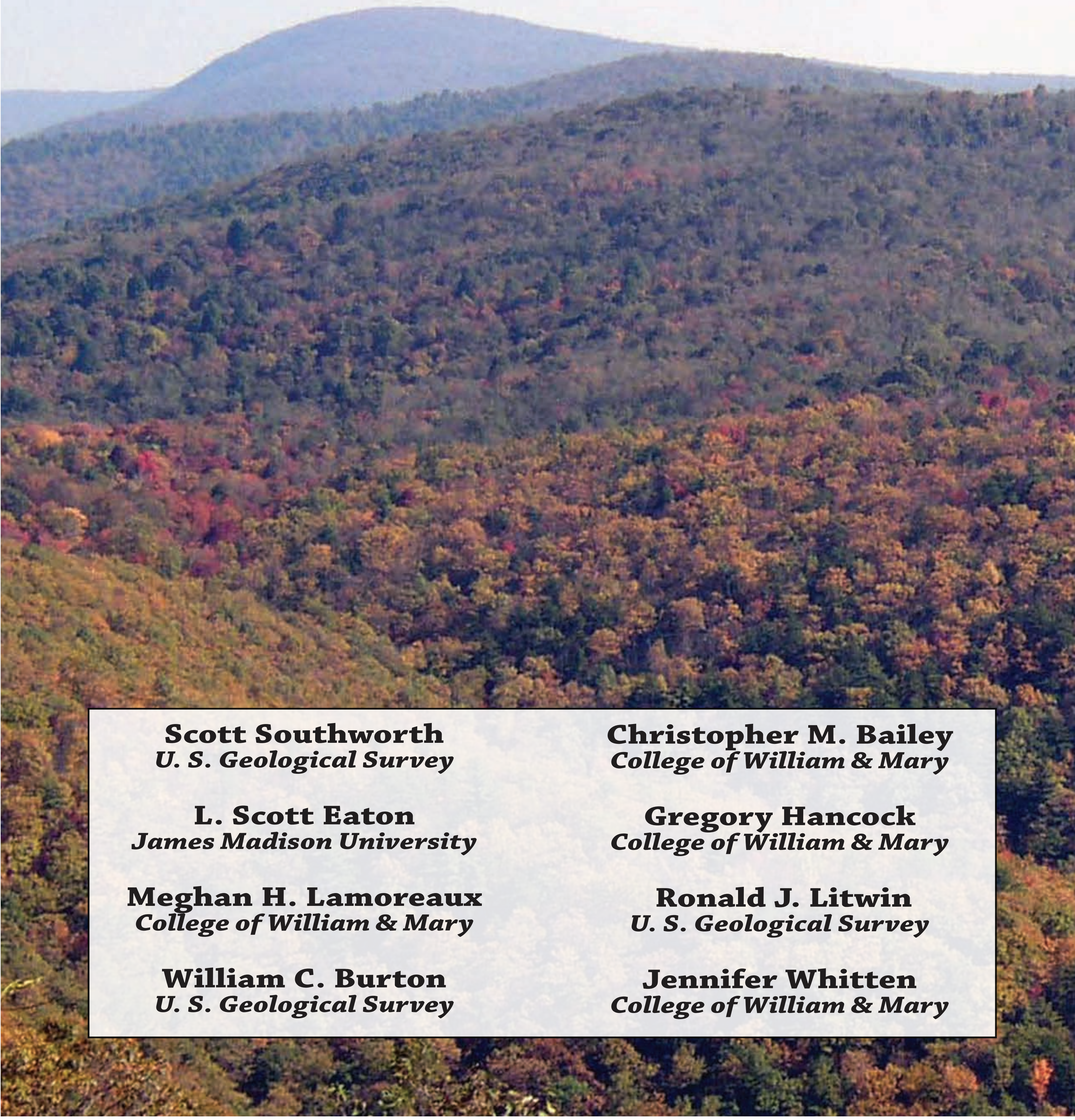
39th Annual Virginia Geological Field Conference
October 2nd - 3rd, 2009

Geology of the Shenandoah National Park Region



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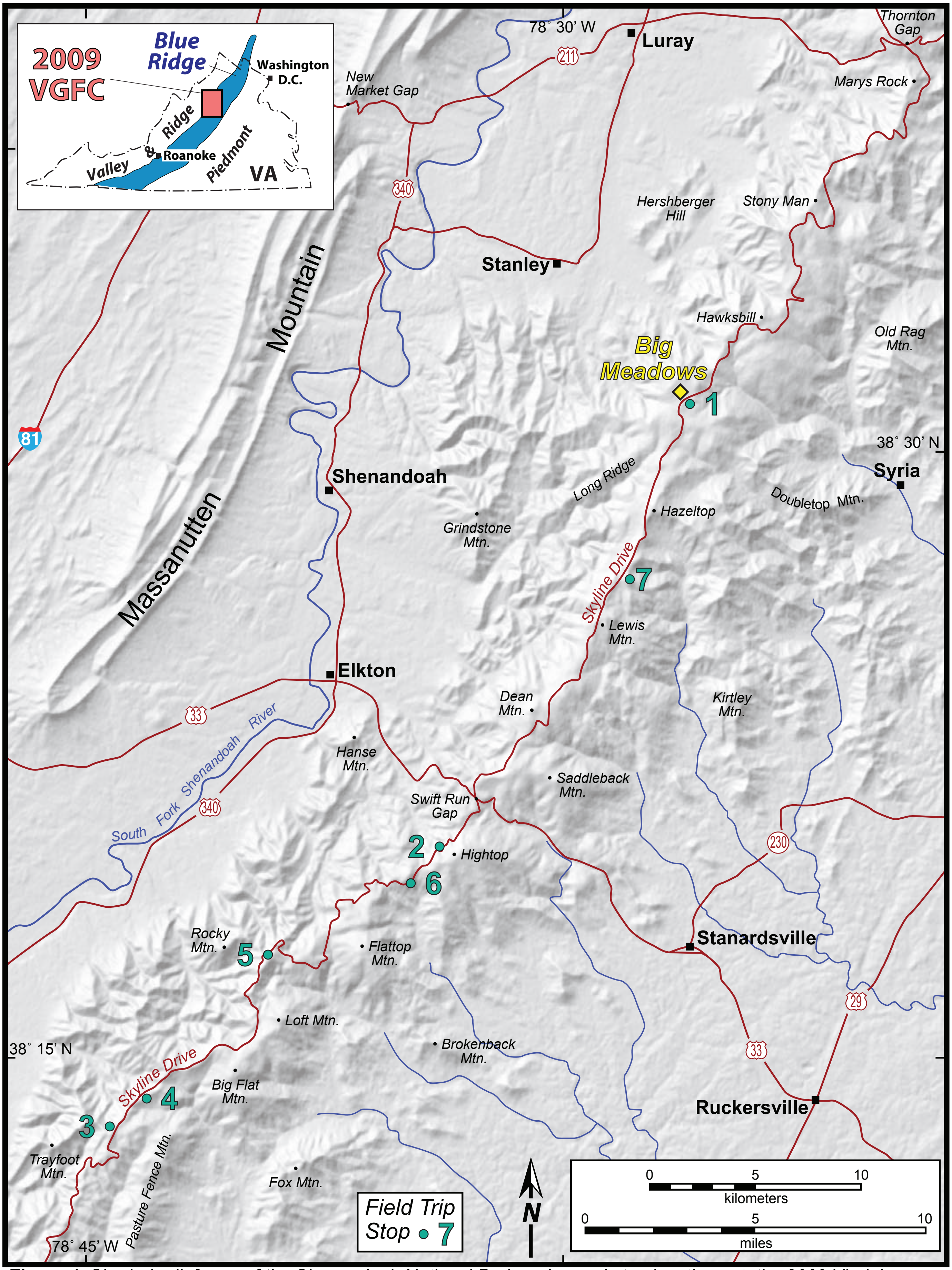
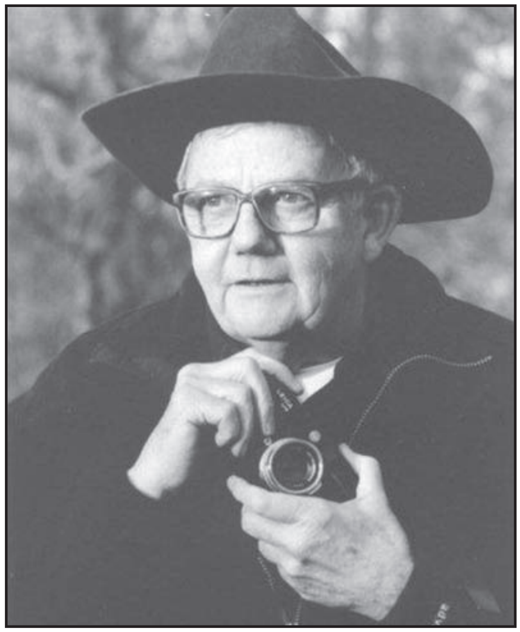


Figure 1. *Shaded relief map of the Shenandoah National Park region and stop locations at the 2009 Virginia Geological Field Conference.*

Dedication

The 39th annual Virginia Geological Field Conference is dedicated to Tom Gathright whose work in western Virginia and Shenandoah National Park has educated and inspired many of us.



Tom Gathright joined the Division of Mineral Resources in 1964. Originally a member of the Ground water Section, Tom conducted groundwater investigations in the Blue Ridge and Valley & Ridge provinces. During this time, he helped to locate and develop several major water supplies along the west flank of the Blue Ridge. Tom's first project with the Geologic Mapping Section was to create a geologic map of Shenandoah National Park. This work was published in 1976 as Bulletin 86, "Geology of the Shenandoah National Park, Virginia." It remains one of the most popular reports related to Virginia geology. Tom subsequently completed geologic maps of ten additional 7.5-minute quadrangles and helped to compile three geologic maps of three 30- x 60-minute quadrangles. As the head of the geologic mapping section, he oversaw the completion of geologic mapping of the southwest Virginia Coalfields. Tom retired in 1991, but continues to work on an occasional basis as a geologic consultant and serves as a member of Virginia's geologic mapping advisory committee.

Tom framing a photograph on a crisp morning at Rockfish Gap.

photos courtesy of Elizabeth Campbell (top) and Gerry Wilkes (bottom).

Introduction

Shenandoah National Park and its magnificent Skyline Drive lie astride the Blue Ridge Mountains in north-central Virginia. Occupying more than 800 km^2, Shenandoah is a long and narrow park with a highly irregular boundary. Established during the Great Depression, the Park was intended to serve as a leafy refuge for urbanites in eastern North America. Created from a patchwork of privately owned farms, orchards, and home sites, the Park has been reforested in the past 80 years with much of the region returned to a wilderness state. In modern times, the Park welcomes over two million visitors per year.

The Blue Ridge Mountains in Shenandoah National Park form a distinctive highland that rises to elevations above 1,200 meters (~4,000') with local topographic relief exceeding 900 meters (~3,000'). The crest of the range forms a drainage divide separating the Shenandoah (Potomac) River drainage to the west, in the Great Valley subprovince of the Valley & Ridge, from southeast-flowing streams of the James and Rappahannock river systems coursing into the foothills region of the Piedmont physiographic province (Fig. 1).

Although a cloak of forest and mantle of soil commonly obscures the underlying bedrock, Shenandoah's topography is dictated by its geology. Rocks exposed in the Blue Ridge Mountains are among the oldest in Virginia and bear witness to more than a billion years of magmatism, sedimentation, sea level oscillation, climate change, tectonic activity, and erosion. Bedrock includes a suite of Grenvillian basement rocks, metamorphosed Neoproterozoic sedimentary and volcanic rocks, and early Cambrian siliciclastic rocks. The Park is situated along the western margin of the Blue Ridge anticlinorium, a regional-scale Paleozoic structure developed at the hinterland edge of the Appalachian fold and thrust belt. The Blue Ridge highlands are the product of differential erosion in the Cenozoic, but post-Paleozoic tectonic activity has influenced the character of the Blue Ridge landscape in discernible ways. Quaternary surficial deposits are common throughout the Park, providing a record of past climate regimes as well as active processes shaping the modern landscape.

Aspects of the Park's geology have been studied since the 1930's (Furcron, 1934; Jonas and Stose, 1939; King, 1950) and many of the counties that encompass the Park were mapped after World War II (Rockingham- Brent, 1960; Albemarle- Nelson, 1962; Greene and Madison- Allen, 1963; Page- Allen, 1967). The first comprehensive treatment of Shenandoah's geology was undertaken by the Virginia Division of Mineral Resources in the late 1960s and culminated in the 1976 publication and geologic map (1:62,500 scale) by Tom Gathright. Gathright's seminal work summarized many of the earlier studies and provided a unifying framework of the Park's geology that is still informative in the 21st century. The hydrogeologic setting of the Park is discussed by DeKay (1972).

Robert Badger's *Roadside Geology of the Skyline Drive* (1999) is a well-illustrated guide aimed at a broad audience. Tollo and others (2004), Eaton and others (2004), and Bailey and others (2006) provide technical field guides with stops in and near the Park. Web resources illustrating the geology of Shenandoah National Park include: William & Mary's *Geology of Virginia/Google Earth* site (*http://web.wm.edu/geology/virginia/ge.php*), James Madison's *Geology of Virginia* site, (*http://csmres.jmu.edu/geollab/vageol/vahist/*) and Callan Bentley's (Northern Virginia Community College) *Geology of Shenandoah National Park* site (*http://www.nvcc.edu/home/cbentley/gol_135/shenandoah/index.htm*).

In the past fifteen years, a multitude of studies focusing on the geology of Shenandoah National Park have contributed new insights about 1) the complexity and chronology of the basement, 2) the structural geometry and deformation history of the region, and 3) Quaternary surficial processes and their efficacy. Nineteen 7.5' quadrangles in and around Shenandoah National Park have been mapped (or remapped) at 1:24,000 scale in the last decade (Fig. 2). A new 1:100,000 scale geologic map of the Shenandoah National Park region and an accompanying report, compiling much of this research, was just published by the U. S. Geological Survey (Southworth and others, 2009; online at *http://pubs.er.usgs.gov/usgspubs/ofr/ofr20091153*).

The 2009 Virginia Geological Field Conference will highlight new research, visiting exposures along and near the Skyline Drive. The trip will be broad, encompassing bedrock and structural geology as well as surficial geology and rates of erosion at local and regional scales. Shenandoah National Park is too vast for a single day traverse; thus the Conference will focus on the geology between Big Meadows (milepost 51) and Blackrock Summit (milepost 85) in the central and southern part of the Park. In addition to emphasizing new findings in Shenandoah we hope this trip will foster discussion about unanswered aspects about Blue Ridge geology and encourage future research in the Park.

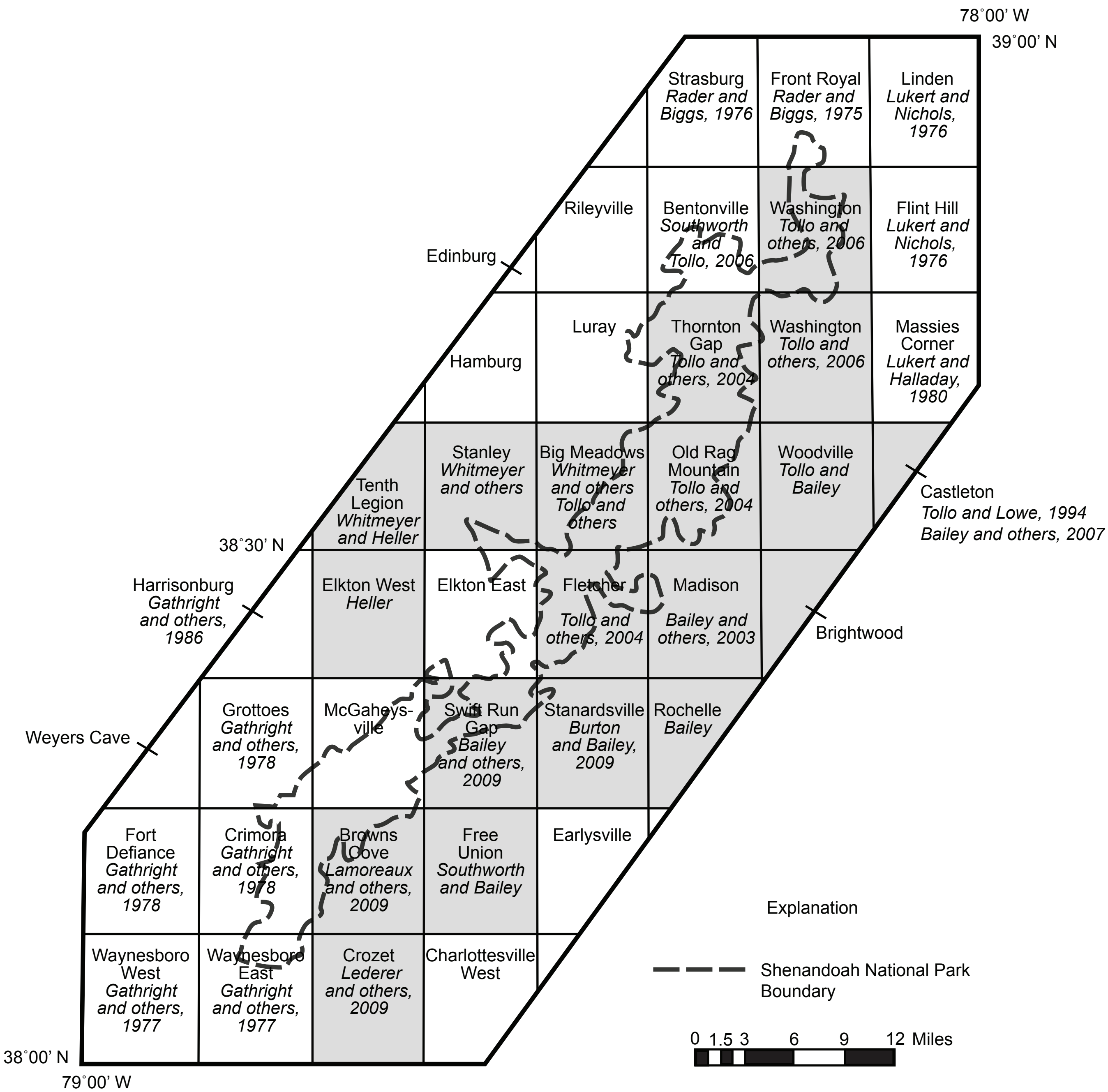


Figure 2. *Index map of 7.5 minute quadrangles surrounding Shenandoah National Park. Authors noted inside quadrangles mapped at 1:24,000 scale. Grey shading indicates recent work.*

Bedrock Geology

Stratigraphy

Shenandoah National Park is underlain by three major geologic units: 1) Mesoproterozoic gneisses and granitoids that comprise the basement, 2) Neoproterozoic metasedimentary and metavolcanic rocks of the Swift Run and Catoctin formations, and 3) siliciclastic rocks of the Early Cambrian Chilhowee Group (Figs. 3 and 4). Basement rocks crop out in the eastern Blue Ridge Mountains and the adjoining foothills, but also underlie peaks such as Old Rag Mountain, Mary's Rock, and Roundtop. The Catoctin Formation forms much of the Blue Ridge's high crest including Hawksbill, Stony Man, Mount Marshall, and Hightop. The Chilhowee Group, exposed in the western Blue Ridge, underlies steep mountains and ridges mantled with thin soil and abundant talus such as Grindstone, Rocky, Trayfoot, and Turk mountains.

