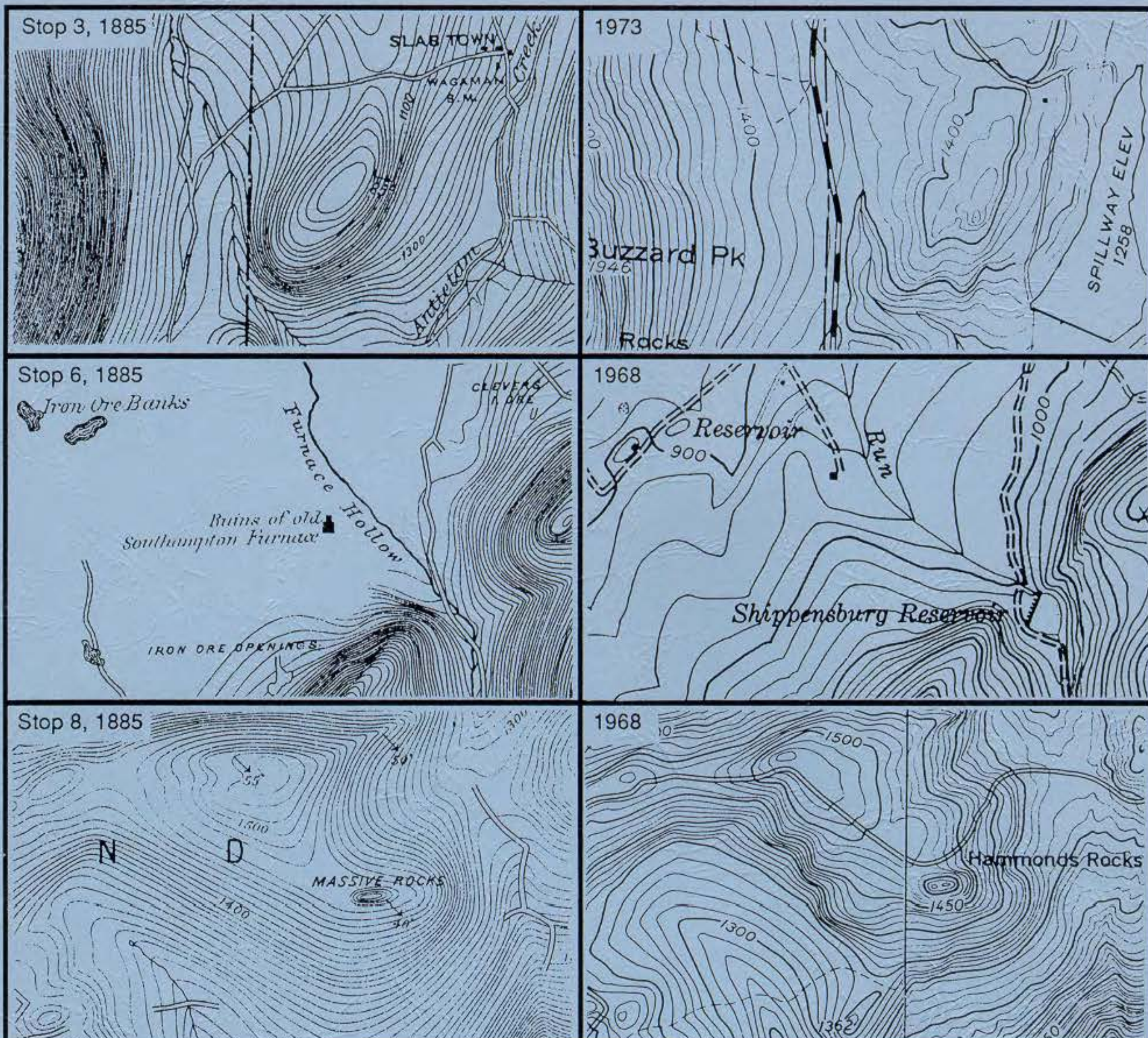
Guidebook

56th Annual Field Conference of Pennsylvania Geologists

Geology in the South Mountain Area, Pennsylvania



Hosts:
Dickinson College
Pennsylvania Geological Survey

September 26, 27, and 28, 1991
Carlisle, Pa.

