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SURFICIAL GEOLOGY OF SHAVER HOLLOW, SHENANDOAH NATIONAL PARK

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INTRODUCTION

At the request of Shenandoah National Park and the Department of Environmental Sciences at the University of Virginia, the US Geological Survey has completed an examination and map of the surficial deposits in Shaver Hollow. The work was carried out as part of the US Geological Survey - National Park Service cooperative agreement implemented in 1994. Shaver Hollow is a small, well defined drainage basin on the west slope of the Blue Ridge about 6.5 miles south of Thornton Gap and can be reached by trail from mile 37.9 on the Skyline Drive. The hollow is drained by the North Fork of Dry Run, and the watershed within the Shenandoah National park is only 2 square miles in area. The area has been the site of extensive investigations by faculty and students at the University of Virginia and by NPS scientists and investigators studying the interaction of atmosphere chemistry, water composition, and the biota of the hollow (Furman and others, written communication, 1997). Modeling of the chemistry of Dry Run surface water, based on atmospheric, biologic, and geologic data, has been attempted with limited success. Better understanding of the surficial deposits and the interaction of streams and springs with near surface materials is needed before more sophisticated models can be devised. Although the bedrock lithology was mapped at a small scale (1:62,000-scale; Gathright, 1976) no examination of the surficial deposits of the hollow was made. The description of deposits contained herein is based on field observations carried out in September - November, 1996. Also included with this report is a 1/12,000-scale map of the surficial geology of Shaver Hollow (figure 1).

PHYSIOGRAPHY

The Shenandoah National Park extends for 105 miles along the summit of the Blue Ridge from Front Royal to Waynesboro, Virginia. The Blue Ridge is a prominent physiographic feature because its principal rocks are metavolcanic lavas and quartzites that are extremely resistant to both chemical and physical weathering (Hack, 1965). The area east of the ridge is underlain primarily by granitic rocks that form a rolling piedmont topography characterized by small, irregular hills and broad valleys. To the west is the prominent, northeast trending Shenandoah Valley, underlain by lower Paleozoic carbonate and clastic rocks. The topographic relief in the Shenandoah National Park is from 2,000 to 3,000 feet on both east and west sides.

For much of the length of the Blue Ridge, the topography consists of a broad, smooth upland; major tracts of land along the ridge summit were used for agriculture in the 19th and early 20th centuries. Topographic slopes for much of the ridge are approximately symmetrical (see for example the topographic slope map, Plate I, in Morgan and others, 1997). However, in the area from Thornton Gap to approximately Fishers Gap, a distance of about 11 miles, the Stanley Fault has thrust granitic rocks west against clastic rocks and cut out all of the volcanic rocks along the west side of the ridge and from several areas (for example, Marys Rock) along the summit of the Ridge. Along this interval, the Ridge is markedly asymmetric, and topographic slopes along the west side have values typically between 30 and 45 degrees, whereas on the east side, slopes range from 14 to 26 degrees. The upper western slopes have prominent outcrops such as the Pinnacles and Franklin Cliffs.

The western flank of the Blue Ridge in this interval between Thornton Gap and Fishers Gap is extremely irregular and is drained by a series of streams flowing to the northwest and spaced along the ridge at one- to two-mile intervals. Each of the streams has eroded deeply into the granitic rocks to form a well defined drainage basin or hollow. The hollows are bounded by short interfluvial ridges parallel to the streams forming promontories or noses between the hollows or recesses. The complex of hollows and noses, coupled with the overall steep gradient of the western Blue Ridge, results in an extremely rugged, deeply dissected topography.

Shaver Hollow, drained by the North Fork of Dry Run, forms a typical drainage basin within this complex. It covers an area of 2 square miles within the Park and lies near the northwest corner of the Old Rag and the southwest corner of the Thornton Gap 7.5' topographic quadrangles. The southwest boundary of the hollow is the Crusher Ridge promontory, and the northeast promontory, beginning at the Pinnacle Picnic Grounds, is unnamed. The Pinnacles, at 3,600 feet altitude, is the highest point along the margin of the basin. About two thirds of the hollow is within the Park; the lower third, below an elevation of 1,680 feet, is on private property.