Basement rocks include granitoid gneisses and granitoids formed during the Grenville orogeny between 1.2 and 1.0 Ga (Fig. 3). Prior to the 1980s rock units in the Blue Ridge basement were commonly mapped as formations and Gathright (1976) mapped two basement units, the Pedlar Formation and the Old Rag Granite, in the Park. Most technical studies published in the past two decades (Rader and Evans, 1993; Southworth and others, 2000; Bailey and others, 2003; Tollo and others, 2004b; Southworth and others, 2009) avoid using the formation terminology; rather Mesoproterozoic basement units are distinguished based on rock type, cross cutting relations, geochemistry, and geochronology. Many general interest publications and websites continue to use the formational lexicon, this is unfortunate because the rich geologic history of the basement is diluted (or worse still, misunderstood) using archaic formation names. For instance, recent mapping and geochronology in Shenandoah National Park demonstrates that the Pedlar Formation contains over 12 different units that that vary in age by ~150 million years (Southworth and others, 2009).

Modern U-Pb zircon geochronology reveals that the basement complex is comprised of three temporally distinct groups of igneous rocks emplaced over a 150 million year interval in the Mesoproterozoic (Fig. 3) (Aleinikoff and others, 2000; Tollo and others, 2004b; Southworth and others, 2009). The oldest group crystallized between 1,190 and 1,150 Ma and are granitoid gneisses with a compositional layering that developed under high-grade conditions. A volumetrically minor group of orthopyroxene-bearing granites crystallized between 1,120 and 1,110 Ma. The youngest group of granitoids was emplaced between 1,090 and 1,020 Ma. Both the older and younger groups are chemically diverse and include pyroxene-bearing charnockitic rocks and alkali feldspar leucogranitoids. Tollo and others (2004) note that the basement suite was derived from melting of lower crustal sources in an intraplate setting.

The Blue Ridge basement is almost entirely of igneous origin (orthogneisses), however recent mapping has identified small (<1 km^2) bodies of quartzite and garnet-graphitic gneisses surrounded by orthogneisses and granitoids (Burton and Southworth, 1993; Southworth and others, 2007; Lederer and others, 2009) to the east of the Park. The mineralogy and texture of these rocks is consistent with arkosic to greywacke protoliths, and as such these rocks are paragneisses. These paragneissic inliers were originally interpreted as blocks of older crust into which the orthogneisses intruded (Southworth and Burton, 1993). Rounded and pitted zircons, interpreted as detrital grains, from quartzite layers at three locations yield U-Pb ages between 1,010 and 810 Ma, indicating that the sedimentary protoliths for these paragneisses postdate the youngest suite of granitoids (Southworth and others, 2008). If these data are accurate, there is a poorly understood episode of early Neoproterozoic (Torridonian?) unroofing and sedimentation preserved in the Virginia Blue Ridge.

In the central and eastern Blue Ridge anticlinorium, the 730 to 700 Ma Robertson River plutonic complex intrudes the Mesoproterozoic basement and a thick sequence of Neoproterozoic metasedimentary rocks (Lynchburg, Fauquier and Mechum River units) overlies the older rocks (Tollo and others, 1996, 2004c; Bailey and others, 2007). These units are generally absent in Shenandoah National Park.

Recent mapping in the northeastern part of the Park reveals thin, discontinuous layers of felsic volcanic rock unconformably overlie the Mesoproterozoic basement and beneath the Catoctin Formation that yield U-Pb zircon ages of 720 to 710 Ma (Southworth and others, 2009).

A late Neoproterozoic cover sequence of metasedimentary and metavolcanic rocks unconformably overlie the basement complex in the Shenandoah National Park region (Fig. 3). The Swift Run Formation is a heterogeneous clastic unit of highly variable thickness (absent to ~300 m) that crops out below metabasalts of the Catoctin Formation and to the east of the Park in outliers surrounded by basement (Gattuso and others, 2009). Common rock types include arkosic phyllite, meta-arkose, phyllite, laminated metasiltstone, and pebble to cobble metaconglomerate. Locally, compositionally mature, cross-bedded, quartz-rich metasandstone crops out. Gathright (1976) and Schwab (1986) interpret the Swift Run Formation to be a non-marine unit deposited in alluvial fan, floodplain, and lacustrine environments. A number of early workers report tuffaceous rocks from the Swift Run Formation, but recent research indicates these are primarily fine-grained metasedimentary rocks. Contemporaneous normal faulting likely influenced the deposition of Swift Run sediments in the outliers (Forte and others, 2005; Gattuso and others, 2009). At a number of locations clastic rocks are interlayered with metabasaltic greenstone, a geometry consistent with a coeval relationship between the Swift Run Formation and the lower Catoctin Formation (King, 1950; Gattuso and others, 2009).

The Catoctin Formation forms an extensive unit in Shenandoah National Park and is characterized by metabasaltic greenstone with thin layers of meta-arkose, phyllite, and epiclastic breccia. Catoctin basalts were extruded over a large region (>4000 km^2) and generated from mantle-derived tholeiitic magmas (Badger and Sinha, 2004). In the Park, basalts extruded primarily as subaerial flows, evidenced by abundant columnar joints and flow-top breccias (Reed, 1955, 1969). In the central and northern part of the Park, nine to sixteen individual flows occur (Reed, 1955; Gathright, 1976; Badger, 1992). The Catoctin Formation is upwards of 700 meters thick in the Park and thins towards the west and southwest. Metadiabase dikes of similar composition to Catoctin metabasalts intrude the basement complex (as well as older Neoproterozoic rocks) and are likely feeder dikes for the overlying Catoctin lava flows. Dikes range from 0.5 to 5 m in thickness and are most common in the northern and central part of the Park.

Badger and Sinha (1988) report a Rb-Sr isochron age of 570±36 Ma from exposures just south of Shenandoah National Park. Zircons from metarhyolite tuffs and dikes in the Catoctin Formation, exposed to the north of the Park, yield U-Pb ages between 570 and 550 Ma (Aleinikoff and others, 1995). Collectively, the geochronologic data suggest that Catoctin volcanism lasted between 15 and 20 million years and occurred during the Ediacaran period in the late Neoproterozoic. Paleomagnetic data from the Catoctin Formation are complex, but broadly compatible with a high southerly latitude (60˚ S) for the Virginia Blue Ridge during extrusion (Meert and others, 1994).

The siliciclastic Chilhowee Group overlies the Catoctin Formation. In Shenandoah National Park, the Chilhowee Group includes the Weverton, Harpers, and Antietam formations and ranges from 500 to 800 meters in aggregate thickness (Fig. 3). Gathright's (1976) nomenclature for the Chilhowee Group includes the Weverton, Hampton (Harpers), and Erwin (Antietam) formations. Weverton, Harpers, and Antietam are units whose type locations occur along the Potomac River approximately 125 km to the northeast, whereas the Hampton and Erwin type locations are located northeastern Tennessee. The contact between the underlying Catoctin metabasalts and the overlying Chilhowee Group has traditionally been interpreted as an unconformity, but its significance remains uncertain (King, 1950; Gathright, 1976; Southworth and others, 2007a).

The Weverton Formation includes quartz metasandstone, granule to pebbly metaconglomerate, laminated metasiltstone, and quartzose phyllite. The Harpers Formation, in Shenandoah National Park, is dominated by green to gray phyllite and thinly bedded metasandstone. Less common rock types include well-cemented quartz arenite and ferrigunous metasandstone. Trace fossils (*Skolithos* and burrowed beds) are rare, but do occur in the upper part of the Harpers Formation. The Antietam Formation consists of distinctive well-cemented quartz arenite, with abundant *Skolithos*, and laminated metasiltstone. Collectively, the Chilhowee Group records a fluvial to shallow-marine transgressive sequence (2^{nd}-supersequence) (Simpson and Eriksson, 1990; Read and Eriksson, *in press*). *Skolithos* is the only well documented fossil in the Chilhowee Group from the Shenandoah National Park region, but along strike to the southwest, *Rusophycus* occurs near the base and *Olenellus* trilobites near the top, bracketing the age of the Chilhowee Group between the earliest Cambrian and early Middle Cambrian (<545 Ma to ~515 Ma). A thick package of Cambro-Ordovician carbonate and shale overlies the Chilhowee Group and crops out in the Shenandoah Valley west of the Park.

Unmetamorphosed diabase dikes cut the Mesoproterozoic basement and the Neoproterozoic to early Cambrian cover sequence. Dikes are volumetrically insignificant within the Park, range from <1 m to ~20 meters in thickness and commonly occurs in north-northwest striking en-echelon sets. The olivine normative, low-titanium diabase is similar to mafic dikes and sills exposed in the Culpeper Mesozoic basin to the east of the Park. A diabase dike along the Potomac River, ~100 km north of the Park, yielded an Ar-Ar age of 200 Ma (Kunk and others, 1992).

Structural Geology

Shenandoah National Park is on the western limb of the Blue Ridge anticlinorium, a regional scale structure extending from south-central Pennsylvania to the Roanoke area. At the latitude of the Park, the Blue Ridge anticlinorium is an imbricated stack of basement thrust sheets emplaced over lower Paleozoic strata during contractional deformation in the late Paleozoic (Mitra, 1979; Evans, 1989). The Blue Ridge province is separated from Valley & Ridge rocks by a family of related, low-angle thrusts that place Mesoproterozoic basement and Neoproterozoic to Cambrian cover rocks of the Blue Ridge on sedimentary units as young as middle Ordovician (Figs. 4 and 5). Northwest of the Park the Massanutten synclinorium forms a regional scale fold complex that exposes rocks as young as Devonian (Figs. 4 and 5).

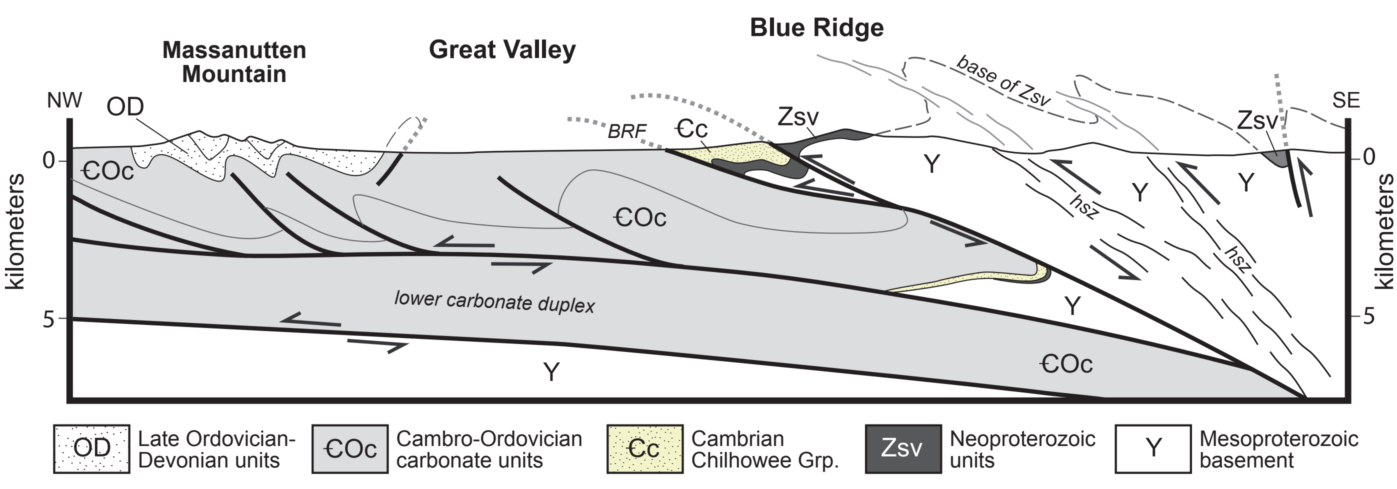


Figure 5. *Geologic cross section of Blue Ridge, Great Valley, and Massanutten Mountain. Compiled from various resources.*

Individual fault segments of the frontal Blue Ridge fault system are 20 to 50 km (12 to 30 mi.) long and tip out or are truncated along strike. The Front Royal and Stanley faults along the western margin of the Park are well delineated (King, 1950; Wickham, 1972; Gathright, 1976). Further south, in the Elkton to Waynesboro area, an unnamed and long unrecognized frontal Blue Ridge fault is concealed beneath surficial deposits along the western foot of the Blue Ridge (Figs. 4 and 5). The frontal Blue Ridge fault system truncates early fold structures and penetrative fabrics in both the hanging-wall and footwall rocks; the deformation style is typically brittle and breccias are well developed at a number of localities. Folds in western Blue Ridge cover rocks are typically asymmetric northwest-verging structures. Axial planar foliation (cleavage) is well developed in fine-grained rocks and dips gently to moderately southeast.

Mylonitic rocks occur in anastomosing zones, up to a kilometer thick, that cut basement rocks along the eastern margin of the Park (Figs. 4 and 5). Mylonite zones in the Park were first recognized by Gathright (1976) and aspects of these rocks have been discussed by Mitra (1977), Bailey and Simpson (1993), and Bailey and others (2002). Asymmetric structures consistently record a top-to-the-northwest sense of shear (i.e. hanging wall up movement). The reverse displacement across Blue Ridge high-strain zones accommodated crustal contraction enabling the relatively stiff basement complex to shorten while cover rocks were folded.

A distinctive coarse foliation or compositional banding is developed in the older Mesoproterozoic basement units (Groups 1 and 2, >1,150 Ma) and defined by aligned feldspars and quartz aggregates that formed at upper amphibolite- to granulite-facies conditions. Younger Mesoproterozoic basement units (Group 3, <1,090 Ma) typically lack a high-temperature fabric. A younger foliation is variably developed in both the basement and cover sequence, this foliation characteristically strikes northeast (010˚ to 050˚), dips to the southeast, and is defined by aligned greenschist-facies minerals. Mitra and Elliott (1980) recognized this fabric (Blue Ridge-South Mountain cleavage of Cloos, 1971) in Silurian and Devonian rocks of the Massanutten synclinorium. At many locations, a penetrative southeast plunging lineation (approximately down the foliation dip) is also developed.

All bedrock units in the Blue Ridge are variably fractured. Majer and Bailey (2000), Hasty and Bailey (2005), and Bailey and others (2006) conducted systematic fracture analysis in the south-central part of the Park. The basement complex is characterized by numerous variably oriented fracture sets. Although dominant sets occur at individual outcrops, there is no systematic pattern across the region. Many fractures in the basement complex are shear fractures with complex kinematics. The cover sequence (Swift Run and Catoctin formations and Chilhowee Group) contains three distinct fracture sets that include: 1) an older, northeast-striking (020°-040°), moderately southeast-dipping set, 2) a younger, east-striking (080°-110°) subvertical set, and 3) a younger north-northwest-striking (340˚-355˚) set. The early joint set was tilted from a subvertical orientation during folding (Hasty and Bailey, 2005). To the northwest, in Valley & Ridge rocks, Engelder (2004) documented two dominant fracture sets: an early, east-striking set that pre-dates folding and a younger, northwest-striking set that post-dates folding. East-striking, subvertical extension fractures in Blue Ridge cover rocks may be equivalent to the early set in the Valley & Ridge (although the Blue Ridge set post-dates folding) and related to dextral contraction during the early stages of the Alleghanian orogeny.

Transverse faults have long been recognized in the Virginia Blue Ridge (King, 1950; Reed, 1955; Gathright, 1976), but their kinematic and temporal significance has remained uncertain. In the south-central part of the Park, transverse faults strike to the north-northwest (330°-350°) and west-northwest (280°-300°) and cut the regional structural trend at high angles (Fig. 4). These faults displace Mesoproterozoic basement, Neoproterozoic to Cambrian cover rocks, and Cambrian to Ordovician units west of the Blue Ridge. In map view, transverse faults displace geologic contacts by up to 300 m. Transverse faults commonly form right-stepping en-echelon sets. Although fault surfaces are not exposed, straight map patterns in areas with significant topographic relief indicate steep dips (>80°). Transverse faults cut previously folded contacts and the apparent strike-slip offset in map view is accomplished by dip-slip movement with maximum displacements of ~100 m. A number of transverse faults, including the recently recognized Simmons Gap and Sandy Bottom faults in Shenandoah National Park, correspond to well-developed topographic lineaments (Fig. 4; Bailey and others, 2009).

North-northwest striking transverse faults and extension fractures are subparallel to a regional suite of Jurassic diabase dikes. Bailey and others (2006) interpreted many transverse faults in Shenandoah National Park as Jurassic structures noting that these faults record minor east-northeast directed extension and possibly developed in a transtensional stress regime. Wieczorek and others (2004) used LIDAR imagery to identify a topographic lineament, which they interpret as a north-northwest-striking fault (Harriston fault) near Grottoes and suggested this structure could have experienced Quaternary slip. Further research is required to evaluate the possibility of neotectonic activity along transverse faults. Many transverse faults are zones of structural weakness and strongly influenced the topographic character of the Blue Ridge Mountains; Simmons, Powell, Smith Roach, and Swift Run gaps are all located on transverse faults.

