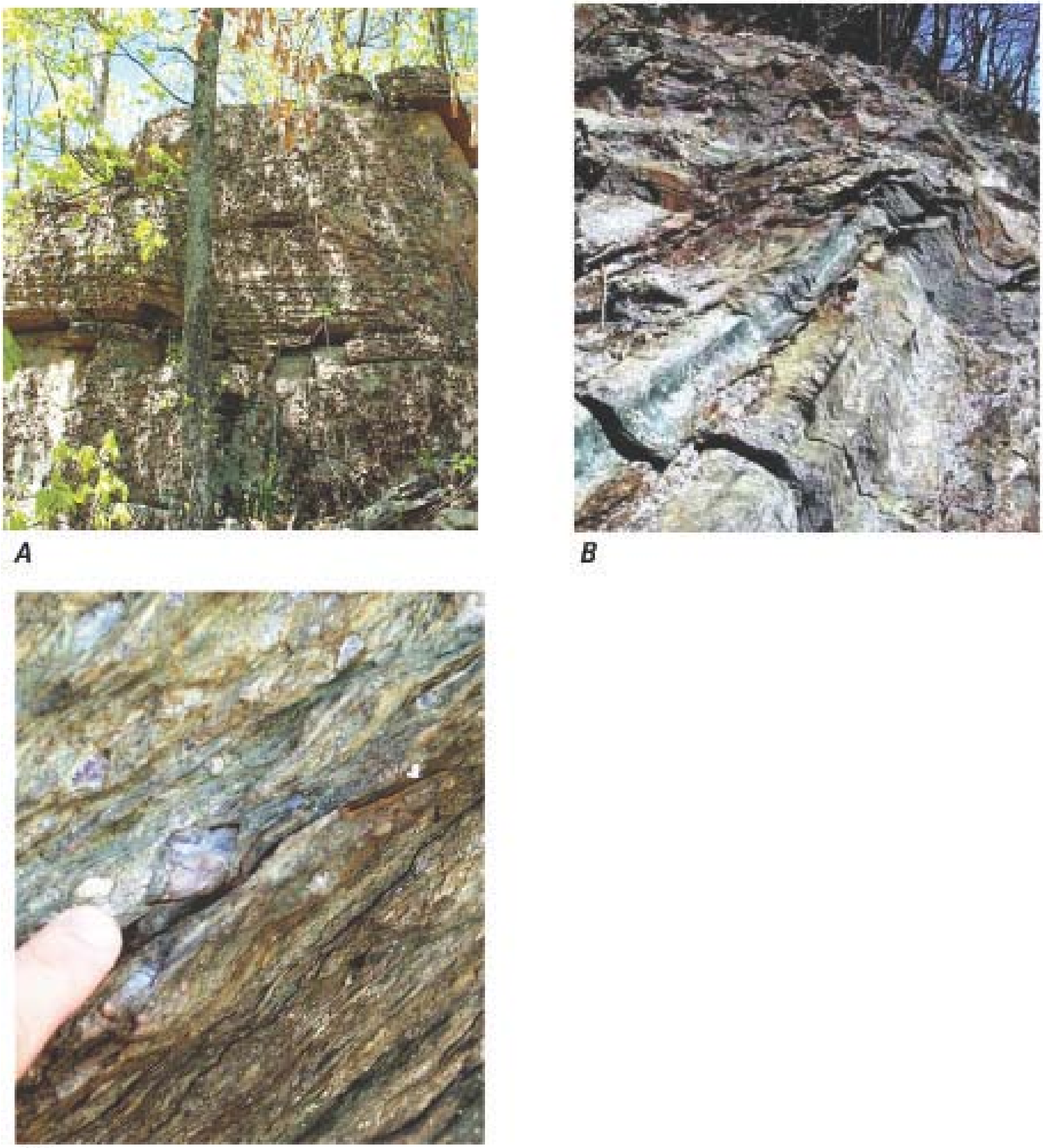


Figure 8. Photographs of Neoproterozoic basal sedimentary rocks. *A*, Conglomerate at the base of the Mechum River Formation (Zmr) exposed on Bessie Bell Mountain (Woodville 7.5-minute quadrangle). Matrix-supported boulders consist mostly of Mesoproterozoic granitoids, but some are derived from the Neoproterozoic Robertson River Igneous Suite. *B*, Boulder bed at the base of the Swift Run Formation (Zsr) in Harris Cove. Clast-supported boulders (outlined in black) consist of unakitic Mesoproterozoic granitoids. Pen for scale. *C*, Matrix-supported pebbles of vein quartz in the basal Swift Run Formation, west of Big Meadows. Brunton compass for scale. (Photograph by Eric Butler, National Park Service).



A B C

Figure 9. Photographs of the Swift Run Formation (Zsr). *A*, Horizontal beds of quartzite in the lower part of the formation on Harris Cove Mountain. *B*, Southeast-dipping cleavage in phyllite is folded with northwest-dipping crenulation cleavage on Powell Mountain. Pen at left for scale. (Photograph by Eric Butler, National Park Service). *C*, Rotated clasts of angular vein quartz in quartz-sericite schist matrix.