Guidebook for the

56th ANNUAL FIELD CONFERENCE OF PENNSYLVANIA GEOLOGISTS

GEOLOGY IN THE SOUTH MOUNTAIN AREA, PENNSYLVANIA

W. D. Sevon and Noel Potter, Jr., editors
Contributors and field trip leaders:

Rolf, Ackermann
John H. Barnes
Albert E. Becher
S. W. Berkheiser, Jr.
Douglas C. Chichester
G. Michael Clark
Gretchen Dockter
Henry W. A. Hanson
Marcus M. Key, Jr.
Steven Lev
Laura Pezzoli
Robert J Schott
Samuel J. Sims
Samuel I. Root
R. C. Smith, II
Thomas Troy
Randal L. Van Scyoc
John H. Way
Candie C. Wilderman
Ellis L. Yochelson

September 26-28, 1991

Hosts: Dickinson College
Pennsylvania Geological Survey

Headquarters Motel: The Embers Convention Center, Carlisle, PA

Cover: Topographic maps at 1 inch = 1,600 feet for 3 stops on the field trip. Older maps are from a series with a contour interval of 10 feet that cover the entire South Mountain area in Pennsylvania. They were surveyed and drafted by Ambrose E. Lehman 1875-1885, and appear in Atlases D5 and D6 of the Second Pennsylvania Geological Survey. Modern maps are:
Stop 3, USGS Iron Springs 7.5-minute quadrangle, C.I. = 20 feet.
Stop 6, USGS Scotland and Caledonia Park 7.5-minute quadrangle, C.I. = 20 feet.
Stop 8, USGS Dickinson and Mount Holly Springs 7.5-minute quadrangle, C.I. = 10 feet.

Guidebooks distributed by:

Field Conference of Pennsylvania Geologists, Inc.
P. O. Box 1124
Harrisburg, PA 17108

TABLE OF CONTENTS

LIST OF FIGURES

LIST OF PLATES

LIST OF TABLES

56TH ANNUAL FIELD CONFERENCE OF PENNSYLVANIA GEOLOGISTS

GEOLOGY IN THE SOUTH MOUNTAIN AREA, PENNSYLVANIA

INTRODUCTION

by
Noel Potter, Jr.
Department of Geology, Dickinson College, Carlisle, PA

Welcome to Carlisle and the 56th Annual Field Conference of Pennsylvania Geologists. In recent years the Field Conference has either focused on a particular problem in a given area or it has looked at the general geology of an area. The 1991 meeting follows the latter pattern with a look at an eclectic range of problems in the South Mountain area south and west of Carlisle. We think we have something for everyone, from hard-rock to soft-rock to no-rock.

Past Field Conferences to South Mountain

This is not the first time that the Field Conference visited South Mountain. In May, 1933, the 3rd Annual Field Conference, based in Harrisburg, had 5 trips, one of which visited South Mountain (Stone and Hickok, 1933). In those days, the guidebooks were shorter and a good bit slimmer than today's. There was a road log, but narrative was terse (e.g., "miles 6.6-7.6 Harrisburg peneplane") and stop descriptions gave only the rock type with no discussion of the geology. I suppose that this gave the leaders the opportunity to say what they wanted when they got there. They didn't have much time to expound or argue--stop descriptions give the time of arrival and departure, typically 5 or 10 minutes per stop. There were 14 stops, not counting lunch. Led by R. W. Stone and W. O. Hickok, IV, the 3rd Conference group visited a variety of outcrops as far south as Waynesboro in the Great Valley and on South Mountain. Their first stop to examine the Mesozoic diabase dike that crosses the Cumberland Valley was on US Route 11 just 0.4 miles east of this year's headquarters at the Embers Inn. They briefly stopped at Pine Grove Furnace, which we will visit on the second day (Stop 10).

The 14th Field Conference in 1948, again based in Harrisburg, had 4 trips, one of which was devoted to South Mountain. That trip was led by Pete Foose, R. C. Stephenson, and F. M. Swartz (Foose and others, 1948). They went through Boiling Springs, Mount Holly Springs, made a stop in Mount Holly Gap to examine the Antietam quartzite (see Yochelson, this guidebook) and several stops in the valley of Mountain Creek near Laurel Lake and Pine Grove Furnace. They then went south and stopped near this year's Stop 9, and to Bendersville, Idaville, and York Springs before returning to Harrisburg via Dillsburg. Again, there is no narrative for the stops.

The 23rd Field Conference dealt with the structural geology of South Mountain in Maryland and was led by Ernst Cloos (1958). The trip did not enter Pennsylvania. By this time there were full written discussions for each stop.