Tectonic History

Rocks and structures exposed in Shenandoah National Park formed in response to tectonic processes associated with the opening of ocean basins and the collision of continents throughout the past 1.2 billion years. Major tectonic events include the Mesoproterozoic Grenvillian orogen, Neoproterozoic Iapetan rifting, multiple episodes of Paleozoic collision, and Mesozoic crustal extension (Fig. 6).

Mesoproterozoic granitoids and gneisses formed in the middle and lower crust along the margin of Laurentia during the long-lived Grenvillian orogeny that culminated in the amalgamation of the Rodinian supercontinent by 1,000 Ma. Older basement granitoids were emplaced during a magmatic interval 30 to 40 My in duration (1,180 – 1,140 Ma). A high-grade deformation event transformed these rocs into granitoid gneisses prior to the intrusion of the youngest plutonic suite by 1,080 Ma. The older deformation event is characterized by many northwest to east-west trending structures, but the kinematics of this event remain poorly understood. Lead isotopes in central and southern Appalachian basement massifs are distinctly different from other Laurentian Grenvillian provinces implying that Blue Ridge basement is not "native crust", but rather was accreted to Laurentia during this long-lived orogeny.

The small exposures of post-1,000 Ma paragneiss require uplift and erosion of the Blue Ridge basement followed by burial, metamorphism, and exhumation before Rodinia broke apart. From 730 to 700 Ma, plutonic and metavolcanic rocks of the Robertson River Igneous Suite were emplaced during an episode of crustal extensional. Although this tectonic event did not result in the formation of an ocean basin, it created significant accommodation space into which the thick sediments of the Fauquier, Lynchburg, and Mechum River sequence were deposited. The evidence of this first episode of Neoproterozoic crustal extension and rifting is preserved primarily to the east of Shenandoah National Park.

Deposition of Swift Run sediments, emplacement of mafic dikes, and extrusion of Catoctin volcanics formed during a rift event that began ~570 Ma. Immediately southeast of the Park (in Greene and Albemarle counties), syn-sedimentary normal faults are associated with outliers of Swift Run and Catoctin formations. The regionally extensive Catoctin Formation may have been generated from plume-related processes that drove rifting along this segment of the Laurentian margin. The non-marine (Weverton) to nearshore-marine (Harpers and Antietam) rocks of the Chilhowee Group record the rift-to-drift transition as a passive continental margin developed along the Laurentian margin of the Iapetus Ocean in the Early Cambrian. Rocks in the Blue Ridge were progressively buried under a sequence of dominantly marine Cambrian and Ordovician sedimentary rocks (>8 km thick) deposited on the Laurentia continental margin.

By the middle to late Ordovician, the Taconic orogeny had began in the Appalachians. Although numerous workers have argued for significant Taconic deformation in the Virginia Blue Ridge, recent geochronological studies provide little evidence of Ordovician activity. A late Ordovician to Silurian clastic wedge was deposited to the northwest of the modern Blue Ridge and significant subsidence continued into the middle Paleozoic (as preserved in Valley & Ridge sedimentary sequences). New Ar-Ar geochronology indicates that pervasive deformation in the basement and cover sequence occurred between ~360 and 310 Ma (late Devonian to Pennsylvanian). Deformation and metamorphism of this age falls between the classic Appalachian Devonian Acadian and Pennsylvanian-Permian Alleghanian orogenies. Zircon fission track data reveals the Blue Ridge had cooled through ~235˚ C by 300 Ma. The frontal Blue Ridge fault system slipped during the late Alleghanian (300 to 280 Ma) when the Blue Ridge massif was relatively cool. Evidence from the southern and central Appalachians suggests that middle to late Paleozoic tectonism was protracted and involved dextral

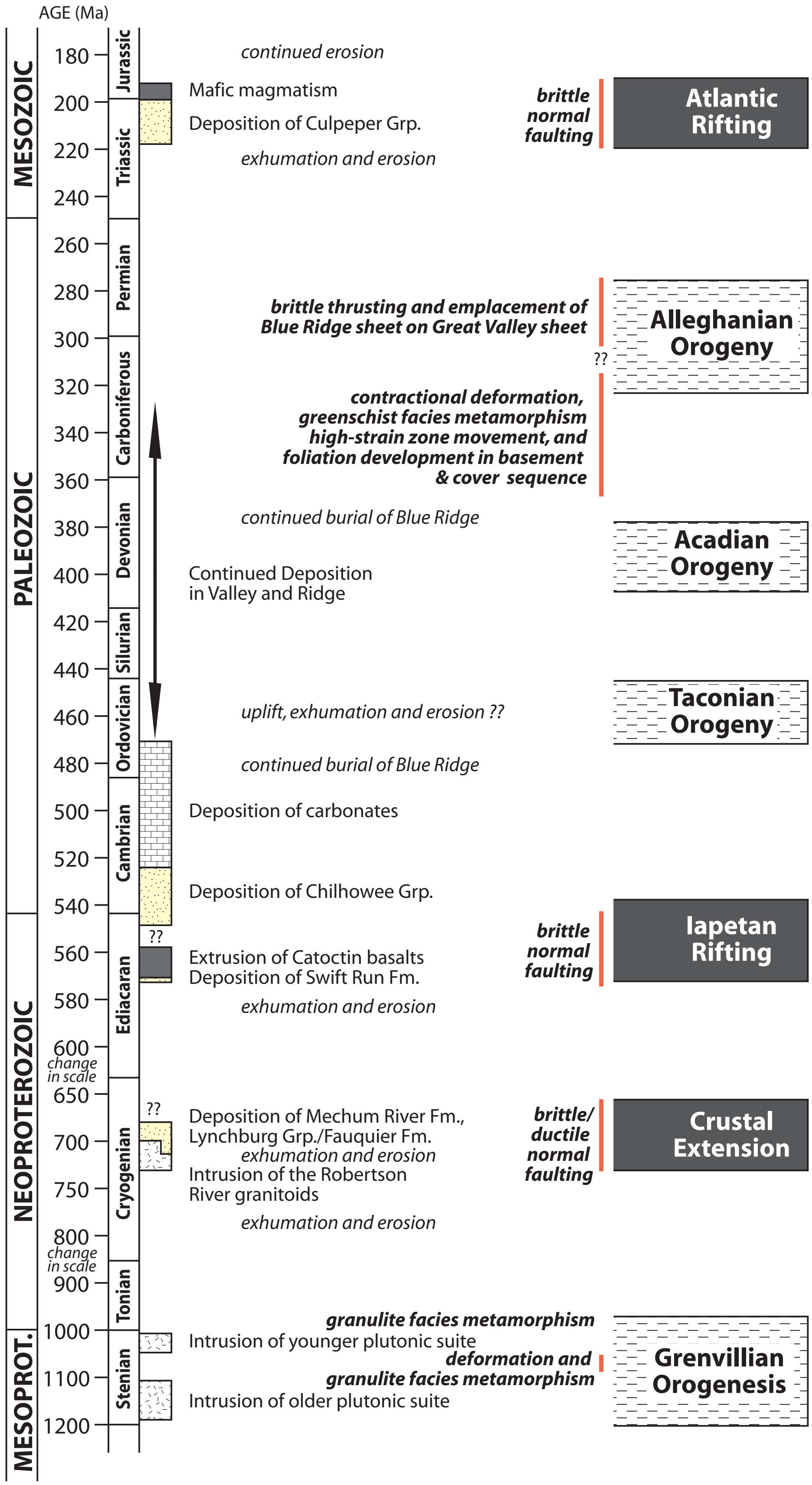


Figure 6. *Generalized geologic and tectonic history of the Blue Ridge (from Bailey and others, 2006).*

contraction that transitioned to orthogonal convergence. The Alleghanian orogeny culminated with collision between North America and Africa, leading to the Pangean supercontinent by the close of the Paleozoic.

In the early Mesozoic eastern North America and northwestern Africa began to rift apart, initiating a process that opened the Atlantic Ocean. In north-central Virginia, this event created the Culpeper basin in the eastern Blue Ridge and western Piedmont. Rifting generated basaltic magmas in the early Jurassic, and in the Blue Ridge to west-northwest-striking diabase dikes were intruded during a phase of modest east-west extension. Apatite fission-track ages reveal that rocks in the Shenandoah National Park region cooled below ~90° C during the Jurassic and Cretaceous indicating that Blue Ridge rocks have been near the surface for over 100 million years. The topographic expression of the Blue Ridge Mountains developed during the late Cenozoic as differential erosion by Atlantic flowing rivers (Potomac, Rappahannock, and James) preferentially removed less resistant rocks in the Valley & Ridge and Piedmont.

Surficial Geology

Unconsolidated surficial materials and their resulting landforms cover large areas of the Shenandoah National Park region. Traditionally, geologists have underappreciated these deposits and their significance. An array of new studies focused on these deposits and surficial processes have yielded significant insights about the Blue Ridge landscape. Recent surficial investigations in the Shenandoah Park region include studies by Sherwood and others (1987), Whittecar (1992), Morgan and others (1999a,b), Litwin and others (2001, 2003), Eaton and others (2003a,b), Smoot (2004), Morgan and others (2004), and Wieczorek and others (2004). Surficial materials contribute to soil character and play a key roll in land use, water resources, and hazards (sinkhole subsidence, debris flows, and flooding). Landforms and surficial deposits in and near the Park reflect processes active over a broad time spectrum and were formed under a wide range of climate conditions.

Surficial deposits include those formed by (1) flowing water (alluvium, terrace deposits, alluvial-fan deposits, and alluvial-plain deposits), (2) gravity and high rainfall events on slopes (debris-flow and debris-fan deposits), (3) gravity and freeze-thaw processes on slopes (stratified slope deposits, colluvium, and block-field deposits), and (4) chemical weathering (sinkholes and residuum) (Fig. 7). In general, alluvial-fan and alluvial-plain deposits are located on the lower slopes and valleys on the west and east sides of the Blue Ridge, respectively, and terrace deposits are located along major rivers in the valleys. Debris-fan deposits are concentrated in coves and hollows on the upper to lower slopes. Colluvium is concentrated in the

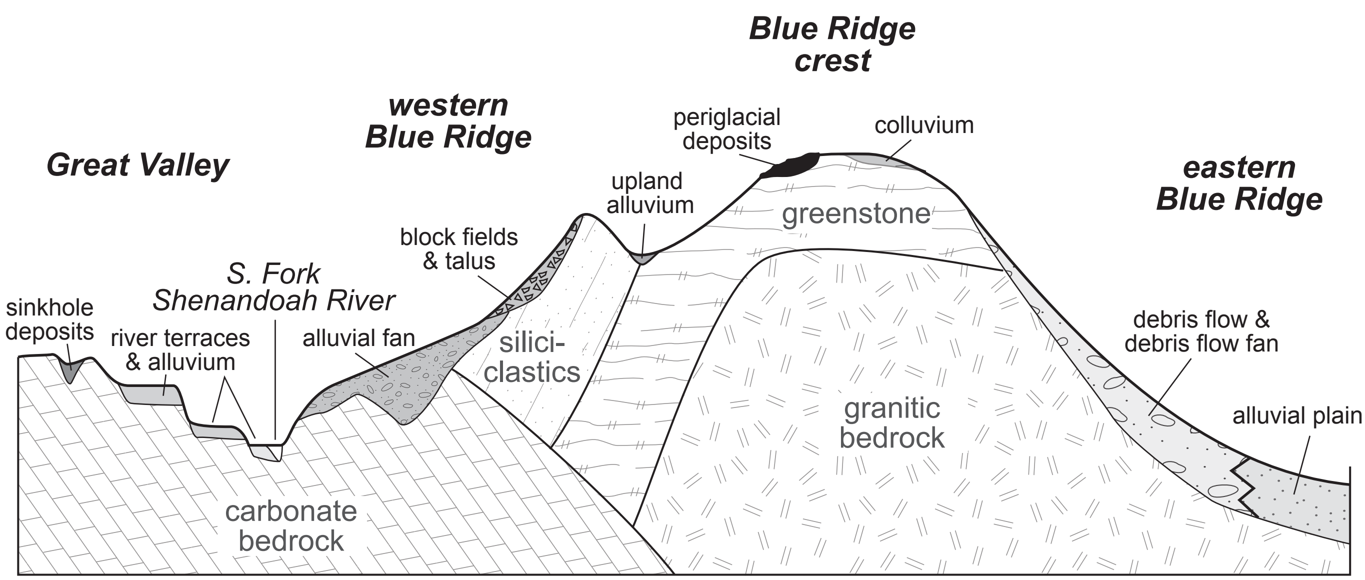


Figure 7. *Simplified model of surficial deposits in the Shenandoah National Park region.*

highlands. Debris-fan deposits are abundant in areas underlain by gneisses and metabasalts, whereas block-field deposits are most abundant in the areas underlain by siliciclastic rocks. Extensive alluvial-fan deposits cover carbonate bedrock immediately west of the Blue Ridge. The South Fork of the Shenandoah River has incised the alluvial fans to form terraces. In addition, sinkholes occur in the carbonate bedrock, including bedrock buried by alluvial fan deposits. In contrast, the lowlands east of the Blue Ridge are characterized by broad alluvial aprons along streams and by some terraces mantled with gravel.

The age of surficial materials in the Shenandoah National Park region ranges from the late Pliocene to the Recent. Weathering characteristics of deposits in high terraces and fans are similar to late Pliocene deposits in the Fall Zone and Coastal Plain (Howard, 1993; Eaton and others, 2001). Debris fans have formed since at least the late Pleistocene, Eaton and others (2003a) obtained ^{14}C ages for organic matter in fans ranging from about 51,000 to 2,000 yr B.P. These old debris fan deposits were eroded and exposed by debris flows generated from a high rainfall event in 1995 that deposited much sediment on lower slopes and floodplains.

Measurements of summit erosion rates in Shenandoah National Park using ^{10}Be

Gregory Hancock and Jennifer Whitten, *College of William & Mary*

Understanding the change in landscape morphology through time requires knowing erosion rates at various points within that landscape, particularly valley and summit erosion rates, and rates averaged over differing timescales. For instance, similar valley and summit erosion rates could suggest a state of dynamic equilibrium, wherein “all elements of the topography are mutually adjusted so that they are downwasting at the same rate” (Hack, 1960, p. 85). Hack (1960) proposed this concept from observations in the central and southern Appalachian landscape, and recent work in the Great Smoky Mountains suggests ridge crest, bare-bedrock erosion rates and longer-term exhumation rates determined from thermochronology are similar to basin average erosion rates, supporting a state of dynamic equilibrium in that landscape (Matmon and others, 2003).

While the concept of dynamic equilibrium is attractive given a lack of substantial orogenesis over the last ~300 My in the Appalachians, disequilibrium may be produced by climatic effects on rates and patterns of erosion in mountainous topography. Peizhen and others (2001) suggest that the late Cenozoic transition to rapidly changing climate conditions 3–4 My ago led to a widespread acceleration of fluvial and glacial erosion rates, leading to relief changes as valley erosion rates increased relative to summit lowering rates in many mountainous regions (e.g, Molnar and England, 1990). Increasing relief has been documented in many other ranges during the late Cenozoic (e.g., Small and Anderson, 1995), which is consistent with disequilibrium induced by late Cenozoic climate change. Along the eastern North American margin, sedimentation rates doubled between the Late Miocene and the Quaternary (Poag and Sevon, 1989) and river incision rates across the region appear to have increased since the late Miocene (Mills, 2000). Hancock and Kirwan (2007) documented summit erosion rates on the eastern margin of the Appalachian Plateau in West Virginia that are well below measured fluvial erosion rates over a comparable period (~200 ky BP). This is consistent with increasing relief and landscape disequilibrium in at least in some areas of the central and southern Appalachians. In contrast to this proposed increased relief, other work in the central Appalachians calls for rapid summit erosion under periglacial climates, leading to an overall decrease in relief. Braun (1989) hypothesizes that periglaciation is the dominant erosional process in the central Appalachian uplands, is most efficient during glacial periods (and, hence, would be made possible as a consequence of late Cenozoic cooling), and occurs at rates greater than fluvial erosion rates. If true, the result of climate cooling in the central Appalachians has been the acceleration of upland erosion rates and a disequilibrium resulting in a decrease in relief during the Late Cenozoic.

These three contrasting hypotheses pose the following question: are the central Appalachians in a state of dynamic equilibrium or not? We address this question by measuring rates of summit erosion at numerous locations within the park, and comparing these rates to previously determined basin-averaged and fluvial erosion rates. To determine erosion rates, we utilize the abundance of the cosmogenic radionuclide ^{10}Be accumulated on bare-bedrock summits. We focus on bare-bedrock summits as these are likely to be the most slowly eroding parts of the landscape, and therefore limit the overall lowering of elevation and relief within the range. To date, we have measured a total of 23 bare-bedrock erosion rates from nine summit locations located within four different lithologies in Shenandoah National Park (Fig. 8).

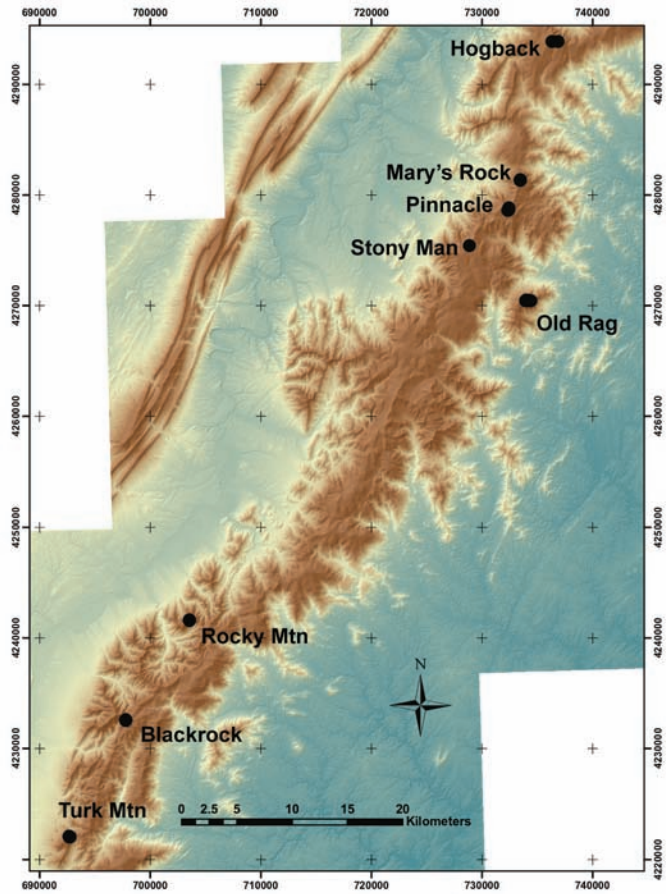


Figure 8. *Sampling locations within Shenandoah National Park.*

Blackrock Summit (Stop 3) is one of the sampling locations from which bare-bedrock erosion rates have been attained. The pale blue to purplish rocks here are well-cemented quartz sandstones and/or quartzites of the Lower Cambrian Harpers Formation (Gathright, 1976; Olney and others, 2007; Southworth and others, 2009). Weathering of this rock produces well-defined fractures oriented along bedding planes. Lowering of this summit likely occurs through removal of blocks defined by these weathered sub- horizontal bedding planes coupled with vertical joints, as well as toppling of isolated tors and direct removal by rain and wind of granular material produced by weathering. Block sliding and tor toppling contribute blocks to an extensive downslope talus field comprised of regularly sized blocks (Gathright, 1976). The absence of vegetation, unstable blocks, and preferred block alignment may suggest recent and/or ongoing downslope creep of blocks in this talus field (Eaton, pers. comm.).