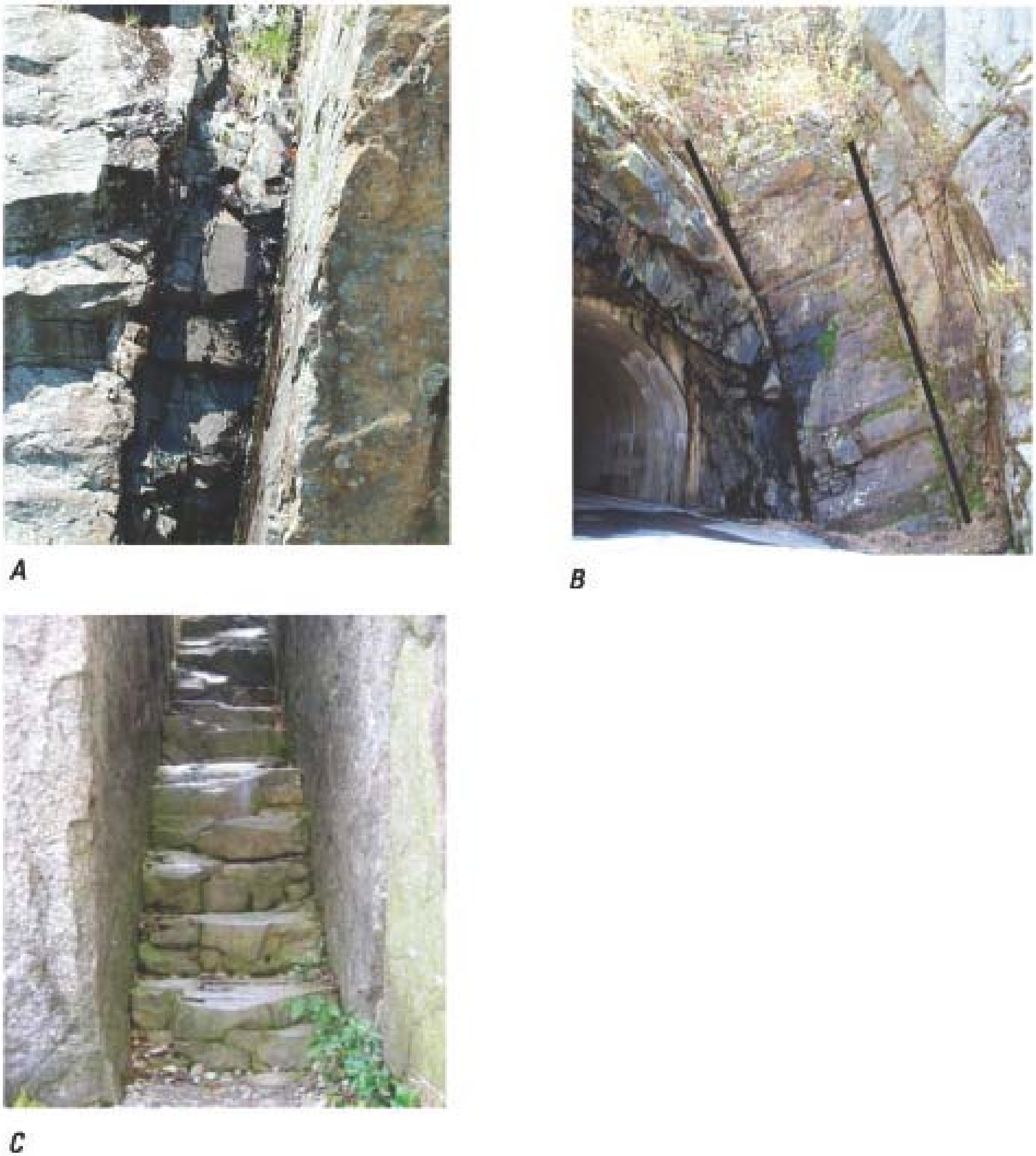


Figure 10. Photographs of Neoproterozoic metadiabase dikes (Zmd). *A*, A 1-m-wide dike intruded orthopyroxene monzogranite-quartz monzodiorite (Yom; 1,044±6 Ma) along Skyline Drive. Groundwater discharges along the contacts. *B*, A 3-m-wide dike in megacrystic orthopyroxene syenogranite-monzogranite gneiss (Yos) at the north portal to the tunnel at Marys Rock strikes northeast and dips northwest. Margins of the dike are shown by dark lines. *C*, A 2-m-wide dike intruded Old Rag Granite on Old Rag Mountain and forms a natural staircase on the trail (Hackley, 2006).

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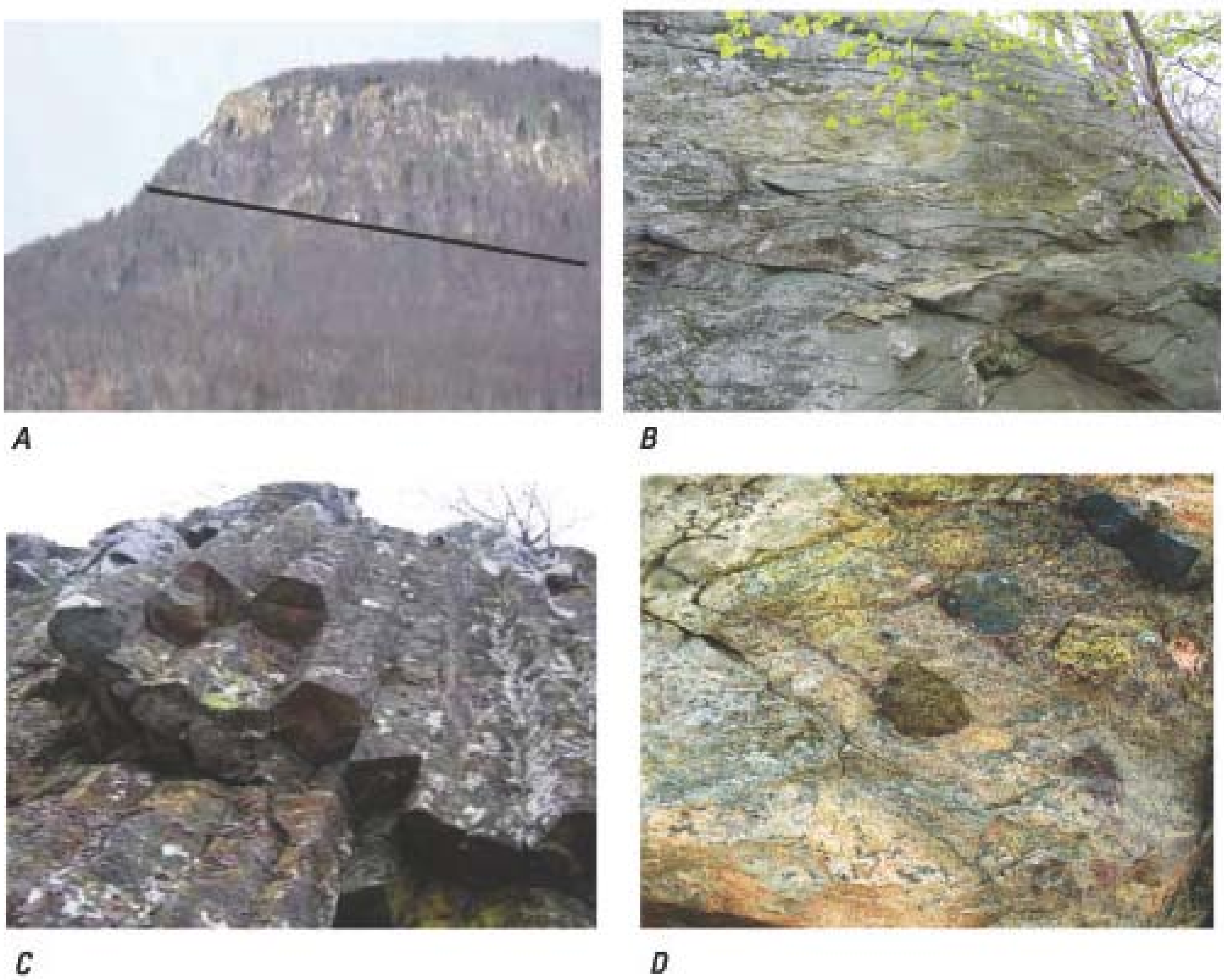

Figure 11. Photographs of Neoproterozoic Catoctin Formation (Zc). *A*, Profile view, looking north at Hawksbill, showing the gentle southeast-dipping homocline of metabasalt flows. *B*, Typical outcrop of "greenstone," which is a metabasalt metamorphosed under greenschist-facies conditions. Pistachio-green layers are epidote and quartz layers, parallel to cleavage, which formed during hydrothermal alteration. *C*, View looking up at the base of columnar-jointed metabasalt on Hawksbill. *D*, Unakitic quartzite in the lower part of the Catoctin Formation contains rounded clasts of metabasalt, granitoid, and vein quartz. (Photographs A, B, and C by Eric Butler, National Park Service).