BEDROCK GEOLOGY

The entire area of Shaver Hollow is underlain by granitic rocks of Grenville (1.0-1.1 Ga) age, intruded by small dikes related to the Catoctin metavolcanic rocks of 570 ma (Badger and Sinha, 1988). These granitic rocks were first recognized by Jonas (1935), who identified them as hypersthene granodiorites. The presence of pyroxene (hypersthene) in a rock of granitic composition generally is indicative of ultra metamorphism in the granulite facies (Turner, 1968; Spear, 1993). Bloomer and Werner

(1955) gave the name of Pedlar to exposures of hypersthene granites along the Pedlar River southwest of the Shenandoah National Park and the name was applied by Gathright (1976) to the principal granitic rock unit within the Park. More recently, Rader and Evans (1993) abandoned the name Pedlar, and the new Geologic Map of Virginia designates these rocks as charnockite and charnockitic gneiss of Middle Proterozoic age. No systematic study has been made of the Virginia granitic rocks that have variously been termed Pedlar, hypersthene granodiorite, or charnockite. Recent studies of these rocks by Bartholomew and others (1991) and by Evans, 1991, do not discuss the granitic rocks within Shaver Hollow or within Shenandoah National Park.

Granitic rocks within Shaver Hollow are medium to coarse grained with about 5 percent dark minerals. Quartz and feldspar are easily recognized in hand specimen, but cleavages on feldspar are often absent suggesting extensive alteration. Dark minerals are mostly chlorite derived from alteration of pyroxene and hornblende. Gneissic layering is common in the rocks cropping out near the Skyline Drive, but in the lower part of the hollow, most rocks are nearly massive and somewhat finer grained. Fresh, unweathered rocks are scarce and are almost all from artificial cuts along the Skyline Drive or from outcrops at small waterfalls in Dry Run. These rocks are only slightly weathered and no saprolite was found anywhere within the study area.

The Catoctin metavolcanic rocks in the park were described by Reed (1955), and the chemistry and alteration were discussed by Reed and Morgan (1971). More recent work by Badger and Sinha (1988) and Badger (1993) established the age of the Catoctin at 570 ma and provided new information on the chemistry and origin of these rocks. The chemistry of these rocks is least altered in the dikes, is typical of tholeiite basalts with moderately high titanium content, and is similar to the chemistry of large, extensive plateau basalts. The Catoctin metavolcanic rocks are present in the hollow as small dikes making up probably less than 5 percent of the bedrock. They are easily recognized in talus and as loose blocks because of the fine grain and pale green color. An outcrop of a Catoctin dike is in the roadcut at mile 37.8 on the Skyline Drive at the head of Shaver Hollow. The dike is five feet thick striking N70W with a near vertical dip. It is identical in hand specimen to the loose blocks found everywhere within the hollow.

Both the hypersthene granite or charnockite and the Catoctin rocks were metamorphosed to the greenschist facies during the Allegheny orogeny near the end of the Paleozoic Era at about 270-300 ma. The high grade charnockite was retrograded with alteration of pyroxenes to chlorite, and feldspar to white micas; the volcanic rocks were altered to chlorite, albite, and epidote. The rocks appear to have moved as a coherent massif westward along the Stanley Fault with little rotation (the dikes are nearly all vertical), and major shear zones seen elsewhere (Bartholomew and others, 1981) and to the east of the park (Gathright, 1976, Plate II) were not seen in the area of Shaver Hollow.

SURFICIAL GEOLOGY

Study of the surficial deposits in Shaver Hollow was hampered by the absence of artificial excavations or recent scouring of the hollow by flood waters. However, abundant excavations from tree throw brought about by Hurricane Fran during September, 1996, revealed soil profiles and colluvium to depths of about 3 feet in many places. Heavy scour action after debris flows, so characteristic of the Madison County area affected by the storm of June, 1995 (Wieczoreck and others, 1996), did not occur during Fran despite rainfall amounts estimated at 7-9 inches during a 12 hour period.

The pattern of bedrock exposure in the hollow is an important guide to the origin and thickness of the surficial deposits (see figure 1). Outcrops are abundant near the summit of the Blue Ridge and along the interfluves bounding the hollow. Outcrops form bold cliffs immediately north and west of the Pinnacles Picnic Grounds, and prominent outcrops occur around the head of the hollow, especially along the Appalachian Trail. However, the principal outcrops in the hollow are not those on the perimeter, but in the thalweg near the center of the hollow at altitudes between 2,400 and 2,000 feet. In this area, the Dry Run flows on bedrock and drops over three waterfalls in a course of about 1,500 feet, each with a drop of about 15 - 20 feet.