To sample this site and others, we collected rock samples from nearly flat bedrock outcrops, and measured the abundance of the cosmogenic radionuclide (CRN) ^{10}Be in quartz extracted from the rock (Fig. 9). The surface area of the sampled bedrock outcrops range from a few 10's to 100's of square meters, and ~1-5 cm thick samples are collected from the bedrock surface. The abundance of ^{10}Be in surface samples is interpreted as a steady-state erosion rate (Bierman, 1994) and calculations of surface production rate corrected for elevation, latitude and horizon blockage are obtained using the methods of Balco and others (2008).

A ***B***

Figure 9. *Examples of sample locations. A) SH-12 on Blackrock Summit. B) SH-28 on Old Rag summit.*

The mean erosion rate all sampled summits in Shenandoah is 8.67 m/My, and rates varies from 1.56 ± 0.16 m/My to 41.19 ± 8.9 m/My (Fig. 10). The average rate from the three surfaces sampled at Blackrock is 17.9 ± 5.1 m/My. Because erosion rates here and in other locations where block removal is important are not, strictly speaking, eroding steadily, we must consider the validity of the steady state erosion assumption. Small and others (1997) model ^{10}Be accumulation on bare-bedrock surfaces to determine how much error is introduced by applying a steady-state assumption to a location where erosion is caused by episodic block removal. The primary factors influencing the error are the long-term erosion rate and the thickness of the block removed (Small and others, 1997). Following the strategy of Small and others (1997), estimates of the maximum error introduced by our use of the steady-state assumption range from ~±1-18% of the actual long term erosion rate, with errors generally decreasing as estimated erosion rate increases (Whitten, 2009). Although this may appear to be a significant error, given that the lowest rates are on the order of a few m/My, a 20% error does not matter when comparing these results to other rates of landscape erosion, as will be discussed later. In addition, such errors are nonsystematic, multiple samples should yield average erosion rates with less overall error relative to the actual average erosion rate (Small and others, 1997). We consider the erosion rates obtained to reflect maximum erosion rates. Most errors introduced by recent geologic events that would change production rates on sampled surfaces (e.g., devegetation and soil stripping) would tend to result in ^{10}Be abundances that would yield higher than average rates.

Our measurements suggest that summit erosion rates in Shenandoah National Park are generally slow, and are closely related to rock type (Fig. 10). The Harpers Formation yields the highest average rate of ~23 m/My, ~4 to 10 times higher than the other three lithologies sampled. The remaining three lithologies show very low erosion rates, and are similar to bedrock erosion rates in both periglacial environments and elsewhere in the Appalachians. Small and others (1997) found an average bare-bedrock erosion rate of 7.9 ± 4.1 m My^{-1} on summit tors in the central Rocky Mountains and Sierra Nevada, USA. Our bare-bedrock summit rates are similar to non-fluvial erosion rates measured in this region, including on bare granite inselbergs in Georgia (2–10 m/My, Bierman and other, 1995); on bare sandstone and conglomerate capped ridges in Kentucky (~2 m/My, Granger and others, 2001); on soil mantled hilltops in West Virginia (4–7 m/My, Clifton and Granger, 2005); on bare sandstone summits along the margin of the Appalachian Plateau in West Virginia (2-10 m/My, Hancock and Kirwan, 2007); and the bedrock-to-saprolite conversion rate beneath regolith in the Virginia piedmont (4.5–8 m/My, Pavich, 1989). In addition, our rates are comparable to an average of 7.8 m/My obtained by Duxbury (2009) obtained from five bare-bedrock summit samples collected within Shenandoah National Park.

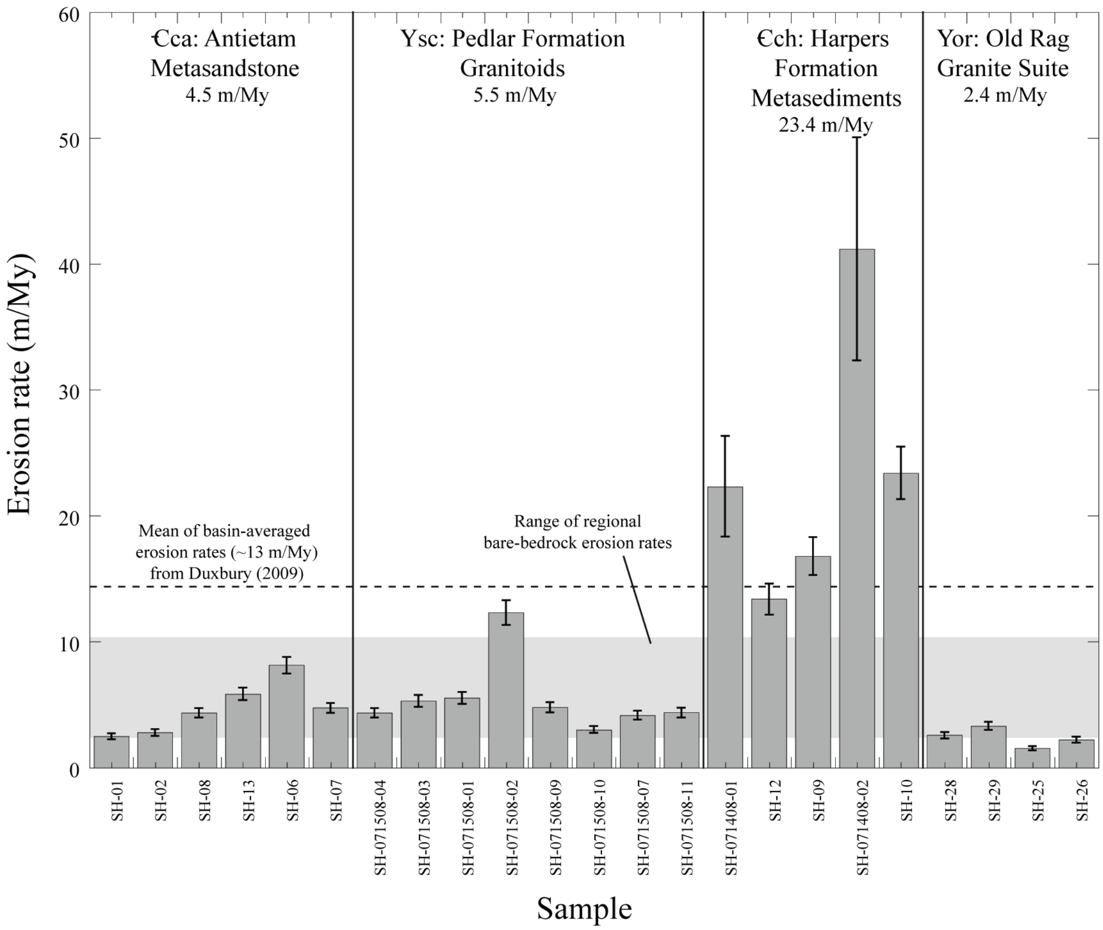


Figure 10. *Summit erosion rates measured in Shenandoah National Park. Geology from Southworth and others (2009).*

These results suggest that bare-bedrock weathering rates on Shenandoah summits developed on lithologies other than the Harpers Formation are low, and are not significantly different from weathering rates obtained from a variety of other non-fluvial settings in the region.

Comparison of our summit erosion rates to basin-averaged rates obtained by Duxbury (2009) suggests variation in the direction of relief change that is dependent on lithology. Overall, the average summit erosion rate (~8.8 m/My) is lower than the average of the basin-averaged rates (~13 m/My) of Duxbury (2009). More importantly, comparison of rates on differing lithologies yields evidence for relief change in some locations within the region. There are some difficulties in direct comparison between this study and Duxbury (2009), as we refer to the formations sampled while she refers to specific rock type sampled, but we can make general comparisons here by assuming the formation from which her rock types must have been derived. The basin-averaged erosion rate for granite-dominated watersheds is 15±4 m/My (Duxbury, 2009), while our summit erosion rates in similar lithologies are 5.5 m/My (Pedlar granitoids) and 2.4 m/My (Old Rag Granite Suite). The average for siliciclastic-floored basins is 11±4 m/My (Duxbury, 2009), while summit erosion rates in similar lithologies are 23±12 m/My (Harpers Formation). The average for quartzite-floored basins is 8.0±2 m/My (Duxbury, 2009), while summit erosion rates in similar lithologies are 4.5 ± 2.1 m/My (Antietam “quartzites”). Taken at face value, these comparisons imply a small but positive change in relief in granitic- and quartzite-dominated watersheds, and a decline in relief in the siliciclastic-dominated watersheds.

Importantly, the basin-averaged samples collected by Duxbury (2009) integrate across entire watersheds, thus including both slowly and rapidly eroding portions of the watersheds. The fact that the basin-averaged erosion rates in two of the sampled lithologies are higher than summit erosion rates suggest that sideslopes and/or channels must be eroding more rapidly than summits. The suggestion that channel erosion rates may be greater than the basin-averaged rates observed by Duxbury (2009) is supported by observations of fluvial incision rates averaged over similar timescales in this region. Fluvial incision rates compiled over a broad region of the east-central and southeastern U.S. average ~30–1000 m/My over ~10^4 to 10^6 years on Piedmont, Blue Ridge/Valley & Ridge, and Appalachian Plateau rivers (Mills, 2000). Incision rates into bedrock obtained from adjacent basins include ~50–160 m/My obtained from cave magnetostratigraphy and ^{10}Be dating of terraces on the James River (Erickson and Harbor, 1998; Hancock and Harbor, 2003), ~20 m/My from CRN dating of fluvial terraces on the New River (Ward and other, 2005), and ~600–800 m/My during the late Pleistocene from CRN dating of strath surfaces on the Potomac and Susquehanna Rivers (Reusser and others, 2004). If these rates are indicative of channel erosion rates within the Shenandoah National Park region, the difference between our bare-bedrock, summit-lowering rates and the fluvial incision rates imply that at least some parts of the Blue Ridge may be currently undergoing a period of growing relief.

Acknowledgements

Putting together a field conference with eight leaders (a.k.a.- co-conspirers) is an exciting proposition. We thank the Park Service for their warm welcome (and the fee waiver!) to Shenandoah as well as the use of the Byrd Visitors Center for the Friday evening session. Julena Campbell, Jim Schaberl, and Tim Taglauer were not only gracious, but extraordinarily helpful with logistics- thanks. VGFC officers, Amy Gilmer and Matt Heller, secured transportation and provisions for the conference. The U. S. Geological Survey has funded much of the research, in particular the EDMAP program has been instrumental in supporting the new mapping.

Our colleagues, John Aleinikoff, Lorrie Coiner, Jorge Dinis, Amy Gilmer, Matt Heller, Bill Henika, Craig Kochel, Mick Kunk, Ben Morgan, Cullen Sherwood, Joe Smoot, Dick Tollo, and Steve Whitmeyer have added much with their insight. A legion of William & Mary students completed senior theses in and around the Park: significant contributions (of blood, sweat, and intellectual energy) have been proffered by Adam Forte, Joe Fuscaldo, Brian Hasty, Adam Gattuso, Colleen Keyser, Graham Lederer, Crystal Lemon, Joe Olney, Owen Nicholls, and Katie Wooton.

ROAD LOG and Stop Descriptions

The field trip road log begins at the bus parking area ~100 meters to the north of the Byrd Visitors Center at Big Meadows, Shenandoah National Park (mp 51). All field trip stops will be in the Park. Stops 4 and 7 involve hikes along well-maintained trails. Participants should be mindful of field conditions and exercise prudence. A number of stops include exposures along the Skyline Drive, although there is an ample shoulder at these outcrops participants must be cautious. ***Stay off the active road way and always watch for traffic***. As with the flora and fauna, rocks in Shenandoah National Park are protected. ***No hammering or collecting of samples is permitted in the Park***. At a few stops unweathered and fresh rock samples will be provided for closer inspection, but remember do not hammer on outcrops. Stop locations are given in decimal latitude and longitude using the NAD 27 datum.

Directions and *comments*

From the bus parking area follow leaders (on foot) to vantage point of Big Meadows.

STOP 1

Big Meadows (38.5168˚ N, 78.4363˚ W)

Late Quaternary to Recent sediments, Neoproterozoic Catoctin Formation

Litwin and Eaton

The Big Meadows site (BMS) in Shenandoah National Park comprises Late Quaternary to Recent sediments that unconformably overlie the late Neoproterozoic Catoctin Formation (Allen, 1963; Gathright, 1976). Previous coring has determined that at least 6 m of unconsolidated sediment were deposited in the topographically lowest part of the BMS (Litwin and others, 2004a), although Quaternary deposits in the meadow are non-uniform in thickness and distribution. BMS is located at an elevation of 1,070 m (3500 ft) and experiences a cool microclimate; that in combination with a shallow perched water table within the alluvial sediments enable it to host a refugial cold-tolerant modern flora. The meadow supports small populations of balsam fir (*Abies balsamea*), red spruce (*Picea rubens*), Red-osier dogwood (*Cornus stolonifera*), and Canadian burnet (*Sanguisorba canadensis*). Each of these plants more typically thrives in environments at least 4-5˚ latitude north of this site, in New England and northward. It also hosts a pine species restricted to the cooler microclimate along the ridge crests of the Blue Ridge and Appalachians, Table Mountain pine (*Pinus pungens*; Little, 1971, 1976; W.B. Cass and R. Engle, NPS (SHEN), pers. comm.). The Virginia Blue Ridge generally is vegetated by a southern extension of the temperate Appalachian oak forest, and this forest surrounds the BMS (Küchler, 1964, 1975; Litwin and others, 2004a). Figure 11 illustrates the general forest structure presently established in eastern North America, and indicates the location of Shenandoah National Park with respect to it. From south to north, these are: 1) southern mixed forest, 2) oak-hickory-pine forest, 3) Appalachian oak forest, 4) northern hardwoods forest, 5) northern hardwoods-spruce forest, 6) spruce-fir forest (minor type), 7) boreal forest, and 8) forest-tundra to tundra (Küchler, 1964, 1975). The modern mean annual temperature (MAT) isotherms for this same geographic area also are shown (Owenby and others, 1992; NCDC, 2002: Litwin and others, 2004a).

Fossil pollen evidence has been recovered from sparsely distributed, small depositional units across the eastern flank of the Blue Ridge, in and around Shenandoah National Park. Recent debris-flow denudation in the Blue Ridge (Morgan and Wieczorek, 1996; Morgan and others, 1999; Eaton, 1999; Wieczorek and others, 2000; Eaton and others, 2003) presented an opportunity to identify and collect climate information from these isolated Late Quaternary deposits. Most of these samples were dated directly using AMS ^{14}C analyses. These samples indicated that the vegetation in the Blue Ridge changed repeatedly with changes in Late Quaternary climate (Litwin and others, 2001, 2004a). Figure 12 shows a map of the Blue Ridge sites from which fossil pollen samples were collected (Litwin and others, 2004a).