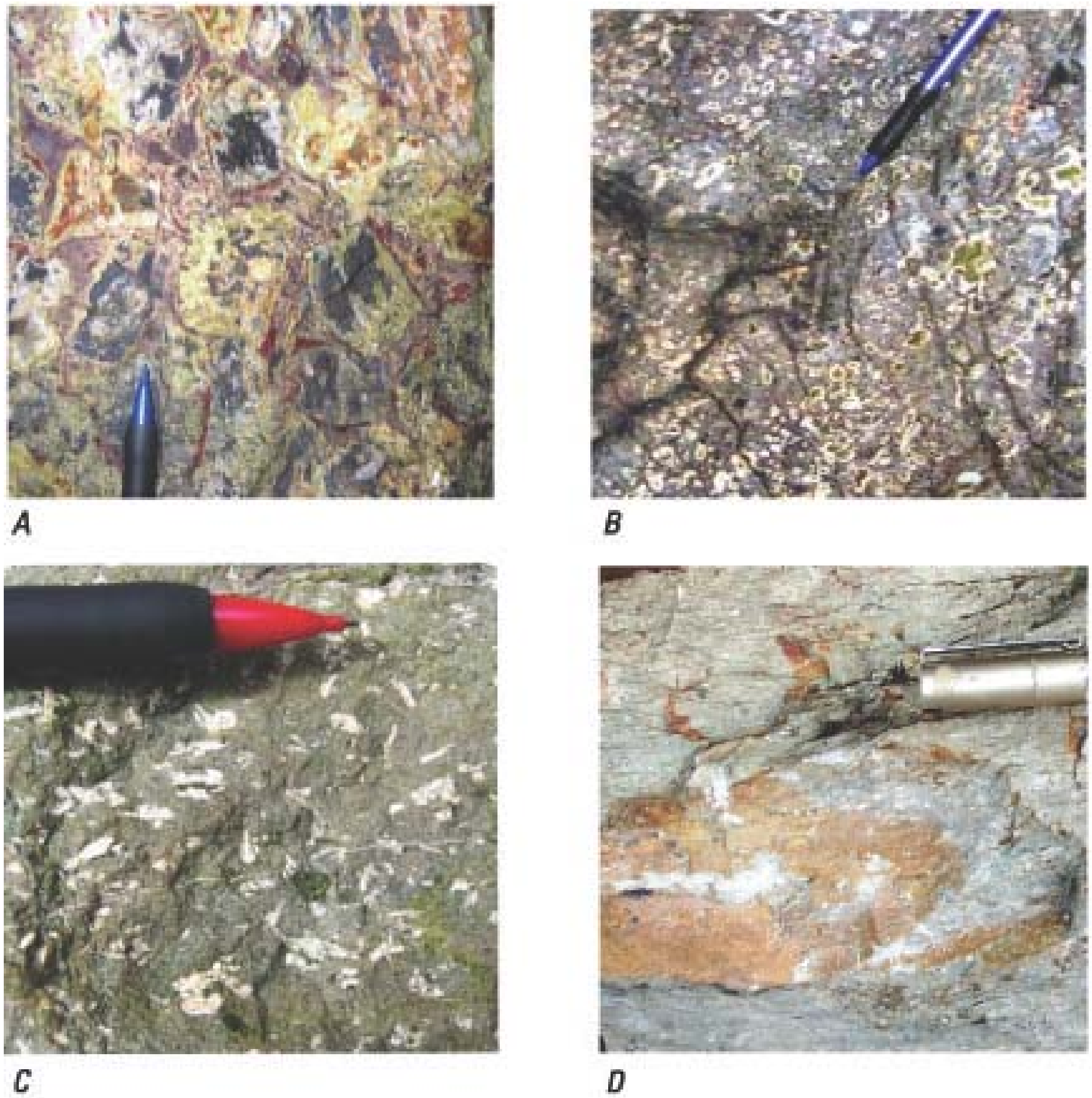

Figure 12. Pens for scale in each photograph. *A*, Fragmental volcanic breccia at the base of metabasalt flows. *B*, Quartz- and epidote-filled amygdules. *C*, White plagioclase laths. *D*, Light-gray metarhyolite with white feldspar phenocrysts (Zcr) within metabasalt near the base of the formation in the extreme northeast part of the map. (Photographs *A*–*C* by Eric Butler, National Park Service.)