Descriptions of surficial deposits in the Blue Ridge and speculations about their origin are aided by consideration of both current dynamic processes affecting the landscape and of changes in the conditions that govern the effectiveness of those processes. Until the collapse of the last ice sheet about 10,000 ybp, much of the higher elevation of the Blue Ridge was probably subjected to a severe climate unlike that of today. Clark (1992) documented many periglacial features in Pennsylvania, Maryland, and West Virginia, all south of the last ice sheet. These include sorted patterned ground, block fields, block slopes and block streams, frost mounds and ground ice scars, eolian sediments, and cryoplanation. Some of these features also exist further south in areas of higher elevation along the Blue Ridge in Virginia and North Carolina (Whittecar and Ryter, 1992).

In his study of the geomorphology of the Shenandoah Valley, Hack (1965) advocated an open system of dynamic equilibrium of all natural processes operating on the landscape. In his view, erosion rates are adjusted so that resistance of the rock, the slope being eroded, and the removal of material all eventually reach equilibrium. Climate change would affect the rate of these processes; if the climate remains constant long enough, the system will return to equilibrium, and evidence of an earlier climatic regime will be removed.

The principal tool cited by Hack for physical weathering in cold climate is the freezing and volume expansion of water. However, Ballantyne and Harris (1994) in their study of the periglaciation of Great Britain concluded that the volumetric expansion due to frost wedging has not been demonstrated anywhere. A more likely hypothesis is the "capillary theory of frost damage" developed by Walder and Hallet (1985, 1986), which is similar to the theory of ice wedges developed in soil under permafrost conditions. According to Walder and Hallet, capillary migration of water to ice in fractures is most likely to be effective within a temperature range of -4 to -15°C. Therefore, if climate conditions prevailed in upland areas of the Blue Ridge allowing the development of deep permafrost, the mechanical breakup of rocks in that region would have resulted from processes that are not operable today.

The study of such a limited area as Shaver Hollow cannot definitively prove that periglacial processes had a profound influence on the landscape, and that these processes produced an effect that has yet to be removed during a period of a more moderate climate. Although many of the surficial deposits at an elevation above 2,100 feet in Shaver Hollow probably formed under conditions of permafrost, a substantially larger area along the summit of the Blue Ridge should be investigated to obtain more conclusive evidence. Similar conclusions on the influence of climate on geomorphic processes were made by Whittecar and Ryter (1992) in a study of a small area along the Blue Ridge about 70 miles south of Shaver Hollow.

The major surficial deposits within Shaver Hollow are described below and each is depicted on Figure 1. Absolute ages for the origin or deposition of these deposits are not available, and in some instances even relative ages are uncertain.

Upland surficial deposits. Along the summit of the Blue Ridge, between Thornton Gap and Big Meadows, there are several, small areas of less than 1/4 square mile with nearly level, flat surfaces. These upland surfaces rarely have slopes greater than 6 degrees. Along the crest of the Blue Ridge near Shaver Hollow these surfaces are found at the Pinnacle Picnic Grounds, the Pinnacle Ranger Station, and Hughes River Gap in the northwest corner of the Old Rag Mountain, VA 7.5' topographic map. All of these surfaces have been extensively altered by human use, first by limited agriculture and then later by construction at CCC camps and construction of parking lots at scenic overlooks along the Skyline Drive. Outcrops are limited within these areas;

Dekay (1972) reported a depth to bedrock of 78 feet for a well drilled at the Pinnacle Ranger Station. Nevertheless, the upland surface areas are bounded to the west by prominent outcrops that form cliffs in a few areas (see Figure 1), especially at the Pinnacle Picnic Grounds and west of the Pinnacle Ranger Station), and along slopes that change abruptly from less than 6 degrees to more than 30 degrees.

There have been a number of published speculations on the origin of nearly level summit surfaces in the central Appalachians. These surfaces have been variously interpreted as being the remnant of a Cretaceous peneplain (Davis, 1890), inherited contact surfaces formed by the removal of overlying Catoctin greenstone (Reed, 1955; Gathright, 1976), or cryoplanation surfaces formed by solifluction during periglacial conditions (Clark, 1992). Davis' system of evolution of topography and the development of peneplains has generally been discredited, and detailed investigations of topography have not supported the concept of accordant summits defining an ancient, continuous, erosion surface. Thompson (1941) analyzed the topography of a large area in the southern Shenandoah Valley and adjacent areas and concluded that the altitudes of the ridges were too variable to support the concept of an ancient peneplain. Monmonier (1968) made a similar study in Pennsylvania using trend surface analysis and concluded that much of the elevation variation on flat upland surfaces can be explained by variations in structure and rock character.