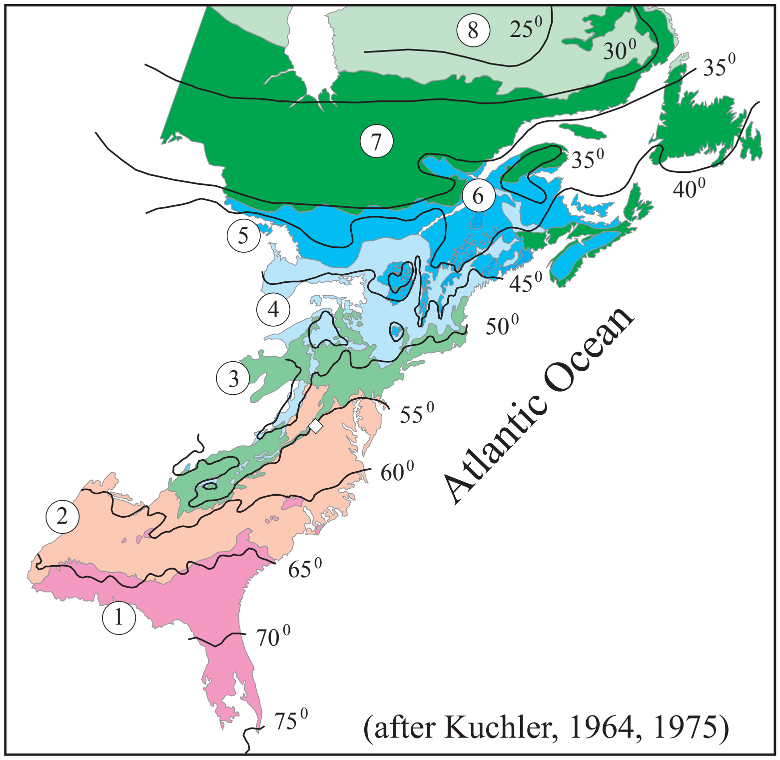


Figure 11. *Generalized forest zonation of eastern North America, showing modern mean annual temperature isotherms and forest zones. From south to north, these are: 1) southern mixed forest, 2) oak-hickory-pine forest, 3) Appalachian oak forest, 4) northern hardwoods forest, 5) northern hardwoods-spruce forest, 6) spruce-fir forest (minor type), 7) boreal forest, and 8) forest-tundra to tundra (Küchler, 1964, 1975).*

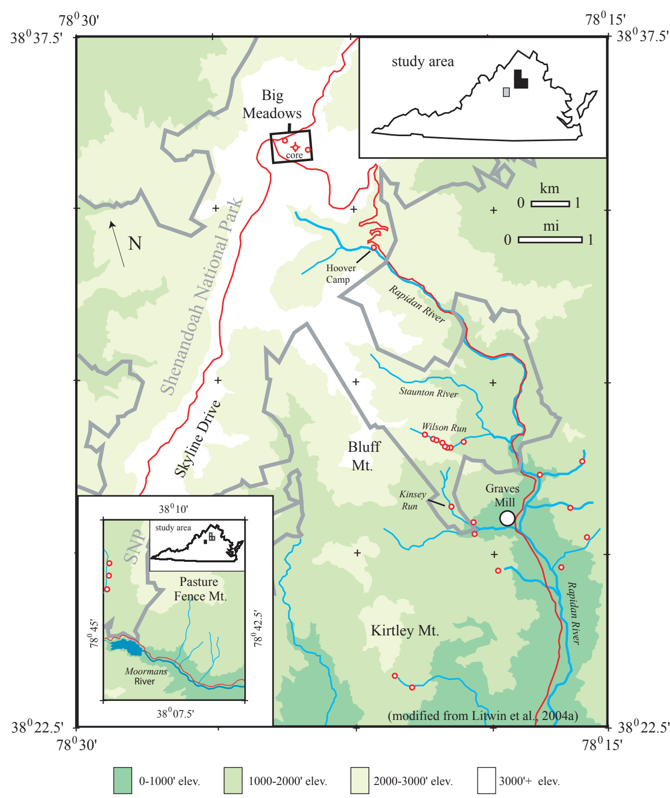


Figure 12. *Site map of study area, Shenandoah National Park and environs, Virginia.*

Figure 13 illustrates a preliminary relative temperature model for the pollen samples that were recovered, and estimates for the forest type that each pollen sample most closely resembled (forest types abbreviated). One of the first observations noted was that the dominant forest type covering the Blue Ridge today, Appalachian oak forest, did not exist in this area for most of the 45 ky of geologic record covered by these samples. The Appalachian oak forest likely first established itself on the Blue Ridge no earlier than ~15 ka, based on the calculated paleo-insolation curve (Litwin and others, 2001; Berger, 1978).

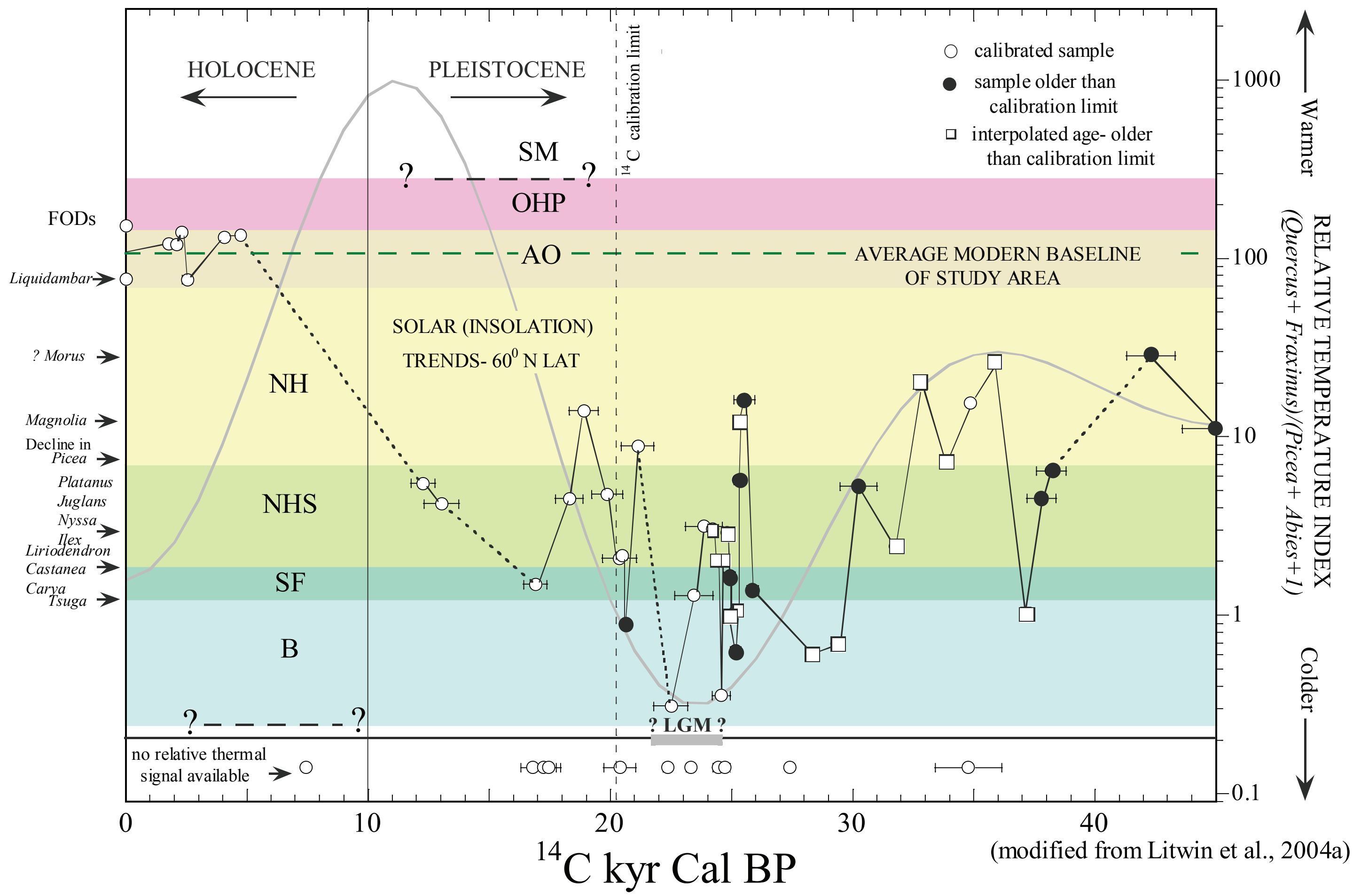


Figure 13. *Blue Ridge dataset summary with reconstruction of prehistoric forest zones, observed first occurrence datums (FODs), and sequence and timing of climate-driven vegetation change (0- 45 ka)..*

This pollen evidence also suggested that during the Last Glacial Maximum (LGM, ~21 to 25 ka) the valleys and foothills surrounding the Blue Ridge most probably were covered completely with boreal forest. With the additional temperature reduction imposed by adiabatic cooling, the ridge crests along the Blue Ridge probably supported only alpine tundra, especially at high elevation sites such as Big Meadows. The modern boreal forest's southerly limit in North America is positioned south of the southerly limit of permafrost. The southerly boreal forest limit occurs at approximately 1.5˚ C MAT (Ecoregions Working Group, 1989; Litwin and others, 2004a). The southerly limit of permafrost occurs at approximately -1.0˚ C MAT (Brown and others, 1997). The vegetation signal from the Blue Ridge pollen data suggested a minimum mean annual temperature shift of ~11˚ C (20˚ F), from the LGM to the present (Litwin and others, 2004a). This thermal difference is bounded between the coldest limit of the present forest in the BM area (Appalachian oak forest, at approximately 12.5˚ C (55˚ F)) and the warmest limit of the coldest forest type suggested by the pollen evidence at Kinsey Run (boreal forest, at approximately 1.5˚ C (35˚ F)).

Evidence of LGM-induced cryogenic processes and cryogenic depositional features occurs throughout in and near Shenandoah National Park. In and around the study area stratified slope deposits are exposed at Hoover Camp (~700 m), and at Kinsey Run (~350' m), which indicate that solifluction and gelifluction processes previously were dominant depositional processes in this area (Smoot, 2004a,b; J.P. Smoot, pers. comm.). Dropstone fabrics (i.e., ice-rafted pebbles and cobbles) also are present in the Kinsey Run section (J.P. Smoot, unpub. data). At Big Meadows, the shallow surface of the BMS has been interpreted as relict sod-bound tundra (J.P. Smoot, pers. comm.). In low-angle light this landform is visible as a series of shallow steps and risers covering the bowl-shaped depression at Big Meadows. Cryogenic processes also dominated the deposition along the Blue Ridge at Black Rock (Eaton and others, 2002). The present landscape in and around the Blue Ridge commonly displays a Holocene-age geomorphic and textural overprinting that incompletely modifies a relict periglacial landscape. In contrast, the fossil pollen evidence suggests that the past floras that have grown on these relict surfaces appear to have transformed regularly in near-equilibrium with past climate shifts over a 45 ky time period (Litwin and others., 2001, 2004b).

The fossil pollen evidence suggests that the vegetation changes on the eastern flank of the Blue Ridge since the LGM have been frequent, abrupt, and strongly expressed. Forests appear to have shifted approximately every ~200 years on average over the past 45 ky (Litwin and others, 2004a). Some of these shifts probably have been quite abrupt. Current analyses of new, long (35 m) high-resolution cores taken along the Potomac River drainage are now be used to test the findings from these Blue Ridge paleovegetation and paleoclimate analyses.

Cumulative Trip Mileage	Point-to-Point Mileage	Directions and *comments*
0.0		Load the buses, find a seat and sit down. Turn right out of bus parking area onto Big Meadows entrance road.
0.1	0.1	Turn right (south) on Skyline Drive (mp 51.2).
14.7	14.6	Cross U.S. Rt. 33 at Swift Run Gap (mp 65.8). *Swift Run Gap is developed along a steeply dipping transverse fault.*
15.9	1.2	*Mass wasting originating in basement rocks exposed on the southeast side of the Skyline Drive blocked the road in 2003. Although the Park Service has attempted to remediate the problem, debris commonly wastes off the cut.*
16.7	1.0	Turn right into Sandy Bottom Overlook (mp 67.8). Unload buses.

STOP 2

Sandy Bottom Overlook (38.3360˚ N, 78.5657˚ W)

Basement Complex- charnockite

Southworth and Bailey

Sandy Bottom Overlook provides a commanding view of the Blue Ridge (northeast), the Shenandoah Valley (west), and Massanutten Mountain (west/northwest). Well-cemented quartz arenites of Silurian age underlie the long linear ridges of the Massanutten Mountain complex. The canoe-shaped prow at the southwestern end of Massanutten reflects the geometry of the gently plunging syncline. Recent work by Shufeldt and others (2008) demonstrates the synclinorium is cut by a number of northwest dipping back thrusts. The mountain front between the Elkton area and north to Stanley is quite sinuous (Figs. 1 and 4). In this area King (1950) recognized the Elkton embayment and Shenandoah salient, but did not place a fault at the foot of the range.

South of Elkton, early Cambrian rocks of the Antietam Formation are brecciated and structurally overlie younger carbonate rocks. The sinuous nature of the mountain front likely reflects the low-angle nature of the Blue Ridge fault and offset associated with younger transverse faults.

Hightop forms the mountain immediately to the east of the overlook (Fig. 14). The "great unconformity" between the Mesoproterozoic basement and Neoproterozoic cover sequence occurs just above the Skyline Drive. The basement is overlain by arkosic phyllite of the Swift Run Formation and at least 250 meters of Catoctin metabasalts. On Hightop, individual Catoctin lava flows dip gently to the southeast, such that the steep northwest flank of the mountain forms the anti-dip slope. The basement and cover sequence are folded into a broad open anticline (Roundtop anticline). Sandy Bottom is underlain by a northwest-striking transverse fault that places basement against the Catoctin Formation (Fig. 14).

Figure 14. *Inset from the Geologic map of the Swift Run Gap 7.5' quadrangle with Stops 2 and 6 (from Bailey and others, 2009).*

The large roadcut exposes a distinctive orthopyroxene-bearing monzogranite. This charnockite is part of a compound pluton that occurs from south of Swift Run Gap to north of Skyland. The rock is composed of 10-30% alkali-feldspar, 30-50% plagioclase, 15-25% quartz, 10-15% orthopyroxene with minor amphibole and clinopyroxene. Rock textures range from porphyritic with alkali-feldspar megacrysts to equigranular. The unit is commonly massive, but does carry a weak foliation at some locations. A U-Pb zircon age of 1,049±9 Ma was obtained from rock at this location and is interpreted as a crystallization age (Southworth and others, 2009). Charnockite exposed immediately east of Hightop Mountain (4 km to the east) yielded a Ar-Ar age from amphibole of 950 Ma indicating that the basement complex in the western Blue Ridge has not been heated above the amphibole closure temperature (450˚ to 500˚ C) since the waning stages of the Grenvillian orogeny.

Cumulative Trip Mileage	Point-to-Point Mileage	Directions and *comments*
16.7	0.0	Bear right (south) on Skyline Drive.
22.3	5.6	Cross Simmons Gap. *Simmons Gap sits along a 10 km-long, steeply dipping transverse fault that strikes 350˚ and extends across the Blue Ridge into the Great Valley.*
31.7	9.4	*Outcrop on the west side of road exposes interbedded granule conglomerate to coarse quartz sandstone and metasiltone/phyllite. of the Weverton Formation. Beds are upright and dip gently to the northwest. Penetrative foliation is well developed in phyllite.*
33.7	2.0	Turn right in Blackrock Summit parking area (mp 84.8). Unload buses and assemble for hike to Blackrock Summit.

STOP 3

Blackrock Summit (38.2200˚ N, 78.7404˚ W)

Harpers Formation- well-cemented quartz sandstone, Quaternary block field

Eaton, Hancock, and Lamoreaux

Blackrock Summit forms an isolated summit tor with outstanding views of the Blue Ridge and Shenandoah Valley. The bedrock crags are composed of a grayish to dusky purple, medium- to coarse-grained, well-cemented quartz sandstone in the Harpers Formation. Well-cemented sandstone (quartzite) is a minor rock type in the silt dominated Harpers Formation (Fig. 15), forming laterally discontinuous lenses up to a kilometer in length and 5 to 15 meters in thickness.

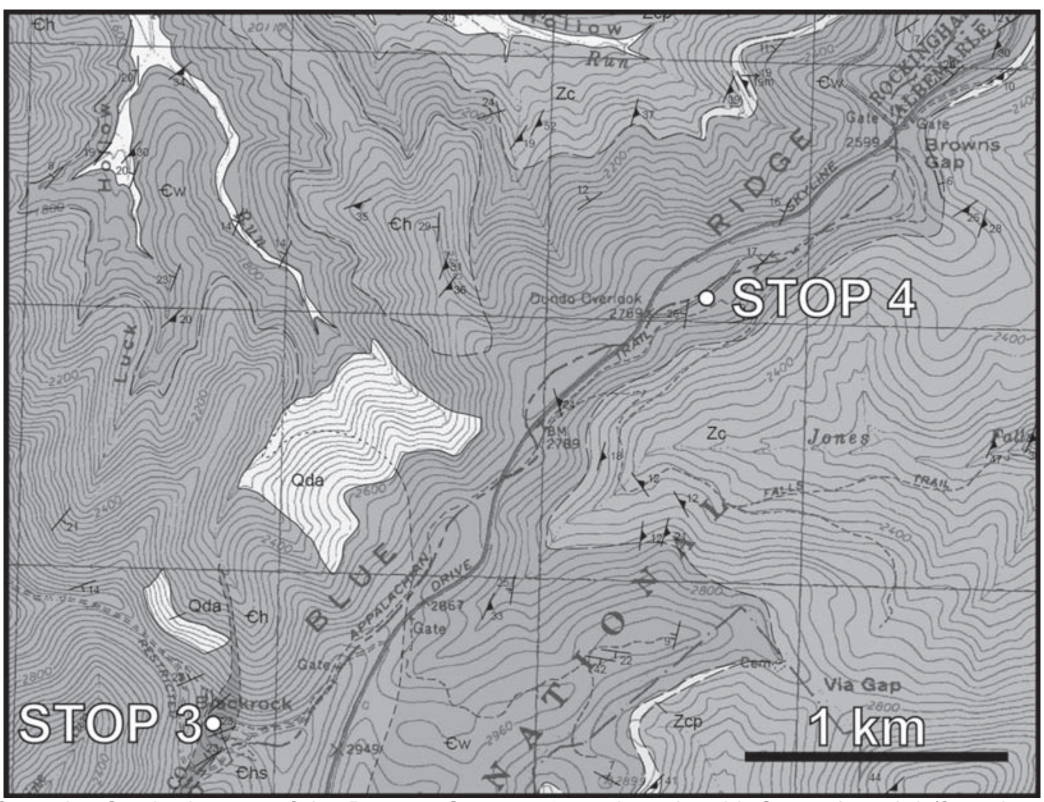


Figure 15. *Inset from the Geologic map of the Browns Cove 7.5' quadrangle with Stops 3 and 4 (from Lamoreaux and others, 2009).*

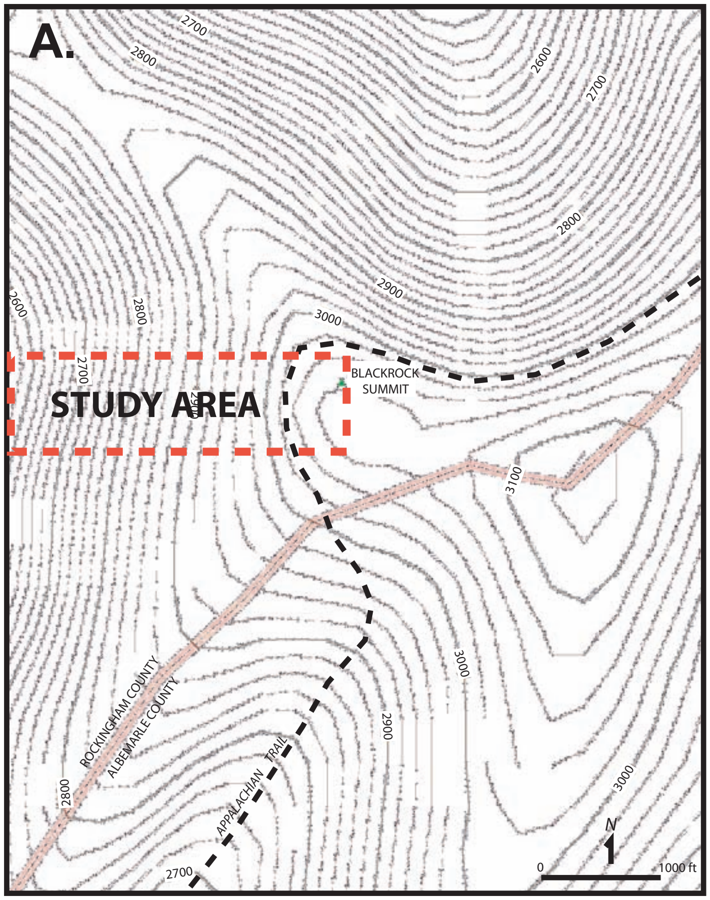
Cross and plane beds are present and the sandstone likely represents a shoaling sequence in the marine. Bedding dips 10˚ to 20˚ to the northwest and is cut by two sets of subvertical fracture sets.