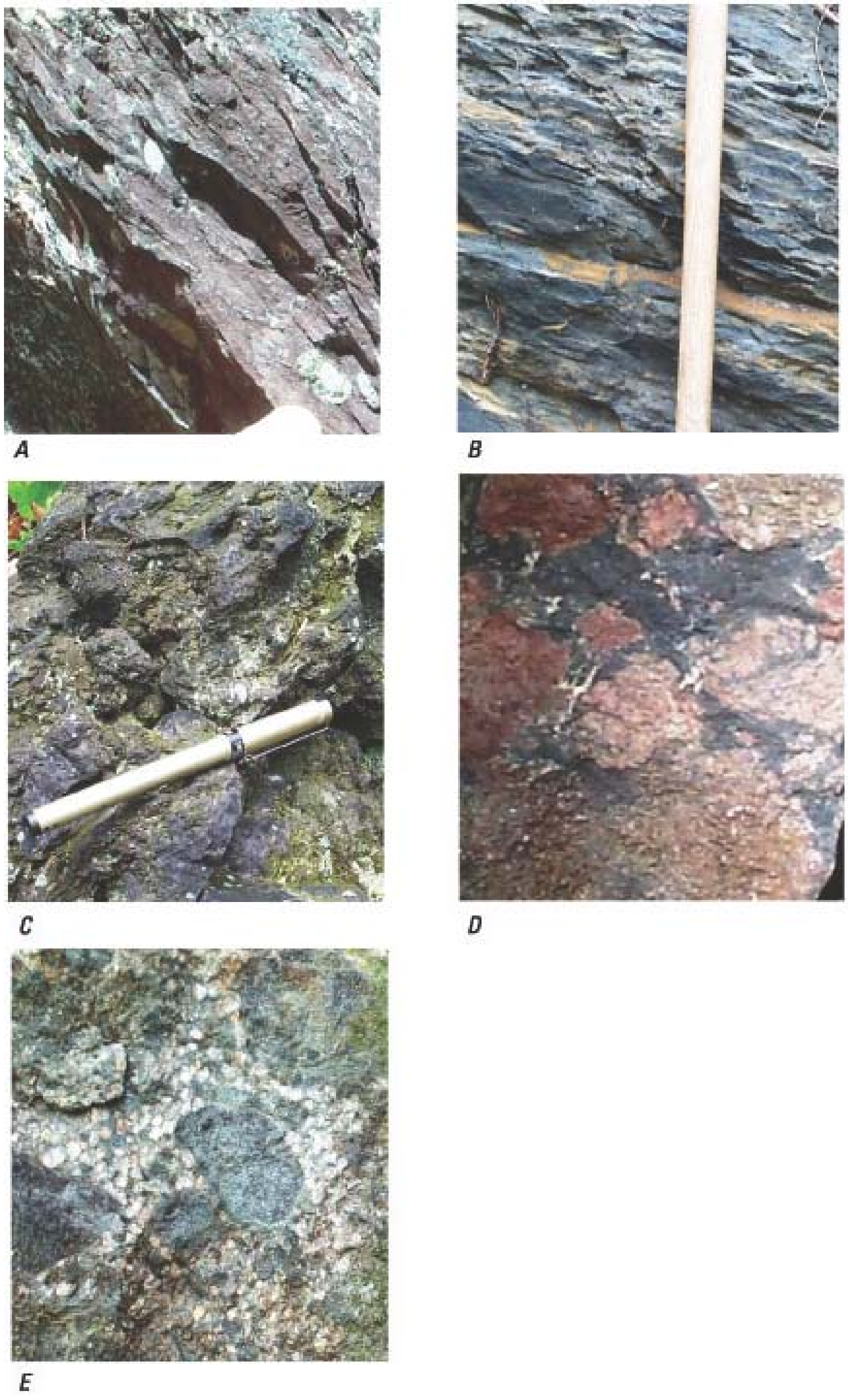

Figure 13. Photographs showing assorted metavolcanic rocks (Zcp) in the upper part of the Neoproterozoic Catoctin Formation. *A*, Red volcaniclastic phyllite with white quartz-filled vesicles (finger points to these) is oxidized and weathered volcaniclastic metabasalt that contains detrital zircons. *B*, The protolith of this dark spotted slate was metasediment derived from a tuffaceous volcanic rock. Hammer for scale. *C*, Angular fragments of scoriaceous metabasalt. Pen for scale. *D*, Rip-up clasts of red scoriaceous, volcanic rock in dark metabasalt matrix. Clasts are about 5 cm across. *E*, Quartz pebbles and rip-up clasts of metabasalt. Clast in center is 4 cm long.

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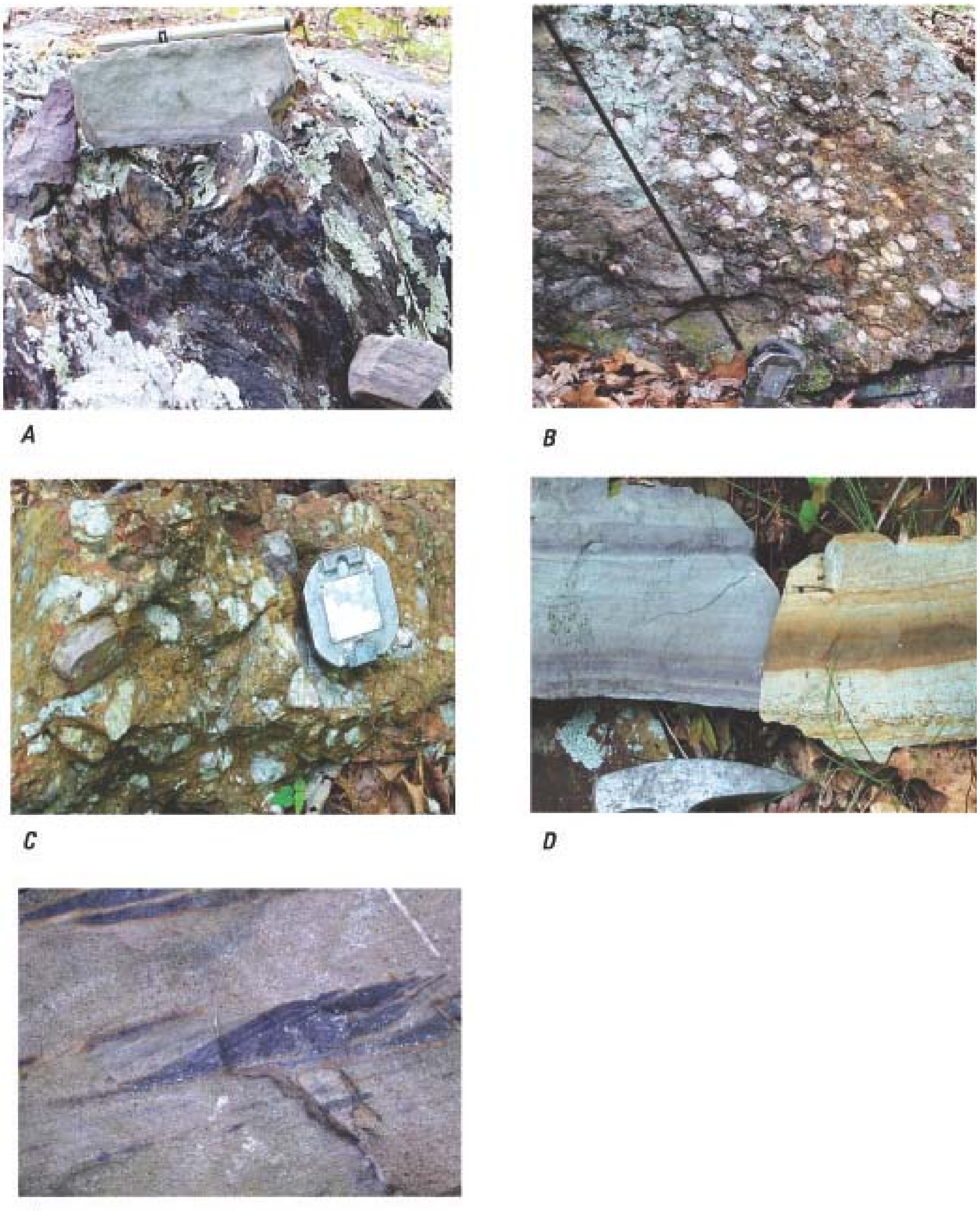

Figure 14. Photographs showing assorted basal rocks of the Cambrian Weverton Formation (Ꞓcw) between Jeremys Run and Overall Run. *A*, Fine-grained crossbedded sandstone overlies purple volcaniclastic rocks of the uppermost Catoctin Formation. Pen for scale. *B*, Basal conglomerate consists of cobbles of mostly vein quartz and minor metabasalt and red jasper. Bedding is overturned and dips to the southeast (parallel to black like). Brunton compass at bottom center for scale. *C*, Clay-rich-matrix-supported conglomerate consisting of vein quartz pebbles. Brunton compass for scale. *D*, Laminated metasiltstones. Hammer head for scale. *E*, Crossbedded quartzite.Dark beds are about 5 cm thick.