The water well data for the Pinnacle Ranger Station indicate deep weathering of bedrock in at least some parts of the surface and suggests that processes that formed the surface have been active over a considerable (but indeterminate) time. A possible origin of the upland surfaces near Shaver Hollow, which would account for deep weathering, is the removal of a flat lying cap of Catoctin Greenstone by erosion from the underlying granitic rocks along the crest of the Blue Ridge. This is most easily envisioned by projecting the nearly flat contact of the Catoctin Formation at Little Stony Man (Mile 38.6, Skyline Drive; see Gathright, 1976, Plate II) to the northeast across the area shown in Figure 1 and just above the present crest of the Blue Ridge.

Evidence for periglacial conditions on these upland surfaces has been obscured by human activity. Much of the ground has been disturbed and direct evidence for periglacial conditions (such as solifluction terraces, patterned ground) was not observed. Scattered scree on upland surfaces of very gentle slope and without any visible outcrops of bedrock suggests that small blockfields did form on these surfaces during cold weather conditions.

Blocky Colluvium. The lower slopes of Shaver Hollow are covered by blocky colluvium, a diamicton with matrix supported clasts that rarely are more than 3 feet in length. Exposures of bedrock are relatively abundant, and individual outcrops in places form the head of small talus fans, but these do not coalesce into readily mappable units. The colluvium is widespread in this part of the Blue Ridge and forms the dominant unit, for example, in the uplands in the Madison Area just to the south of Shaver Hollow, studied by Wieczoreck and others (1996). Unlike the Madison area, the colluvium in Shaver Hollow is not exposed by deep gullies incised by recent storms. However, Hurricane Fran in September, 1996, toppled many trees, especially on Crusher Ridge, and the excavations provided by tree throw reveals profiles in this deposit up to about 3 feet deep. Clasts studied in these pits have weathering rinds averaging about 5 mm. Matrix within these pits makes up about 20 to 35 percent of the diamicton, and Munsel values for the soil matrix are 10YR 4-6/4 (yellow brown). Munsel values of soils and matrix are the same on both ridge crests and in deeper parts of the hollow and do not support evidence for a greater age for weathering on ridge crests.

Although periglacial processes may have been active in the production of this colluvium, evidence for widespread solifluction, resulting in downslope orientation of flat clasts, is not conspicuous. Other processes active in the present climate regime, including creep, in situ weathering, and mass wasting of soil, probably contribute to the development of these deposits.

Talus. The upper slopes of Shaver Hollow, from the edge of the upland surfaces previously described at about 3,200 feet down to an elevation of approximately 2,100 feet, is covered by an extensive talus deposit. The slopes covered by talus are steep, commonly in excess of 30 degrees, especially where outcrops of bedrock are exposed within the sheet. The talus is composed of blocks of granite gneiss and charnockite, and a small fraction of the blocks are derived from Catoctin dikes. The talus lies immediately below the bedrock outcrops that rim the northwest margin of the upland surfaces and extends downward in a complex of cone sheets and fans. The presence of slope outcrops of bedrock indicate that the talus sheet is not thick, and probably averages between ten and twenty feet thick. Well developed parts of the talus sheet may be seen along the Appalachian Trail northwest of the Pinnacle Ranger Station and on the trail from mile 37.9 on the Skyline Drive down to the Shaver Hollow shelter site and spring at an altitude of about 2,800 feet. The ratio of talus slope to the headwall producing the debris is very large, but caution should be made in using this to characterize the talus because of midslope outcrops which can replenish the supply of debris downhill; there is no unique point source for debris along a given line of slope.

Talus blocks typically are angular to moderately rounded, two to three feet in length, have edge/edge contacts, lack a soil matrix, do not exhibit weathering rinds more than about 1 mm thick (therefore, never had a soil matrix with extensive chemical exchange), contain many voids, and are unstable. Quantitative measurements of the talus sheet were not made, but qualitatively, larger blocks (greater than 6 feet in length) appear to be more abundant near the lower margin of the talus sheet.