Exposures of bedrock and block slopes near the summits of the Blue Ridge Mountains in central Virginia reveal a variety of deposits suggestive of periglacial slope processes. The summit of Blackrock at an elevation of 933 meters (3,060'), is a shattered tor consisting of gently dipping (12°), broken quartzite blocks and minor, less disturbed bedrock of the Harpers Formation (Fig. 16). The tor, consisting of isolated quartzite columns (3-8 m high), has undergone dislocation by cambering, leaving openings up to several hundred cm between prominent orthogonal joint sets (Fig. 17). Below the ridge top where the slopes gradually increase, some remaining displaced bedrock columns show rotational movement which probably facilitated toppling collapse and production of large amounts of blocky debris. Prominent block slopes extend downward from both the east and west sides of the tor. On the west (dip slope) side the block slope extends over 500 m downslope on gradients that vary between 18-35°. The longitudinal profile of the block slope is an undulating surface of weakly developed benches and escarpments (Fig. 18).

Detailed measurements of clast volume and orientation show four sequences where boulder deposits exhibit an increase in mean clast volume downslope, consistent with discrete events and landforms. The thickest accumulations of debris are associated with large individual blocks that exceed 4 m in length (Fig. 18). The fabric of the deposits is largely open framework, some sites exhibit distinct layers of finer materials underlying coarser materials. The majority of the sites show a preferential downslope clast orientation of the long axis (Fig. 16). The lower half of this block slope narrows into a slightly sinuous path for ~150 m, and abruptly terminates with a distinct snout just above a small first order tributary of Paine Run. Additionally, circular shallow depressions measuring 1-3 m in diameter on the block slopes suggest an ice-rock mixture prior to melting and subsidence of these deposits, producing the depressions.

These landforms and deposits are similar to solifluction and relict rock glacier deposits documented in the literature. Our continuing investigations will focus on the mechanisms and chronology of the slope processes, and their association with climatic events.

Lowering of Blackrock Summit likely occurs through removal of blocks bound by weathered sub- horizontal bedding planes and vertical joints, as well as toppling of isolated tors and direct removal by rain and wind of granular material produced from weathering. To estimate the erosion rate at this site, rock samples were collected the surface of bedrock outcrops, and the abundance of the cosmogenic radionuclide (CRN) ^{10}Be in quartz (Fig. 9). ^{10}Be abundance in surface samples is interpreted as a steady-state erosion rate (Bierman, 1994) and calculations of surface production rate are corrected for elevation, latitude and horizon blockage. The average erosion rate from three surfaces sampled at Blackrock is 17.9 ± 5.1 m/My, which is nearly 10m/My greater than the average for eight other summit sites across the Park.



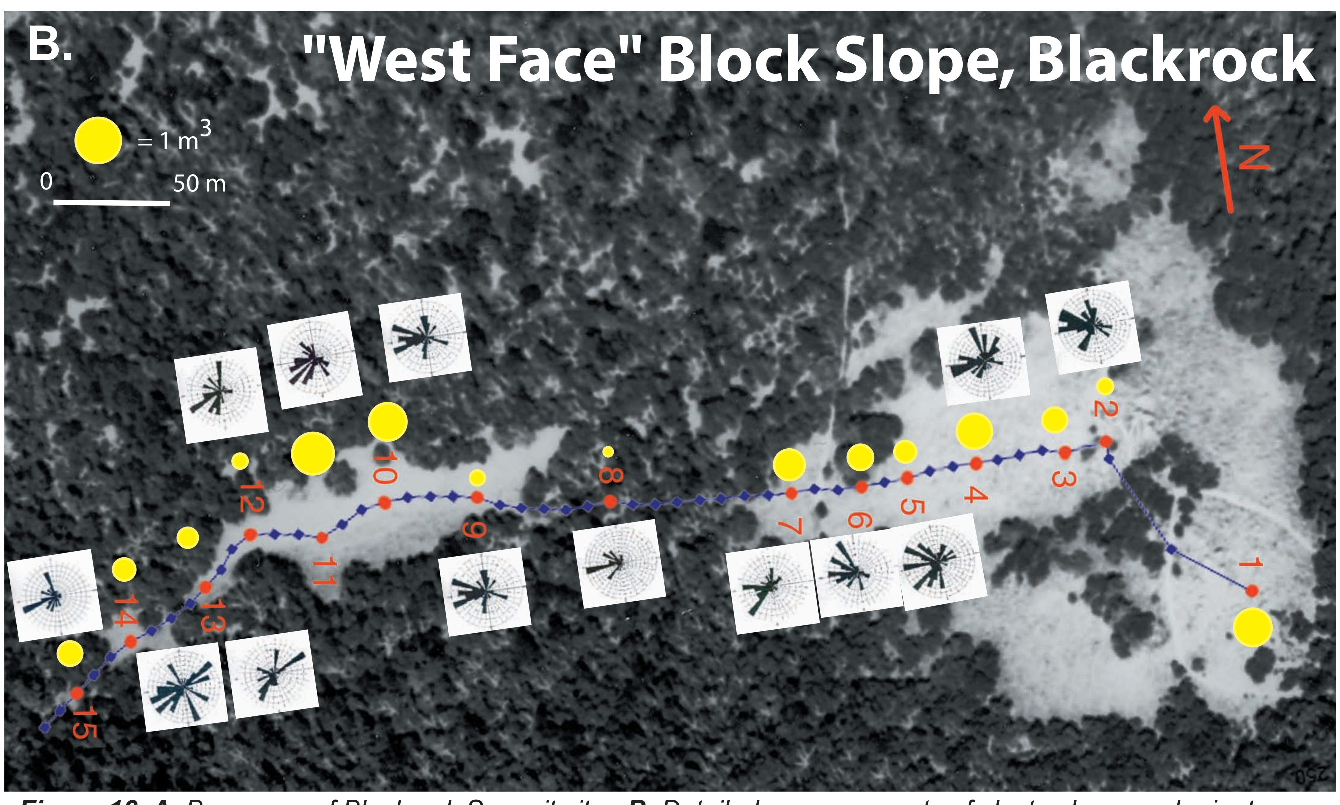


Figure 16. ***A.*** *Base map of Blackrock Summit site.* ***B.*** *Detailed measurements of clast volume and orientation show four sequences where boulder deposits exhibit an increase in mean clast volume downslope, suggesting discrete events and landforms.*

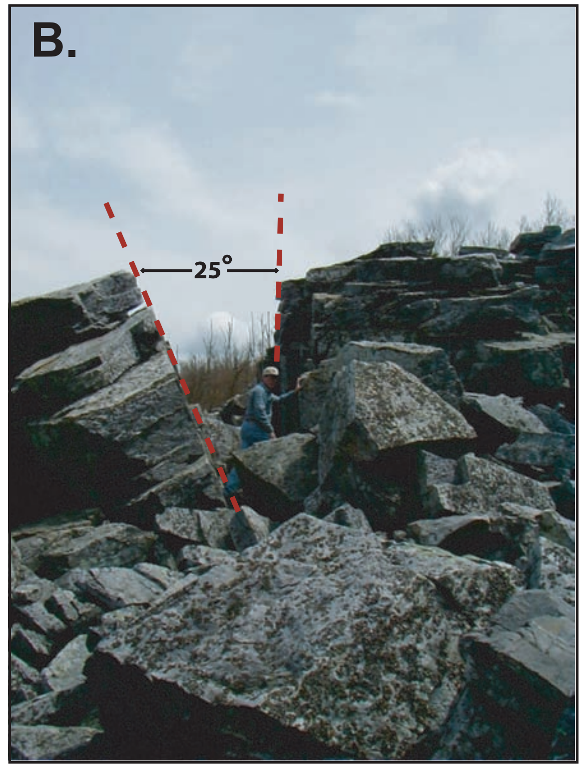
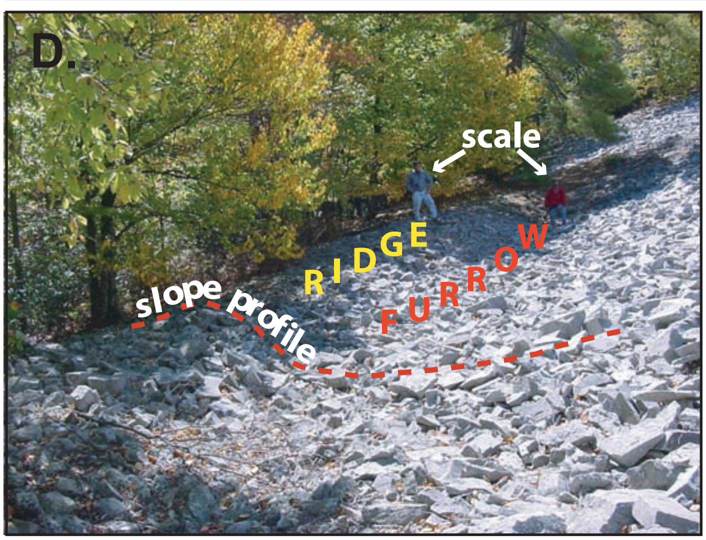
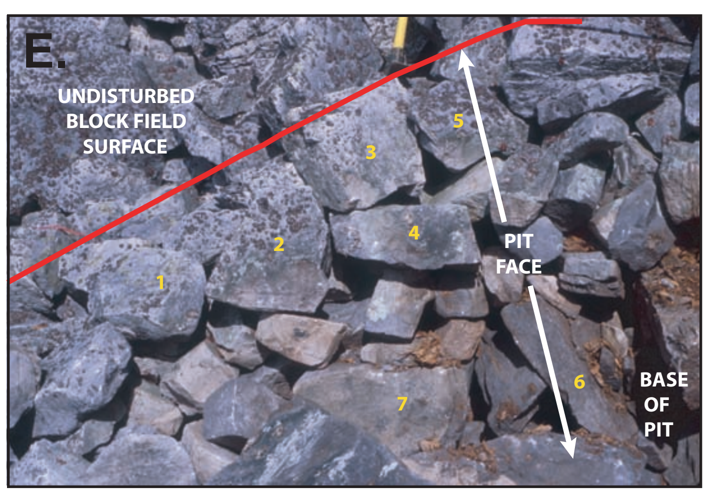


Figure 17 A. *Shattered tor (height ~10 m) Quartzite blocks break along A) bedding planes and B) joint planes.* ***B.*** *Quartzite columns displaced by cambering and toppling. The tilt of the lines demonstrates the toppling of the nearly horizontal beds after cambering has opened and separated the blocks along joint planes.* ***C.*** *Prominent block slopes west of Blackrock summit.* ***D.*** *"Curious" ridge and furrow feature at the toe of the block slope shown in C. Note the geologists for scale. Landform may be a protalus rampart.* ***E.*** *Excavation pit of boulder front individual numbered boulders are identified in Figure 18. The line marks the upper boundary of the excavation.*





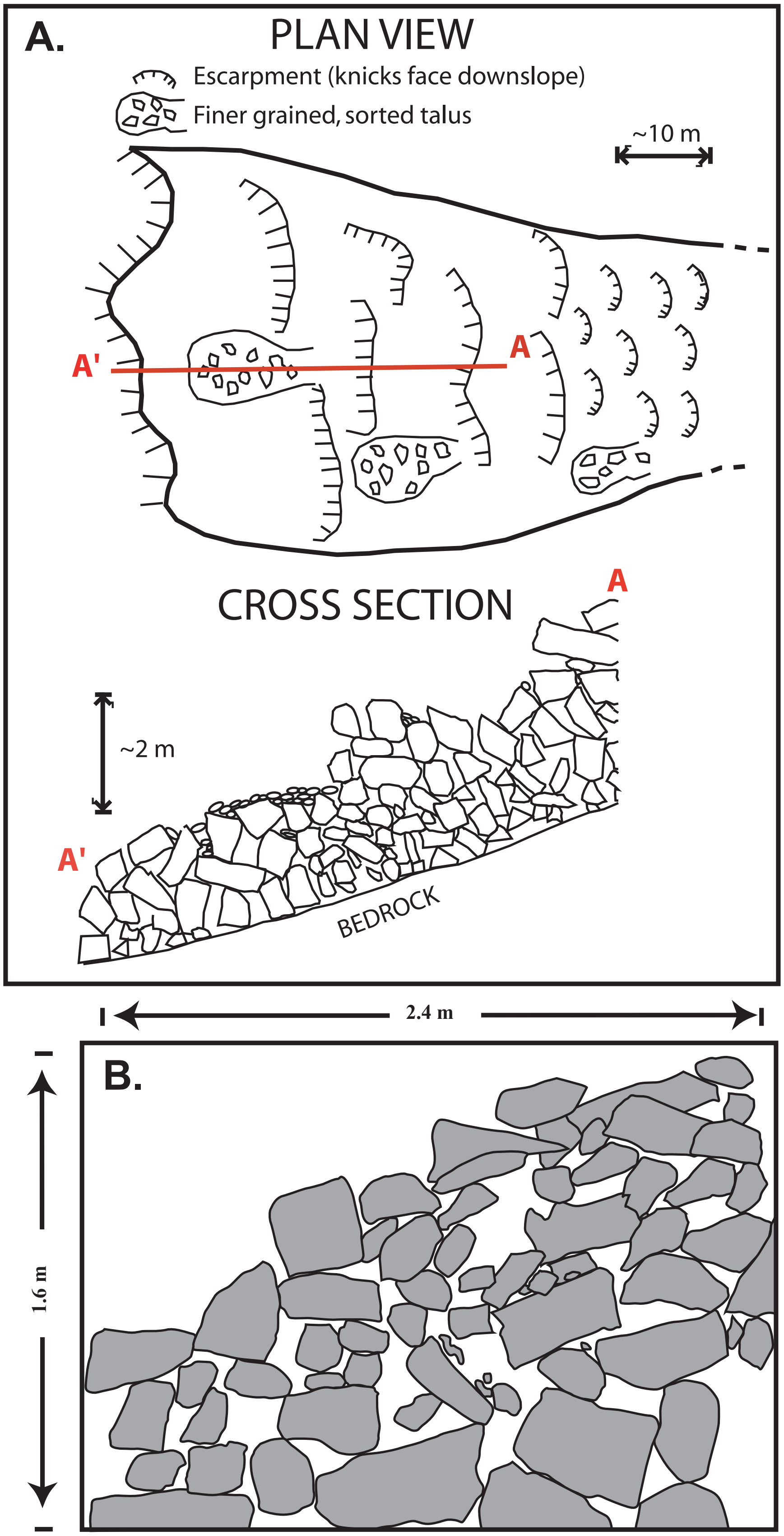


Figure 18. *A. Plan and cross section views of the topography depicting weakly-developed benches and escarpments.* ***B.*** *Longitudinal profile of boulder front. The fabric of this deposit is dominantly open framework.*

Cumulative Trip Mileage	Point-to-Point Mileage	Directions and *comments*
33.7	0.0	Load buses and go north on Skyline Drive (mp 84.8).
34.8	1.1	Turn right into Dundo Picnic area (mp 83.7).
35.1	0.3	Park and unload buses.

STOP 4
Dundo Picnic area- Lunch (38.2349˚ N, 78.7180˚ W)
Weverton Formation- well-cemented coarse-grained sandstone
Southworth and Lamoreaux

The Dundo Picnic area will serve as our lunch stop; low outcrops of well-cemented coarse-grained sandstone to granule conglomerate in the Weverton Formation occur immediately to the northwest of the picnic area and along the Appalachian Trail to the east of the picnic area (Fig. 15). Coarse-grained sandstones in the Weverton Formation underlie much of the Blue Ridge's crest in the area, although blue-gray quartzose phyllite makes up a significant part of the Weverton Formation in the southern part of Shenandoah National Park. Bedding dips gently to the northwest and the penetrative foliation dips to the southeast.

In the 1960s wells were drilled to obtain water for the Dundo site. A 190-meter deep test well at the picnic area penetrated 42 meters of quartzose phyllite, metasandstone and metaconglomerate interpreted as the Weverton Formation, 30 meters of chlorite schist, phyllite, and conglomerate underlain by 116 meters of metabasaltic greenstone collectively interpreted as the Catoctin Formation (DeKay, 1972). The well yielded 25 gallons per minute and the static water level was 110 meters below the surface.

A purple to maroon spotted phyllite (with the spots being sericite and chlorite blebs) is commonly, but not everywhere present, at the top of the Catoctin Formation. The chemistry and texture of the spotted phyllite indicates that this rock originated as a basaltic metatuff. The absence of a laterally persistent spotted phyllite beneath the siliciclastic rocks of the Weverton Formation is consistent with an unconformable contact between the Catoctin Formation and Chilhowee Group. Traditionally, the Swift Run/Catoctin sequence upwards through the Chilhowee Group has been interpreted to record the rift to drift transition associated with the opening of the Iapetus Ocean, however the Catoctin/Weverton unconformity requires modification of this tectonic model.

Cumulative Trip Mileage	Point-to-Point Mileage	Directions and *comments*
35.1	0.0	Load buses and continue out of picnic area
35.5	0.4	Turn right (north) on to Skyline Drive.
38.3	2.8	*Outcrop on the southeast side of road exposes purple, porphyritic metabasalt of the Catoctin Formation.*
41.1	2.8	Turn left into Rockytop Overlook (mp 78.1). Unload buses.