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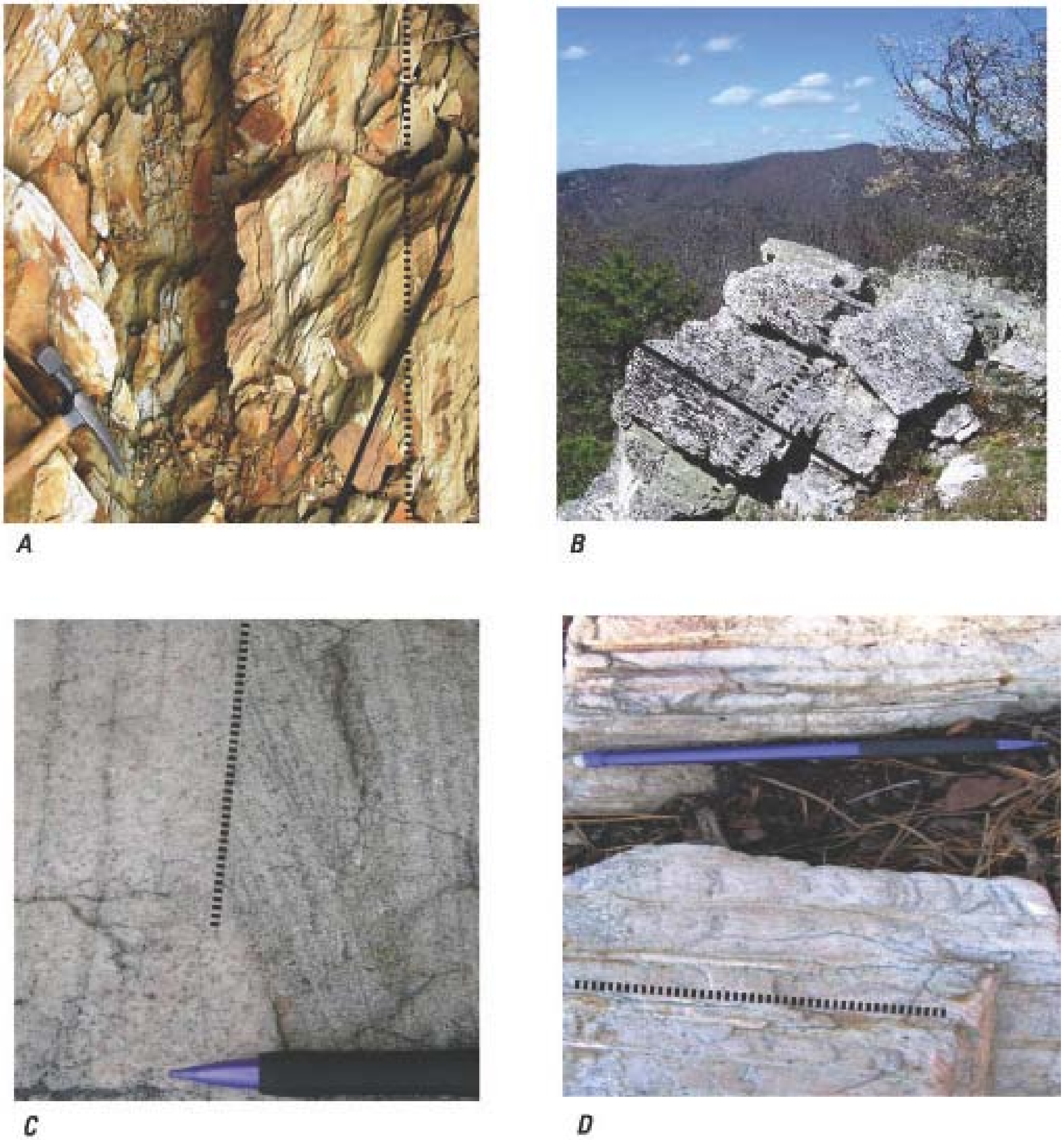

Figure 15. Photographs of the Cambrian Harpers (Ꞓch) and Antietam Formations (Ꞓca). *A*, Metasiltstone of the Harpers Formation. Bedding (indicated by dashed line) is vertical and cleavage (indicated by solid line) dips steeply from upper right to lower left. Hammer for scale. *B*, Outcrop habit of quartzite of the Antietam Formation at Calvary Rocks. Bedding (indicated by dashed line) dips steeply to the west and strong cleavage (indicated by solid line) dips east (to the lower right). *C*, Crossbeds in quartzite of the Antietam Formation are truncated by beds parallel to the near-vertical dashed line. Stratigraphic top is to the left. Pen for scale. *D*, Trace fossil *Skolithos linearis* is a fossil worm burrow that is preserved in the Antietam Formation. Differential erosion makes the infilled tubes stand out in relief. Pen parallel to the tubes provides scale. (Photographs *B–D* by Eric Butler, National Park Service).

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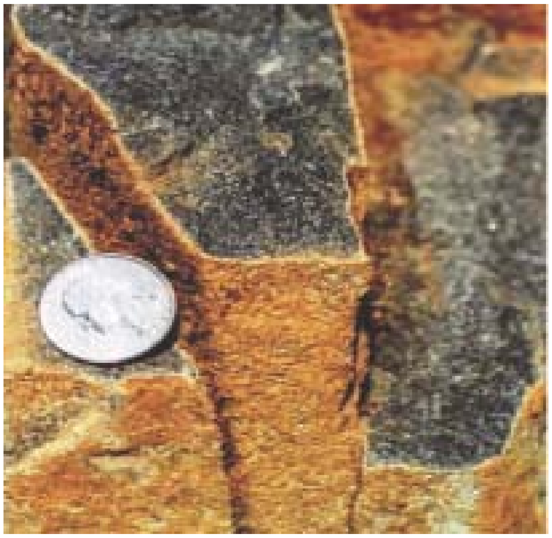

Figure 16. Photograph of Jurassic diabase (Jd) that intruded Mesoproterozoic rocks. The diabase intruded rocks of the Blue Ridge during continental extension that was associated with the opening of the Atlantic Ocean in the Mesozoic Era. The dikes are readily recognized as spheroidally weathered boulders that have a rind of iron-weathered material. Quarter for scale.

A

B

Figure 17. Photographs showing structure in Mesoproterozoic basement rocks. *A*, Mesoproterozoic mylonitic foliation in Flint Hill Gneiss (Yfh) dips to the lower left and was crenulated and kinked by Paleozoic cleavage, which dips to the lower right. Keychain for scale. *B*, Tectonic breccia in orthopyroxene monzogranite-quartz monzodiorite (Yom) along the west of Branch Creek. Brecciation is interpreted to be Neoproterozoic in age as the contacts between basement and cover rocks are not displaced. Outcrop is 3 m wide.