The talus deposits tend to become broader below 3,000 feet; at 2,800 feet, the talus sheet is continuous across the entire hollow. Nevertheless, thick accumulations of talus in stream courses do not occur. There are no rock streams in any of the small drainages that feed into the North Fork of Dry Run. The floors of these small streams are armored with water laid cobbles. In these small stream beds, cobbles up to 1 foot in diameter were moved during Hurricane Fran in September, 1996. At the center of the hollow at about elevation 2,300 feet, the small streams converge to form the North Fork of Dry Run. It is also a point of convergence for the talus sheet in a hollow with contours sharply concave outward. However, the talus is entirely missing at the point of convergence, and the stream is flowing on bedrock as described previously.

The talus sheet does not extend below an altitude of approximately 2,100 feet as a continuous, coalescing deposit. The limit of the sheet is clearly defined between Crusher Ridge and the North Fork of Dry Run, but poorly defined around the nose of the unnamed ridge that extends up to the Pinnacle Picnic Grounds (see Plate I). At its lower terminus, the talus becomes thinner and is not marked by thickening in pressure ridges.

Debris Fans. Debris fans, as used in this report, represent the constructional land forms resulting from down slope deposition of prehistoric debris flows. A well developed, composite fan with evidence of multiple debris-flow events fills the floor of Shaver Hollow from an altitude of about 1,900 feet down stream beyond the study area toward Morning Star and coalesces with a similar fan which fills the floor of Tutwiler Hollow. Figure 1 shows the distribution of the debris fans and a relative sequence for their development is given in the paragraphs on interpretive history.

Most of the fan can be studied and interpreted only from its surface morphology. This is most apparent with the rapid decline of slope as well as the gradient of Dry Run, which drops from about 12 degrees near the head of the fan at 1,900 feet to about 4 degrees at the cabin of Darwin Lambert at an altitude of 1,600 feet. The rubbly matrix-supported ground is characteristic of the fan as are the irregular contour lines and poorly defined stream channels, which help to delineate the deposit on the topographic map. Early settlers built massive stone walls as much as 10 feet thick in an effort to clear lower parts of the debris-flow deposits for agricultural use, and these are still visible on areal photographs.

In one locality, the North Fork of Dry Run has cut into the side of a debris-flow deposit about 500 feet south of the Park boundary. During the flood that accompanied Hurricane Fran, the fabric in the deposit was exposed along the bank of the stream. The cut is about 10 feet high and reveals only one debris-flow deposit. Cobbles average 1 foot in length but generally range from about 0.5 to 1.5 feet with rare boulders as much as 3 feet in length. The clasts are partially supported with a sand matrix with virtually no silt or clay. The boulders are loosely held by the matrix and weathering rinds are about 2 mm.

In size, morphology, thickness, and general aspect, the debris-flow deposits in Shaver Hollow resemble the deposits exposed in Madison County during the June, 1995 storm (Wieczoreck and others, 1996; Morgan and others, 1997). The lack of cementation in the matrix, the lack of reddening in the soil matrix, and the thinness of the weathering rinds suggests that the debris flows are of Holocene or perhaps latest Pleistocene age, but the absence of carbon 14 dates makes that estimate conjectural.

Alluvial deposits. With the decrease in gradient, the North Fork of Dry Run flows over a narrow alluviated area beginning at an altitude of about 1,700 feet. The floor of the stream is armored with imbricated cobbles and boulders. The deposit is thin and discontinuous within the Park; an outcrop of bedrock occurs just north of the Park boundary and is the site of stream sampling by researchers at the University of Virginia. Evidence of frequent flooding is conspicuous with boulders perched against trees and numerous minor changes in the stream channel. During the flood that accompanied Hurricane Fran, boulders up to 3 feet in length were moved within the alluviated area. Downstream, roads, bridges, houses, and a dam were heavily damaged during this storm.

INTERPRETIVE HISTORY

An interpretive history of the surficial deposits in Shaver Hollow can be made in terms of relative ages of materials.

The upland surface probably has a composite origin partly related to stripping of the protective cap of Catoctin Formation from the mountain ridge. The stripping was perhaps intensified by the structurally elevated position of the basement and cover, a result of Paleozoic movement along the Stanley Fault. If so, a cap of Catoctin Formation vanished completely before the present sequence of deposits was accumulated in Shaver Hollow. Catoctin metavolcanic rocks are nearly absent from the extensive talus deposits in the hollow except for a few boulders probably derived from dikes. The upland surface was probably exposed during the late Pleistocene and modified by permafrost action which included solifluction and planation. If so, virtually the only evidence that remains is the morphology of the surface itself, with low slope values and very minor block fields.