STOP 5
Rockytop Overlook (38.2789˚ N, 78.6651˚ W)
Harpers Formation- phyllitic siltstone and thinly bedded metasandstone
Bailey and Hancock

The view of the Big Run watershed (to the south and west) is expansive. Big Run forms the largest drainage

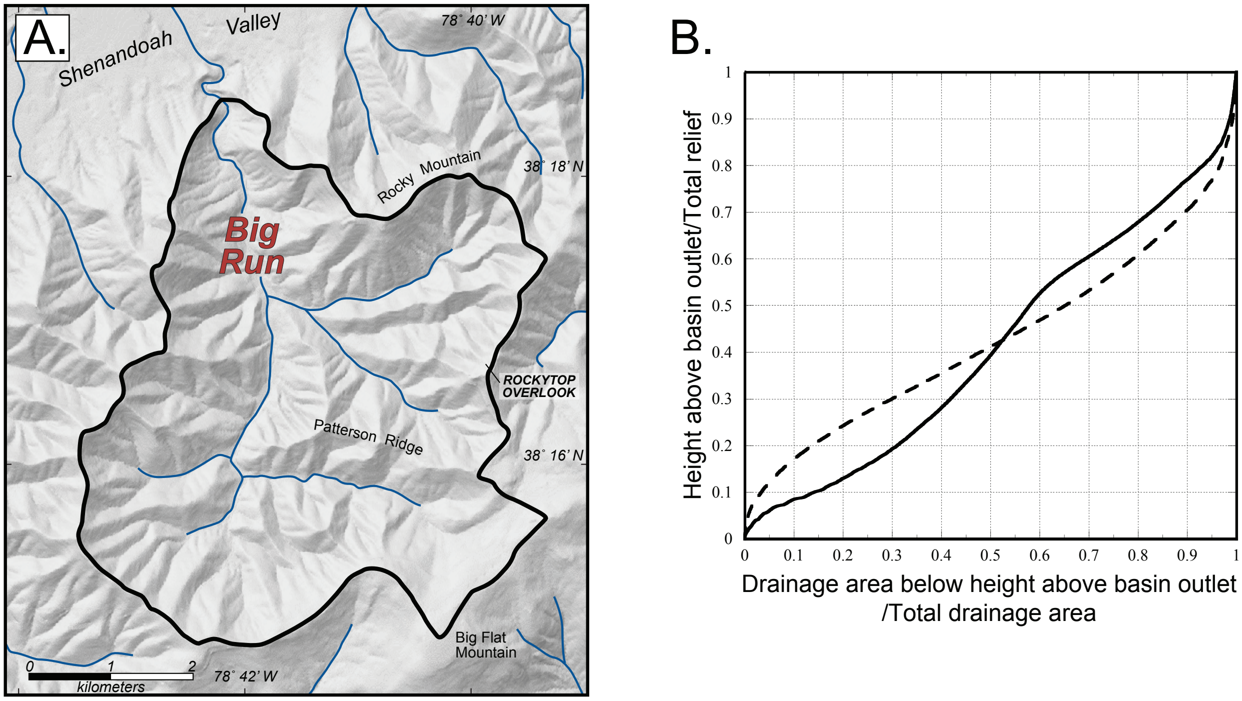


Figure 19. *A. Shaded relief map of Big Run drainage basin.* ***B.*** *Hypsometric curves for the Big Run (dashed line) and Doyles River (solid line).*

entirely within the Park (Fig. 19A). Most of the drainage basin is underlain by Chilhowee siliciclastics (mostly Harpers Formation metasiltstone), although greenstone of the Catoctin Formation crops out at the base of the slopes in the southern part of the basin. Bedding generally dips to the northwest, but a series of north-northeast plunging, asymmetric folds occur in the Antietam Formation along Brown and Rocky mountains.

As is typical for drainage basins developed in the Chilhowee Group, slopes are steep with ~60% of the Big Run basin having slopes in excess of 35% (20˚ slope). The elevation/area relationship (hypsometric curve) for the Big Run basin is quite different from the neighboring Dolyes River drainage (developed primarily on the Catoctin greenstone and basement complex) (Fig. 19B).

The large roadcut on the east side of the Skyline Drive exposes typical Harpers Formation lithologies. Brown to grayish phyllitic siltstone is predominant, but thinly bedded sandstone is also present. A penetrative northeast-striking foliation (cleavage) is well developed and dips moderately to the southeast, bedding is

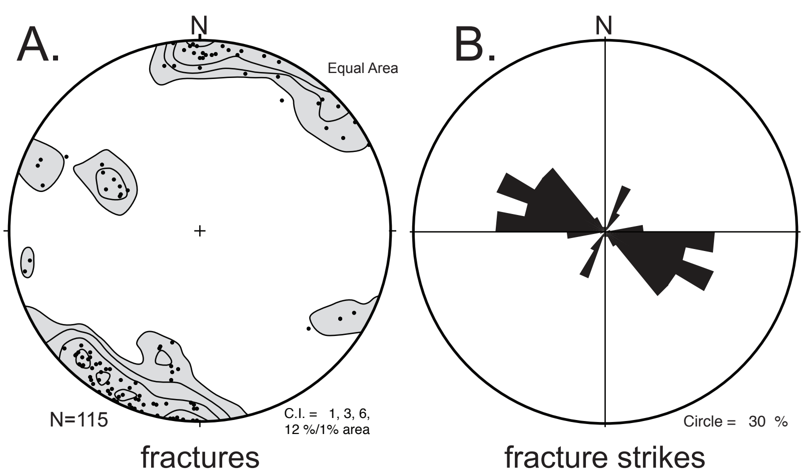


Figure 20. ***A****. Contoured stereogram of poles to fractures in the Harpers Formation between miles 76-79* ***B****. Rose diagram of fracture strikes 10˚ bins.*

upright and dips gently to the northwest at the southern end of the roadcut. A set of steeply dipping joints cut the rock; the dominant set strikes to the west-northwest (Fig. 20) and individual fracture surfaces are commonly ornamented with plumose structures (best viewed when the fracture surface is obliquely illuminated). West-northwest-striking extension fractures are common in the Chilhowee Group. This fracture set formed after the Blue Ridge cover sequence was folded and developed under a regional stress field with the maximum compressive stress oriented ~290˚±10˚.

Cumulative Trip Mileage	**Point-to-Point Mileage**	**Directions and *comments***
41.1	0.0	Load buses and Turn left (north) on to Skyline Drive.
44.7	3.6	Loft Mountain Overlook (mp 74.5) *A diverse array of lithologies are exposed in strongly deformed Catoctin Formation at the overlook. Nice view to the southeast with Simmons Gap fault between overlook and prominent foothills that are underlain by the basement complex.*
49.3	4.6	Cross Powell Gap. *Powell Gap sits along a steeply dipping transverse fault bounding the southwest margin of the Bacon Hollow graben. The gap is underlain by granitoid gneiss; phyllitic arkosic of the Swift Run Formation is exposed in roadcuts 0.2 miles to the south.*
49.9	0.6	Turn right into Bacon Hollow Overlook (mp 69.3). Unload buses.

STOP 6
Bacon Hollow Overlook (38.3215˚ N, 78.5822˚ W)
Basement complex- granitoid gneiss and low-silica charnockite
Bailey and Southworth

Bacon Hollow Overlook is cut into the southern slopes of Roundtop Mountain and affords views to the southeast. In the far distance, approximately 40 km away, the Southwestern Mountains form a low linear ridge underlain by the Catoctin Formation on the southeastern limb of the Blue Ridge anticlinorium. Neoproterozoic metasedimentary rocks of the Lynchburg Group and the basement complex underlie the low terrain in the middle distance. In the foreground, the southern rampart of Hightop is to the left, Bacon Hollow in the center, and Flattop Mountain to the right (note the eclectic array of buildings dotting the upper slopes of Flattop).

Bacon Hollow forms a steep-walled, 500-meter deep valley drained to the south by the Roach River. The valley floor and lower slopes are underlain by granitoid gneiss with an east-striking high-grade foliation. The basement is overlain, on both Hightop and Flattop, by the Swift Run and Catoctin formations. A set of newly recognized, north-northwest striking faults bound the U-shaped valley and pass beneath Powell and Smith Roach gaps (Fig. 14) (Bailey and others, 2009). Slip on these steeply dipping faults appears to have "down dropped" rocks in Bacon Hollow (Fig. 14). Bacon Hollow is, in essence, a graben. Total displacement on the bounding faults is <100 m and considerably less than the modern topographic relief. The north-northwest-striking faults are parallel to Jurassic diabase dikes in the region. We posit that these faults were active during Mesozoic extension, albeit along the western (distal) margin of the rift. Bacon Hollow may have been a structural trough in the Mesozoic and the modern drainage inherited from the structural geometry of the area. In the 19th and early 20th century copper was mined along the southern rampart of Hightop, this deposit occurs in the Catoctin Formation and may be localized along fractures associated with the north-northwest-striking transverse faults.

Granitic gneiss is exposed at the overlook and along the large cuts on the Skyline Drive. The rock is composed of 30-40% perthite, 30-40% quartz, 5-20% plagioclase and up to 15% biotite. This rock forms an extensive unit along the eastern slopes of the Blue Ridge in southern Shenandoah National Park. The chemical composition of this unit ranges from alkali-feldspar granite to monzogranite and at some locations the rock is leucocratic. A U-Pb zircon age of 1,150±23 Ma was obtained from this exposure and the granitic gneiss is part of the older plutonic suite (Southworth and others, 2009). A charnockite dike intrudes the granitic gneiss in the cut ~120 meters southwest of the overlook and the summit of Roundtop is underlain by charnockite (observed at Stop 2, Fig. 14)

A medium- to coarse-grained gneissic foliation is evident. This fabric is defined by mm- to cm-scale, elongate aggregates of feldspar and quartz (individual grains are subequant) and aligned biotite that formed under high-grade (but solid-state) conditions. At this location foliation strikes east-northeast to east and dips 25˚ to 50˚ to the north. In the Swift Run Gap quadrangle high-temperature foliation in the granitic gneiss is folded (Fig. 21A); this fabric was formed and folded prior to the intrusion of the ~1,050 Ma charnockitic pluton. A younger foliation, defined by aligned greenschist minerals, is also developed at many locations in the basement complex. In contrast to the high-temperature fabric, the greenschist-facies foliation consistently strikes to the northeast and dips moderately southeast (Fig. 21B).

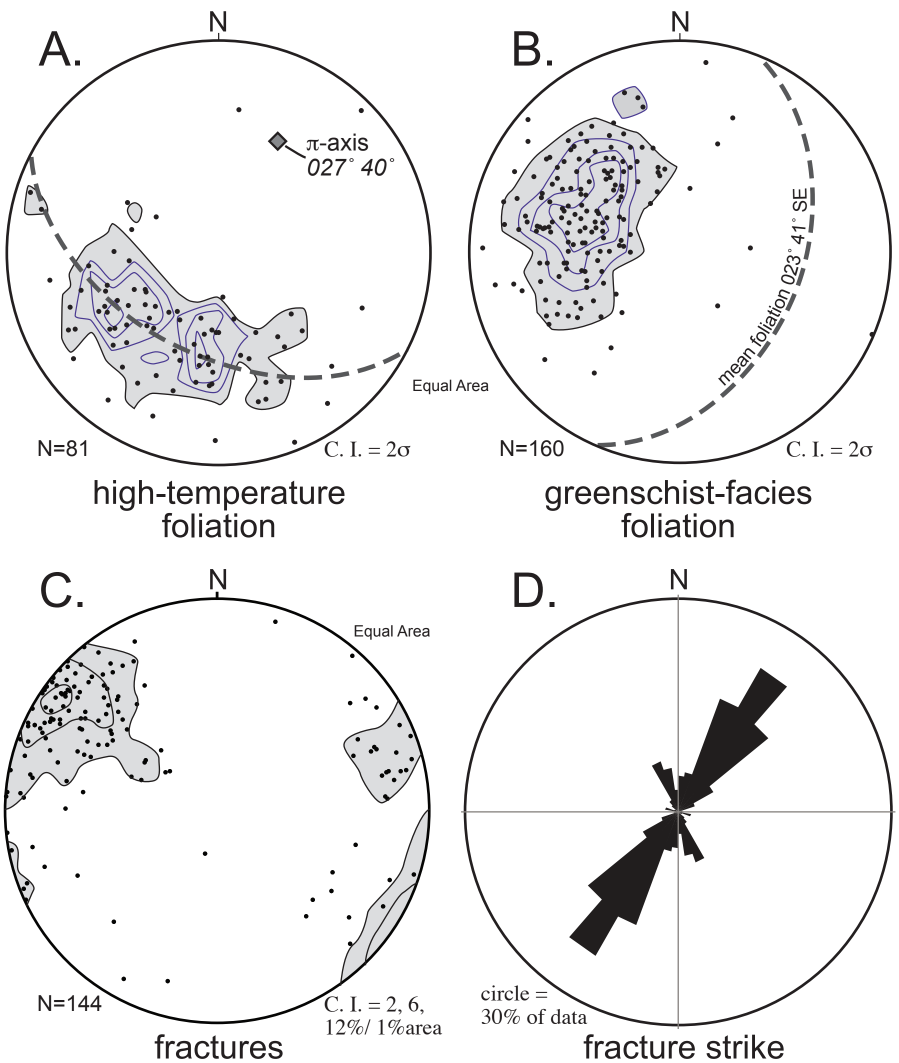


***Figure 21. A.** Contoured stereogram of poles to high-temperature foliation. **B.** Contoured stereogram of poles to greenschist-facies foliation. **C** & **D**. Contoured stereogram of poles to fractures in the granitic gneiss*

The granitic gneiss is cut by numerous fracture sets and in the exposures along the Skyline Drive northeast-striking fractures are common (Fig. 21C, D). These fractures display a wide range of dips. A number of northwest-striking and northeast-striking fractures are coated with Fe-oxide slickensides that record hanging wall down slip. The significance of these shear fractures is unclear. Regionally, the basement complex displays much greater variably in fracture orientation than the cover sequence.

Cumulative Trip Mileage	Point-to-Point Mileage	Directions and *comments*
49.9	0.0	Load buses and turn right (north) on to Skyline Drive.
53.4	3.5	Cross U.S. Rt. 33 at Swift Run Gap (mp 65.8).
54.8	1.4	Hensley Hollow Overlook (mp 64.4). *Outcrop on southeast side of road exposes well-foliated arkosic Phyllite in the Swift Run Formation. Immediately to the southwest of the overlook, foliated and altered basement charnockite is exposed. The contact is interpreted to be unconformable at this location.*
56.5	1.7	Turn right into South River picnic area (mp 62.7).
56.9	0.4	Park and unload buses.

REST STOP
South River Picnic Area (38.3813˚ N, 78.5173˚ W)
Catoctin Formation- foliated greenstone
Burton

Cumulative Trip Mileage	Point-to-Point Mileage	Directions and *comments*
56.9	0.0	Load buses and continue out of picnic area.
57.1	0.2	Turn right (north) on to Skyline Drive.
63.0	5.9	Pull off on right (east) side of Skyline Drive across from Meadow School parking area (mp 56.8). Unload buses and assemble for hike.

STOP 7
Bearfence Mountain Traverse
Bailey

Stop 7 involves a hike along the Appalachian Trail over Bearfence Mountain. The total distance of the hike, from the drop off point to the pick up point, is 2 km (1.4 mi.) with a vertical climb of ~100 m (300 ft). The Appalachian Trail is well maintained and the hike requires a modest level of effort. Those who **do not** wish to participate in the hike should remain with the buses and travel north to the pick up point. Figure 22 is a 1:4,000 scale geologic map of the stop and should be consulted as a guide to locating the described outcrops. The decimal latitude and longitude of the outcrops is given using the NAD 27 datum.

Hike north along the Appalachian Trail. Prior to the first switchback there is abundant float of arkosic phyllite from the Swift Run Formation. Beyond the first switchback, small outcrops of basement granitoid occur.

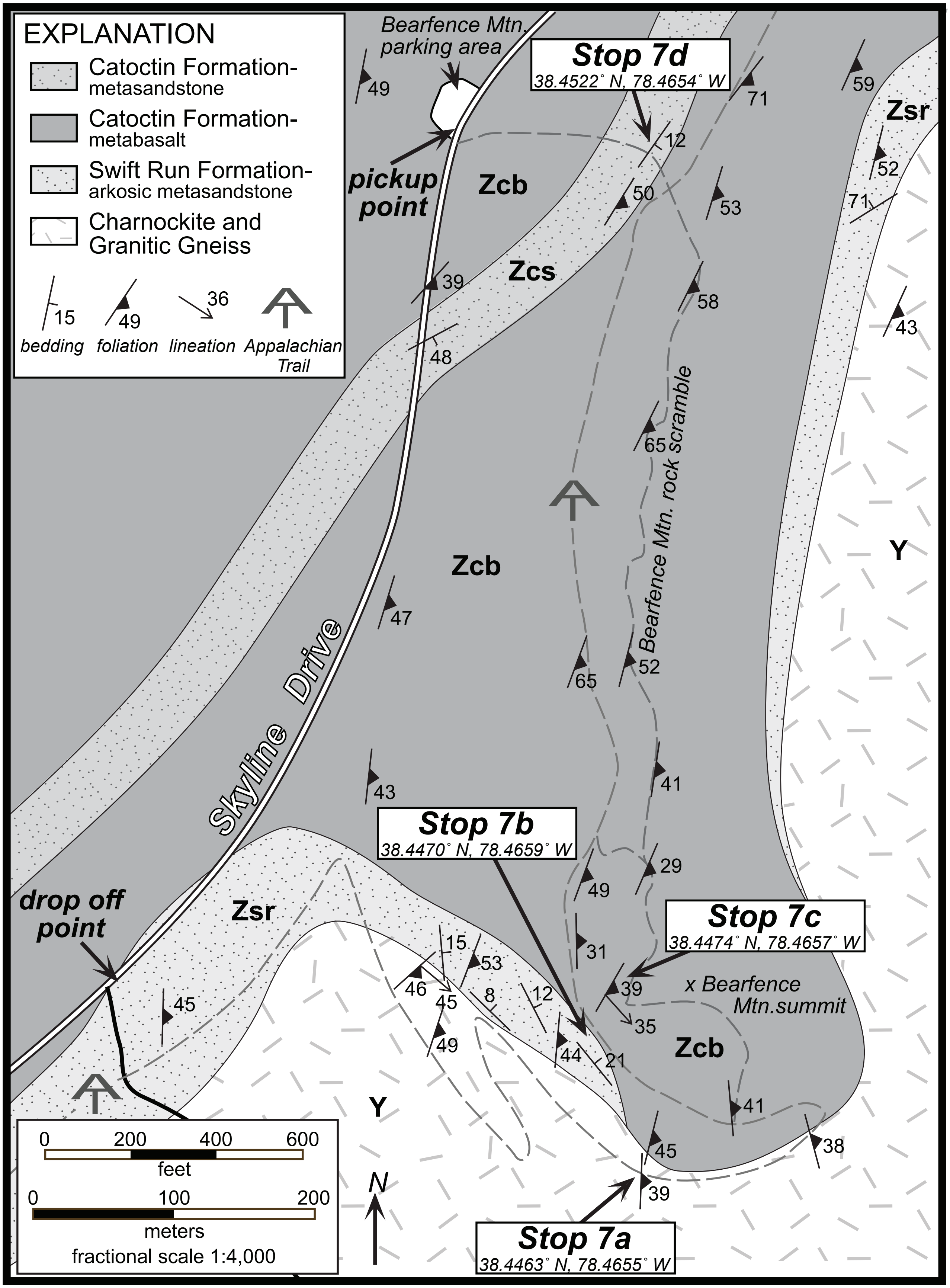


Figure 22. *Geologic map of Bearfence Mountain area.*

Pass the third switchback and continue ~160 m onwards to a large outcrop immediately south (right) and just below a rock wall on the trail (Fig. 22).