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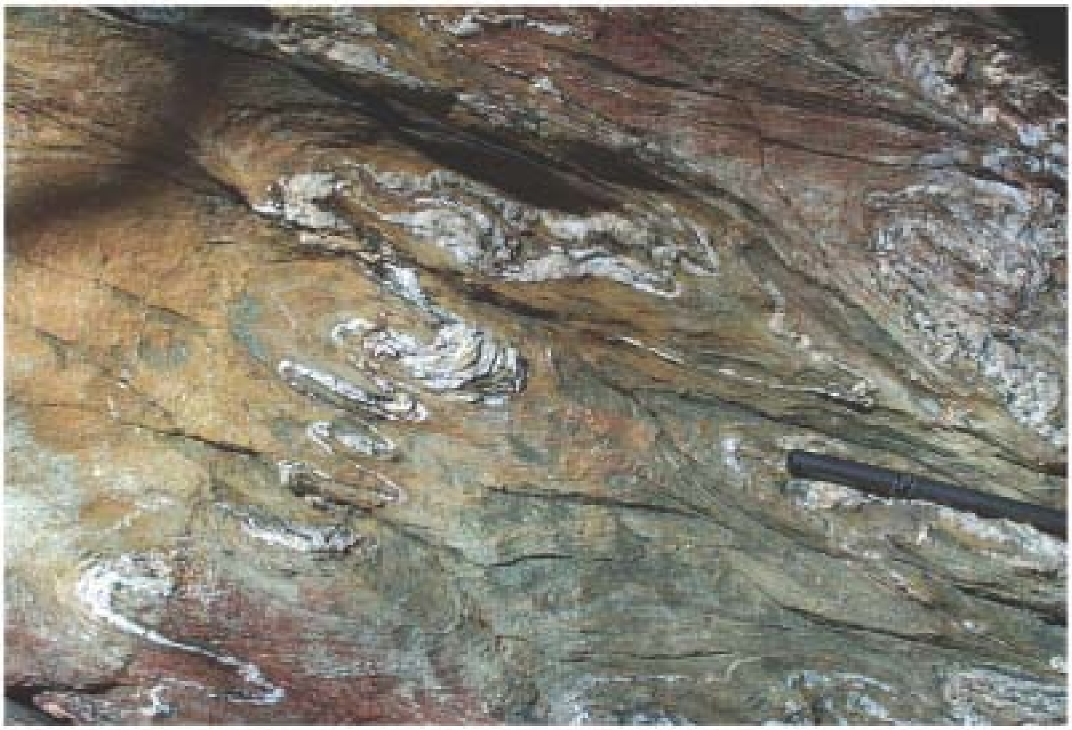

Figure 18. Photograph showing folded seams of vein quartz in quartz-sericite phyllite of the Swift Run Formation (Zsr). In the footwall of the Elkton fault, southwest of Old Rag Mountain. Vein quartz is likely parallel to bedding laminations. Cleavage is axial planar to the folds and dips to the southeast (lower right, parallel to pen used for scale).

Figure 19. Photograph showing sheath folds of Paleozoic mylonitic foliation in the Garth Run high-strain zone. Photograph by Peter Berquist, College of William and Mary.

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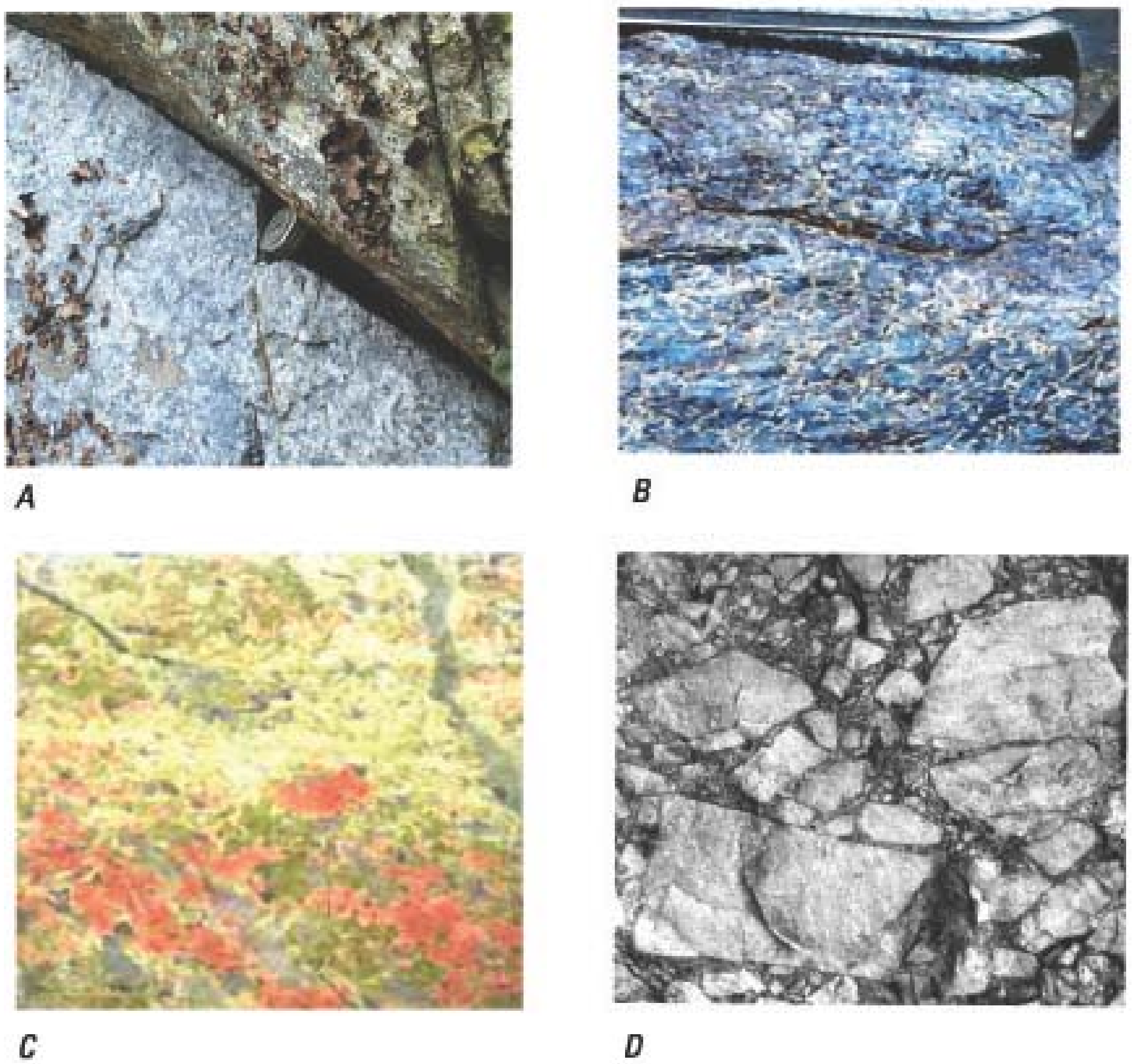

Figure 20. Photographs showing interesting rocks and minerals in the Shenandoah National Park region. *A*, Outcrop of a blue quartz vein that intruded gneiss. Brunton compass (for scale) located on contact, just right of center. *B*, Close-up of part *A* showing texture of blue quartz. The blue color is due to titanium in the quartz. Hammer for scale. *C*, Unakite, a byproduct of hydrothermal alteration, is an ornamental stone consisting of pink potassium feldspar and green epidote. Unakite is the official State rock of Virginia. *D*, Breccia consisting of quartzite from the Antietam Formation (light fragments) cemented with manganese. Quartzite fragments are 2 cm long. Photograph by T. M. Gathright, II.