Hack (1965) in his study of the geomorphology of the Shenandoah Valley made a study of scree developed on Chillhowie quartzites and siltstones in the southern part of the Shenandoah National Park. However, he attributed bouldery deposits of granites to weathering, removal of soil by mass wasting, and the subsequent collection at the surface of core stones which armor the hillsides and valleys as forest covered scree and stone streams. The extensive talus deposits in Shaver Hollow cannot be attributed to lag deposits; weathering rinds are almost completely missing, the boulders are angular, and in most cases, the talus can be traced upslope to their source at outcrops of granitic rock. The large volume of talus material relative to the vestiges of outcrops that produced the talus suggests that cold weather conditions greatly accelerated the production of rubble and that the process had almost completly destroyed these outcrops by the end of the Pleistocene. Although Hack (1965) specifically used a model of talus production and removal to demonstrate his concept of dynamic equilibrium, it is difficult to envision dynamic production and removal of 3-foot-diameter boulders from very minor outcrops in the present forest-covered, lichen-rich talus sheet. Rather than being derived from the substrate as core stones, the talus lies stratigraphically over the blocky colluvium, was derived from disruption of outcrops, but was not produced or deposited in any quantity below an elevation of about 2,100 feet.

The absence of talus in the center of Shaver Hollow and the presence of an extensive composite debris-flow fan downstream suggests that the talus sheet from the center of the hollow provided the debris for the downstream fans. If so, a succession of debris-flow events, perhaps at the end of the Pleistocene or in early Holocene time, stripped most of the debris out of the center of the hollow, leaving the stream to flow directly on bedrock. Similar relationships are commonly observed in the debris-flow fans and upstream channels produced by the June, 1995 storm in Madison County (Wieczoreck and others, 1996). The oldest preserved debris-flow fan in Shaver Hollow tilts about 3 degrees away from the present drainage toward an earlier drainage. That older stream drained just west of the site of the original Shaver cabin now occupied by Mr. Darwin Lambert. The lower part of the present fan was produced by a massive debris flow, which removed parts of the earlier fan just east of the Lambert cabin and deposited materials downstream. The uppermost part of the Shaver Hollow fan consists of minor, multiple headwall and side-slope, debris-flow and landslide contributions into the area cleaned and enlarged by the previous, large debris flow. Some of these minor fans may have formed from recent events.

NOTE ON HYDROLOGIC SAMPLING

The investigation of the surficial geology in Shaver Hollow was undertaken as a result of a suggestion by scientists at the University of Virginia to provide the basis for further research into water/rock interactions. This information can be used to discriminate between rock and atmospheric effects on water chemistry.

The upper part of the hollow should be sampled for water chemistry at the Shaver Hollow shelter/spring site. This site should be ideal also for dating the water using chlorofluorocarbon techniques (Busenberg and Plummer 1992). A second upper hollow site should be near the waterfalls at the conjunction of minor streams to form the North Fork of Dry Run. Both of these sites will sample waters which are probably young (Plummer, personal communication, 1996) and which have primarily been in contact with bedrock and talus armored with only 1-2 mm of weathering rinds.

Further downstream, within the composite debris-flow fans, residence times of water may be much longer with prolonged exposure to more weathered material. Estimates of the quantity of water that is spring fed and recycled into the stream should be made to calculate the influence of the fan on water chemistry. Springs just west of the Lambert residence should be sampled for water chemistry and age. Both of these data should be known before chemistry of surface water from the collection site near the park boundary can be cast into a more complete model.

REFERENCES

Badger, R.L., 1993, Fluid interaction and geochemical mobility in metabasalts: an example from the Central Appalachians: Journal of Geology, v. 101, p. 85-95.

Badger, R.L., and Sinha, A.K., 1988, Age and Sr isotopic signature of the Catoctin Volcanic Province: Implications for subcrustal mantle evolution: Geology, v. 16, p. 692-695.

Ballantyne, C.K., and Harris, Charles, 1994, The Periglaciation of Great Britain: Cambridge University Press, 330 pp.

Bartholomew, M.J., Gathright, T.M., and Henika, W.S., 1981, A Tectonic model for the Blue Ridge in Central Virginia: American Journal of Science, v. 281, p. 1164-1183.