STOP 7A
(38.4463˚ N, 78.4655˚ W)
Basement complex- altered granitoid gneiss

The overhanging outcrop of granitoid gneiss is strongly epidotized (Fig. 23A). Epidote is a secondary mineral produced by the reaction of Fe-bearing minerals and feldspar in the presence of water. Prior to alteration the basement was a pyroxene-bearing charnockite, but few mafic minerals (other than epidote) are still present. The granitoid gneiss is cut by numerous hematite-stained fractures that are cut by white quartz veins. Foliation strikes to the north and dips ~40˚ east. Did the foliation develop prior to alteration?

This exposure is less than 20 meters (map distance) from the overlying greenstone. One possibility is that epidote was generated by near-surface hydrothermal alteration when the overlying Catoctin lavas were erupted in the Neoproterozoic. Another possible scenery involves the formation of epidote during regional metamorphism and deformation in the Paleozoic. In Shenandoah National Park altered epidote-rich basement is common near the "great unconformity", but is not confined to the contact zone. At a number of locations east of the Park epidote-rich coronas form around leucocratic basement inclusions that are surrounded by biotite-rich granitoids, and clearly developed during regional metamorphism (Wadman and others, 1997).
Continue upwards along the Appalachian Trail. Pass low outcrops of altered basement along the trail, to the left (uphill) greenstone float of the Catoctin Formation is common. At ~140 m (0.1 mi) the trail goes through a switchback and outcrops of greenstone occur. At this location the Catoctin Formation is directly above the basement and siliciclastic rocks of the Swift Run Formation are not evident. At 60 m beyond the switchback the Appalachian Trail is joined on the right by a trail to the summit of Bearfence Mountain. Continue along

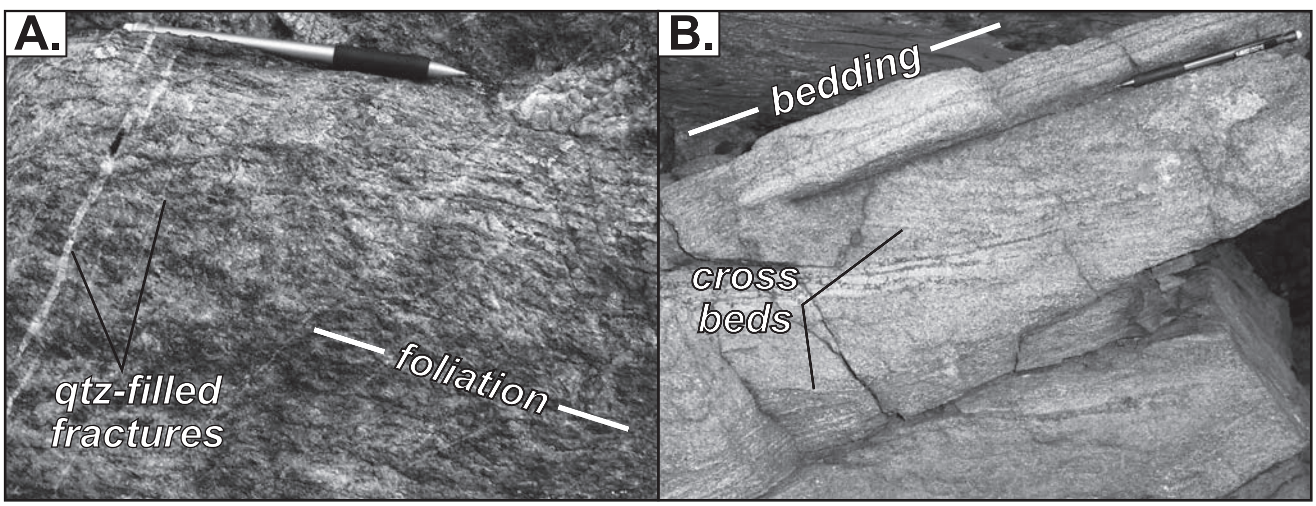


Figure 23. A *Altered granitoid gniess at stop 7a. B Arkosic meta-sandstone at stop 7b.*

the Appalachian Trail, the trail is nearly flat over this section. 120 m past the trail junction flagging denotes the way to outcrops 10 to 20 m southwest (left) of the Appalachian Trail (Fig. 22).

STOP 7B
(38.4470˚ N, 78.4659˚ W)
Swift Run Formation- arkosic metasandstone

Toppled blocks of Catoctin greenstone directly overlie outcrops of arkosic metasandstone in the Swift Run Formation. The medium- to coarse-grained Swift Run metasandstone contains abundant feldspar and quartz with rock fragments and heavy minerals forming a minor component. Cross and plane bedding is common

and graded bedding evident at a few locations (Fig. 23B). These rocks are clast supported, well to moderately sorted, and contain much less sericitic mica (phyllitic) than most Swift Run lithologies. The provenance is the granitic basement complex and sedimentary structures are consistent with a fluvial to fluvial braidplain depositional environment.
Although the Swift Run Formation is missing to the southeast (near Stop 7A), there is at least 50 meters of metasandstone exposed to the west-northwest (down slope) above the basement complex. Dramatic thickness changes over a short distance are a characteristic of the Swift Run Formation and could reflect significant local topography at the time of deposition or later removal of an originally more laterally expansive unit.

Bedding is upright and dips 10˚ to 20˚ to the northeast (Fig 23B). A penetrative foliation dips ~45˚ to the southeast and is refracted across grain-size changes in beds. Ar-Ar ages from syn-tectonic sericite in Swift Run lithologies exposed to the southeast of the Park range from 350 to 320 Ma (Bailey and others, 2007; Gattuso and others, 2009) and demonstrate that the penetrative deformation (foliation forming) event in the western Blue Ridge occurred prior to or at the earliest stage of the Alleghanian orogeny (325 to 280 Ma).

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Return to the Appalachian Trail and continue north (left), passing the base of greenstone cliffs on the right. The trail descends gently and 140 m after stop 7B a side trail intersects on the right (Fig. 22).

To journey on to Stop 7C (time and spirit permitting) take the side trail upwards to the right. 50 m on the trail joins the Bearfence Mountain summit trail, continue to the south for another 80 m to the rocky promontory (although great views also occur along the rock spine to the north).

STOP 7C
(38.4474˚ N, 78.4657˚ W)
Catoctin Formation- foliated greenstone

This location provides quality vistas of the western Blue Ridge and Massanutten Mountain beyond. The rock underfoot is Catoctin greenstone. Foliation strikes towards 030˚ and dips 45˚ southeast. Chlorite blebs, evident on exposed foliation surfaces, are elongate and define a downdip lineation that likely corresponds to the direction of maximum stretching. Among the greenstone cliffs on the Bearfence crest, foliation dips vary between 70˚ and 30˚, and in places, well-developed shear bands are present. Badger (1999) reports columnar joints from a number of spots along the Bearfence Mountain ridge crest.

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Retrace steps and return to the Appalachian Trail. Bear right and continue north, the trail descends and greenstone cliffs are above and to the right of the trail. After ~440 m the Appalachian Trail intersects the Bearfence Mountain trail again. Turn left (northwest) onto this trail. After 20 m step off the trail to the right (Fig. 22).

STOP 7D
(38.4522˚ N, 78.4654˚ W)
Catoctin Formation- arkosic metasandstone

Laminated to thinly bedded quartzose metasandstone and pebble meta-conglomerate are exposed in the low outcrop. The rock is quite similar to the Swift Run metasandstone at Stop 7B. Bedding is upright and dips ~15˚ southeast, whereas foliation dips 45˚ to 50˚ southeast. Previous workers have interpreted these metasandstones as part of the Swift Run Formation exposed in the nose of a northeast-plunging overturned anticline. Field relations indicate that greenstone occurs both above and below the metasandstone layer (Fig. 22). An alternative interpretation, consistent with the field data, is that the metasandstone forms a layer within the Catoctin Formation.

Metasedimentary rocks in the Swift Run and Catoctin formations are comparable and sourced from the same protolith; lithologic similarity and geometric relations indicate the Swift Run and lower Catoctin are contemporaneous units. Field relations at Bearfence Mountain serve to highlight the difficulty of mapping Blue Ridge formations based solely on lithology.

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Continue west and gradually downhill for 150 m, emerge from the woods to the Skyline Drive and the Bearfence Mountain parking area.

Cumulative Trip Mileage	Point-to-Point Mileage	Directions and *comments*
63.0	0.0	Buses will continue north on Skyline Drive.
63.4	0.4	Bearfence Mountain parking area (mp 56.4) Load buses and proceed north on Skyline Drive.
68.5	5.1	Turn left off Skyline Drive into Big Meadows complex.
68.6	0.1	Turn right towards parking area for Byrd Visitors Center.
68.8	0.2	Park in bus parking area. Unload buses and say goodbye.

End of Road Log

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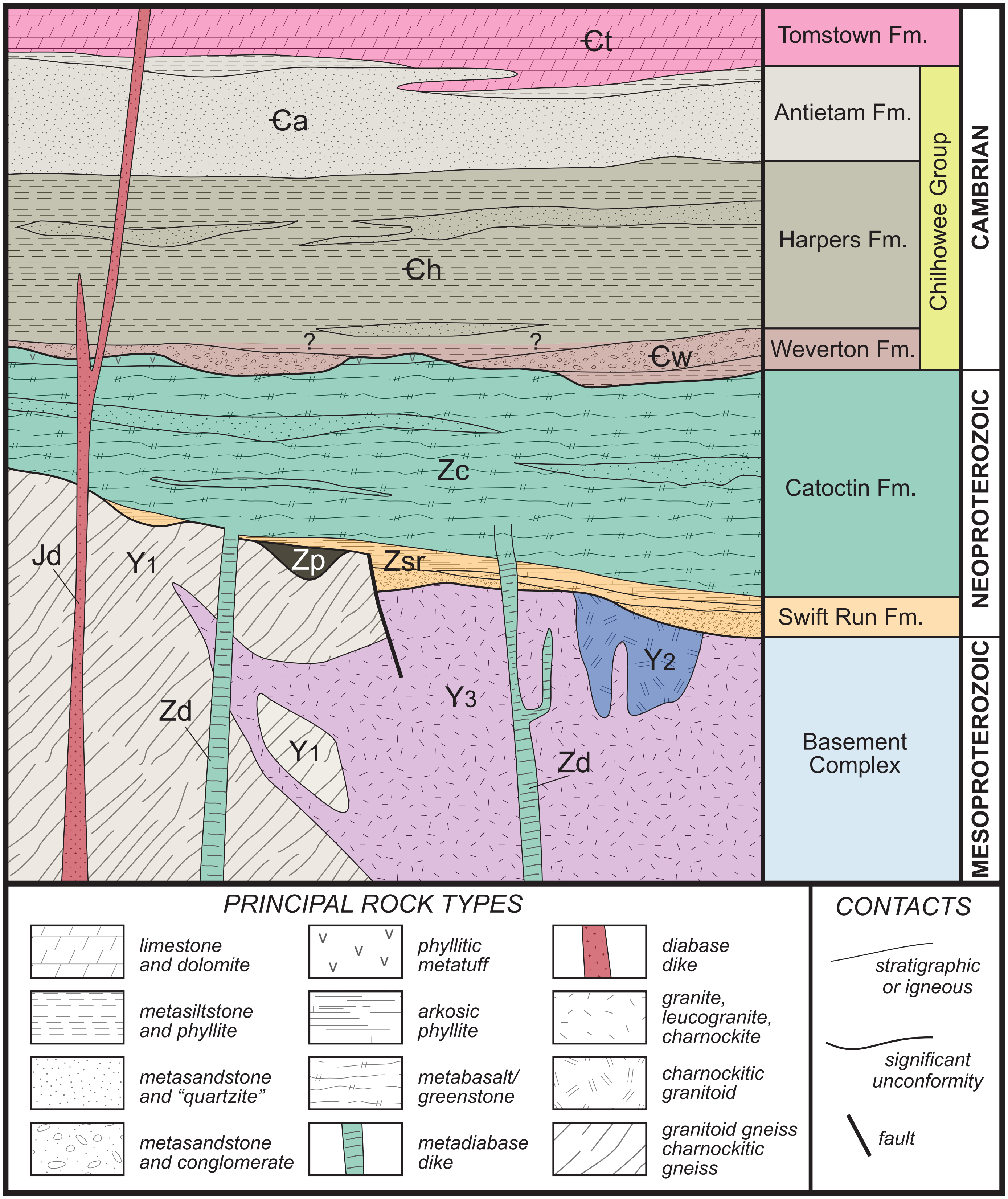

Figure 3. *Generalized stratigraphic section of rock formations in the Shenandoah National Park region, Virginia. Based in part on Figure 6 from Gathright (1976). Zp- Neoproterozoic garnet-graphite paragneiss.*

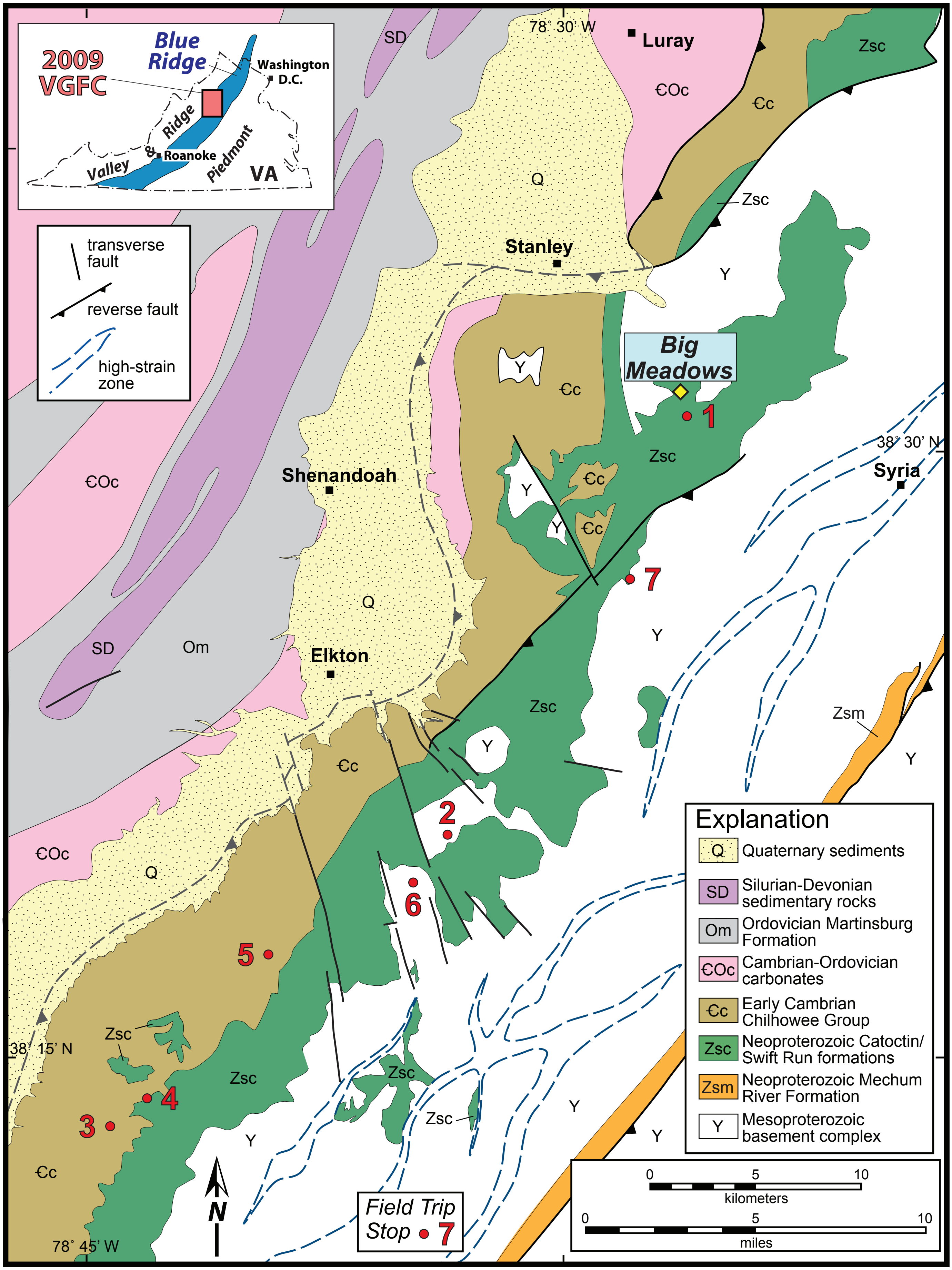

Figure 4. *Simplified geologic map of the Shenandoah National Park region and stop locations at the 2009 Virginia Geological Field Conference.*