A

B

Figure 21. Photographs showing surficial deposits and landforms. *A*, Exposure of an alluvial-fan deposit (Nf) along Naked Creek. A 5-m-thick layer of quartzose gravel overlies residuum derived from dolomite in the Tomstown Formation. *B*, View looking southwest into the modern Thornton River alluvial valley at Sperryville.

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Figure 22. Photograph of a debris-fan deposit (Nd) near Graves Mill. The old fan deposit consists of very poorly sorted large boulders (see jeep for scale) and cobbles of granitoids in a fine-grained matrix of sand, silt, and clay. Photograph by Gerry Weiczorek, USGS.

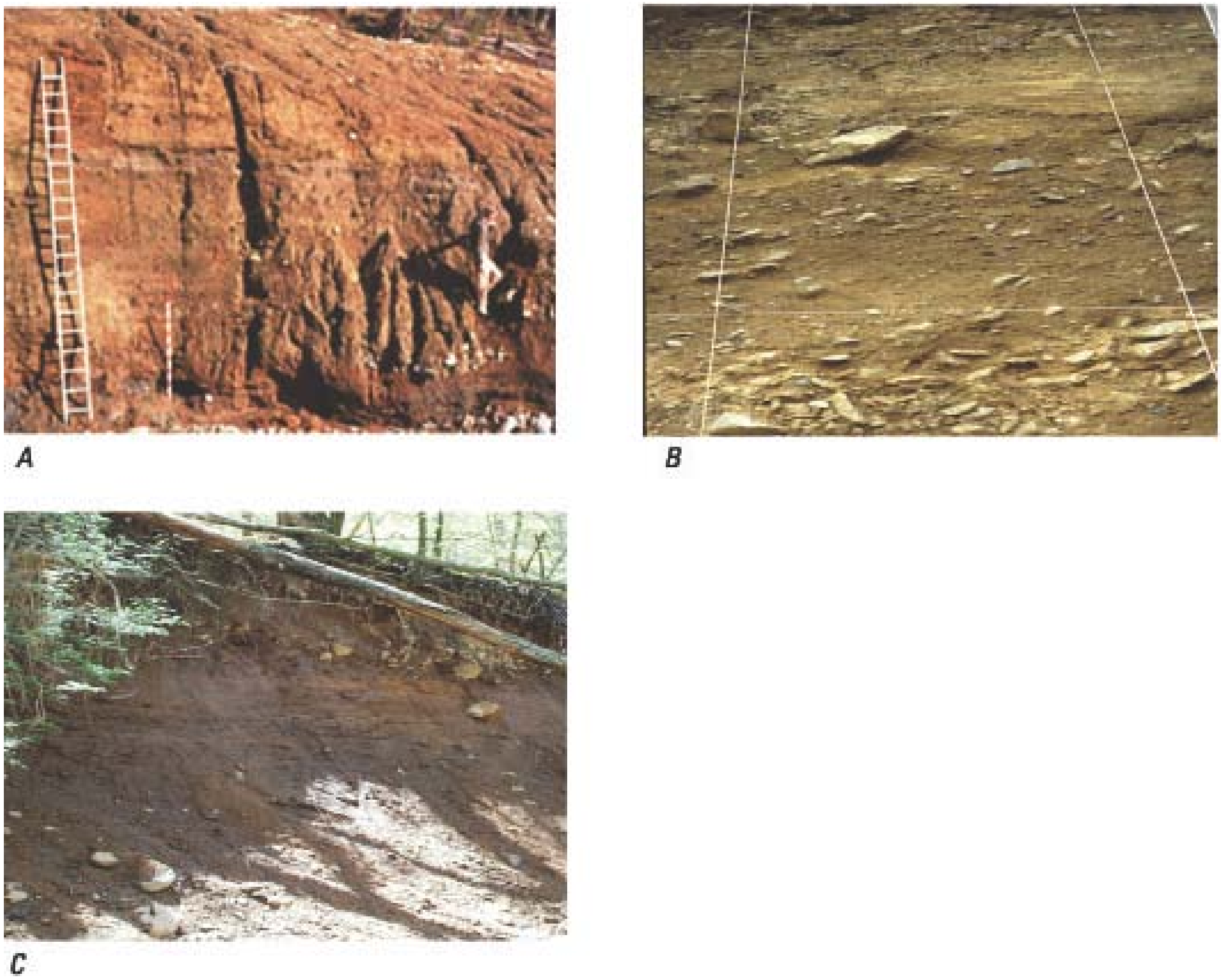

Figure 23. Photographs showing stratified slope deposits. *A*, Stratified slope deposit exposed by the 1995 debris flow along Kinsey Run. Photograph by Gerry Weizcorek. *B*, Matrix-supported diamictite of a stratified slope deposit along the channel of Rapidan River at an altitude of about 700 m. The 5-m-thick section was exposed by floods. The section is beneath a deposit of boulders in the floor of a hollow (Smoot, 2004). *C*, Stratified slope deposit adjacent to Naked Creek.