Bartholomew, M.J., Lewis, S.E., Hughes, S.S., Badger, R.L., and Sinha, A.K., 1991, Tectonic History of the Blue Ridge Basement and its cover, central Virginia. p. 57-90 in Geologic Evolution of the Eastern United States, Schultz,A., and Compton-Gooding, E,eds.: Virginia Museum of Natural History, Martinsville, Va., Guidebook 2, 304 p.

Bloomer, R.O., and Werner, H.J., 1955, Geology of the Blue Ridge region in central Virginia: Geological Society of America Bulletin, v. 66, p. 579-606.

Busenberg, E., and Plummer, L.N., 1992, Use of chlorofluorocarbons (CCl_3F and CCl_2F_2) as hydrologic traces and age-dating tools: Example - the alluvium and terrace system of central Oklahoma: Water Resources Research, v. 28, p. 2257-2283.

Clark, G.M., 1992, Central Appalachian periglacial geomorphology: a field excursion guidebook: The Pennsylvania State University, Agronomy Series Number 120, 248 p.

Davis, W.M., 1890, The geologic dates of origin of certain topographic forms on the Atlantic slope of the United States: Geological Society of America Bulletin, v. 2, p. 545- 581.

Dekay, R.H., 1972, Development of ground-water supplies in Shenandoah National Park, Virginia: Virginia Division of Mineral Resources, Mineral Resources Report 10, 158 p.

Evans, N.H., 1991, Latest Precambrian to Ordovician metamorphism in the Virginia Blue Ridge: origin of the contrasting Lovingston and Pedlar basement terranes: American Journal of Science, v. 291, p. 425-452.

Gathright, T.M., 1976, Geology of the Shenandoah National Park, Virginia: Virginia Division of Mineral Resources, Bulletin 86, 93 p.

Hack, J.T., 1965, Geomorphology of the Shenandoah Valley, Virginia and West Virginia and origin of the residual ore deposits: U.S. Geological Survey Profesional Paper 484, 84 pp.

Jonas, A.I., 1935, Hypersthene granodiorite in Virginia: Geological Society of America Bulletin, v. 46, p. 47-60.

Monmonier, M.S., 1968, Trends in upland accordance in Pennsylvania"s Ridge and Valley section: Proceedings of the Pennsylvania Academy of Science, v. 42, p. 157-162.

Morgan, B.A., Wieczoreck, G.F., Campbell, R.H., and Gori, P.L., 1997, Debris-flow hazards in areas affected by the June 27, 1995, storm in Madison County, Virginia: U. S. Geological Survey, Open File Report 97-438, 15 p.

Rader, E.K., and Evans, N.H., eds., 1993, Geologic map of Virginia - expanded explanation: Virginia Division of Mineral Resources, 80 p.

Reed, J.C., Jr., 1955, Catoctin Formation near Luray, Virginia: Geological Society of America Bulletin, v. 66, p. 871-896.

Reed, J.C., and Morgan, B.A., 1971, Chemical alteration and spilitization of the Catoctin greenstones, Shenandoah National Park, Virginia: Journal of Geology, v. 79, p. 526-548.

Spear, F.S., 1993, Metamorphic phase equilibria and pressure -temperature-time paths: Mineralogical Society of America Monograph, 799 p.

Thompson, H.D., 1941, Topographic analysis of the Monterey, Staunton, and Harrisonburg Quadrangles: Journal of Geology, v. 49, p. 521-549.

Turner, F.J., 1968, Metamorphic Petrology: McGraw-Hill Book Company, New York, 403 p.

Walder, J.S., and Hallet, B., 1985, A theoretical model of the fracture of rock during freezing: Geological Society of America Bulletin, v. 96, p. 336-346.

Walder, J.S., and Hallet, B., 1986, The physical basis of frost weathering: toward a more fundamental and unified perspective: Arctic and Alpine Research, v. 18, p. 27-32.

Whittecar, G.R., and Ryter, D.W., 1992, Boulder streams, debris fans, and Pleistocene climate change in the Blue Ridge Mountains of Central Virginia: Journal of Geology, v. 100, p. 487-494.

Wieczorek, G.F., Morgan, B.A., Campbell, R.H., Orndorff, R.C., Burton, W.C., Southworth, C.S., and Smith, J.A., 1996, Preliminary inventory of landslide, debris-flow, and flooding effects of the June 27, 1995, Storm in Madison County Virginia: U. S. Geological Survey Open File Report 96-13, 8 p.