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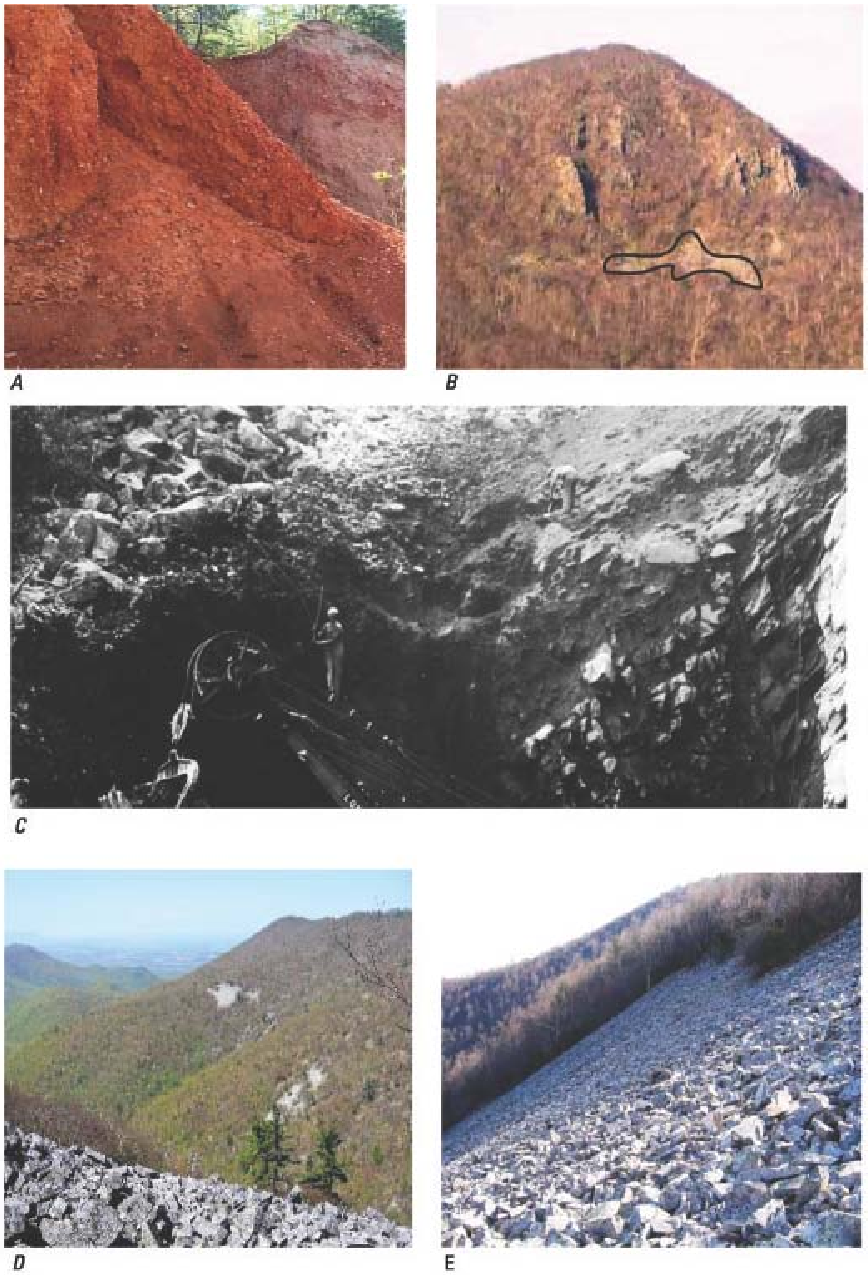

Figure 24. Photographs showing types of colluvium (Qc). A, Colluvium consisting of pebble- to cobble-sized fragments of quartzite of the Antietam Formation in a red residuum derived from carbonate rocks of the Tomstown Formation. The colluvium mantles most of the steep slopes and forms aprons along the west slope of the ridges that are underlain by the Chilhowee Group adjacent to the Page Valley. *B*, In the highlands, colluvium consists mostly of subangular clasts and boulders of metabasalt that are locally derived from outcrops, such as on the north side of Hawksbill (outlined area). *C*, Diamictite and colluvium uncovered during construction above the tunnel at Marys Rock, at an elevation of 800 m. Photograph courtesy of the NPS. *D*, Quartzite block-field deposits (Qb) are well developed on the ridges underlain by the Chilhowee Group in the southwestern part of the Shenandoah National Park area. *E*, Lichen-covered subangular blocks of quartzite derived from the Antietam Formation make up this quartzite block-field deposit (Qb).

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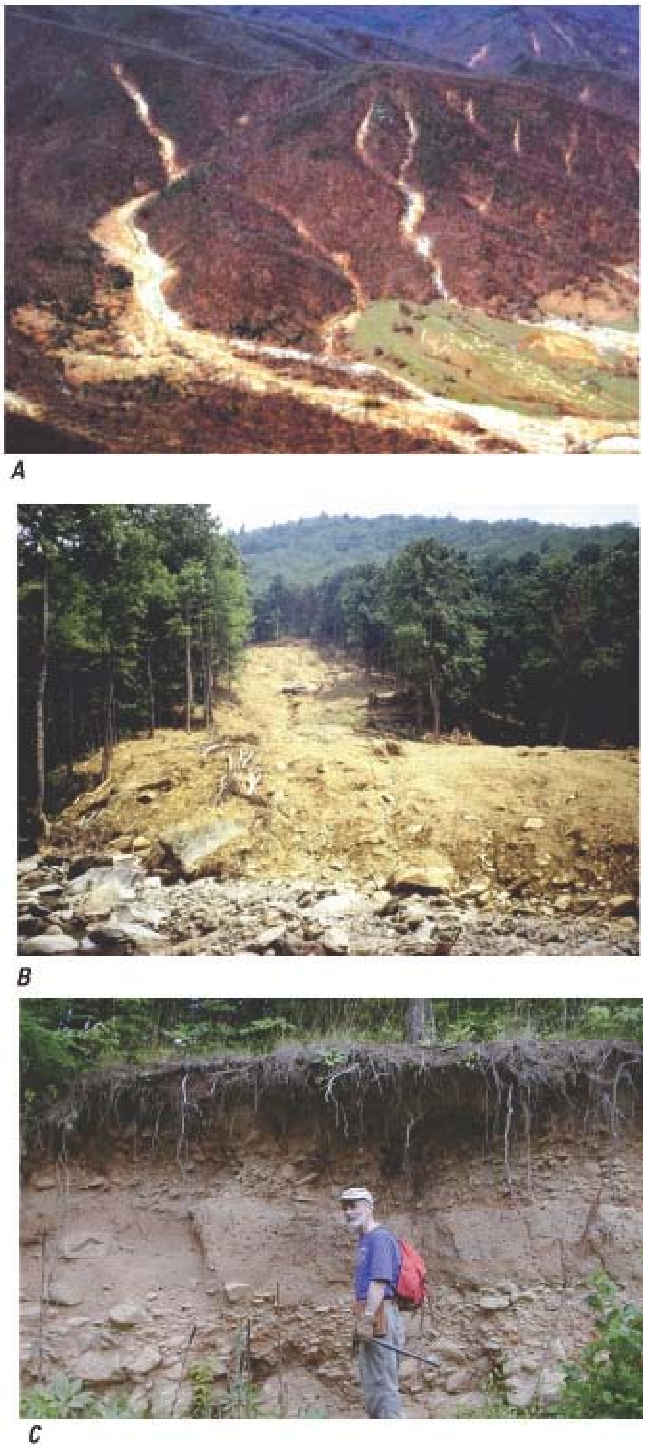

Figure 25. Photographs showing debris-flow deposits (Qdf). *A*, Deposits resulting from a storm in 1995, which caused debris flows and floods in the area between the Robinson and Conway Rivers in Madison County. *B*, A close-up of the debris-flow chute seen in the extreme left part of the view in part *A*. Masses of bedrock, soil, and vegetation moved rapidly downslope and incorporated more material as they descended. *C*, Close-up of ancient stratified debris-flow deposit exposed by debris flows in 1995. (Photographs by L. Scott Eaton, James Madison University).