The 31st Field Conference, led by Dave MacLachlan and Sam Root (1966), compared the tectonics and stratigraphy of the Cumberland Valley west of Harrisburg and the Lebanon Valley east of Harrisburg. Although the trip route passed along the valleys of Mountain and Conococheague Creeks within South Mountain, all of the Cumberland Valley stops were in the Cambro-Ordovician rocks south of Chambersburg.

It has been nearly a quarter century since the 33rd Field Conference last visited South Mountain in 1968 (Cloos and others, 1968). The theme that year was "the geology of mineral deposits in south-central Pennsylvania," and 3 of the stops were within South Mountain. There were stops at (1) a quarry in metabasalt used for roofing granules near Charmian, southwest of Gettysburg, (2) the former Benders Quarry southwest of Mount Holly Springs in the Antietam quartzite, and (3) the Philadelphia Clay Products Co. pit in clays weathered from the Tomstown dolomite at Toland on the south side of South Mountain southwest of Mount Holly Springs. This pit is now owned by Hempt Brothers and used for crushed quartzite. Our last stop (Stop 11) this year will be in the former Benders Quarry (now owned by Pennsy Supply). The pit has expanded considerably from the "mom and pop" operation of 23 years ago.

This Year--An Overview

A series of papers in the first half of the guidebook provide summaries of most aspects of the geology of South Mountain, and a few details that add to the story.

The metavolcanics have long been a subject of interest in the South Mountain area. Indeed, Florence Bascom's 1893 paper on the "acid volcanic rocks" of the area was published in volume 1 of the Journal of Geology. Until recently, the metavolcanics in Pennsylvania have not received as much attention as further south in the Blue Ridge. Smith, Berkheiser, and Barnes discuss the Catoctin metabasalts of the Pennsylvania part of South Mountain and contrast them to those further south. They focus on South Mountain, but also discuss and present geochemical data on metabasalts from elsewhere in southeastern Pennsylvania. They suggest possible implications for distinguishing Precambrian from Paleozoic metabasalts in the Piedmont.

The stratigraphy and sedimentology of the Lower Cambrian Chilhowee Group (Loudoun, Weverton, Harpers, and Antietam Formations) is reviewed by Key. Historically the age of the Chilhowee has been a problem. It is now well established lowermost Cambrian, between radiometric ages on the underlying Catoctin Formation and biostratigraphic ages in the Weverton and Antietam Formations. In an interesting historic note, Yochelson tells how in 1892, while on vacation with his family in Mount Holly Springs, C. D. Walcott found trilobite fragments that established the age of the Antietam in Pennsylvania as Early Cambrian.

Almost any geology student in the country will have at least briefly encountered South Mountain in structural geology when they heard or read of Ernst Cloos' classic studies of stretched oolites (actually in the Great Valley to the northwest) and other indicators of strain. Root and Smith present an overview of the structural geology and history of South Mountain. If you are new to the South Mountain area, you could do no better for an introduction than to read the first few pages of their paper. That will set the scene for the whole trip. Additional comments about the relationship of South Mountain to regional structural geology offered by MacLachlan are food of additional thought.

Surprisingly little has been done on the geomorphology of South Mountain. It deserves better treatment. Clark reviews what is known of the geomorphic evolution of the South Mountain area. There are features on many mountain crests that may be periglacial relics from Pleistocene cold climates, and the alluvial and colluvial diamicts that flank the mountain to the north and west beg further study.

Chichester discusses work being done on modeling what happens to some of the abundant water which leaves South Mountain and enters the Great Valley by way of the apron of gravels that flank the mountain. The quartz-rich rocks of South Mountain do not buffer acid precipitation that is so much in the news these past few years. Most of us are aware of the problem in general, but tend not to think of it as site-specific, depending upon the the geology of a given area. Wilderman reviews what is known of the effects of acid deposition at two sites on South Mountain.

Finally, the South Mountain area has had a rich history of yielding mineral resources, though not in the flashy way of giant mines. Berkheiser and Smith review what has and is being mined from South Mountain. The range of products is remarkable for a relatively small area. Van Scyoc reviews the details of production of sand and crushed aggregate from two sites that we will visit on the trip. Way traces the history of iron making that goes back to Colonial times as background for our stop at Pine Grove Furnace State Park on the second day.

The Trip

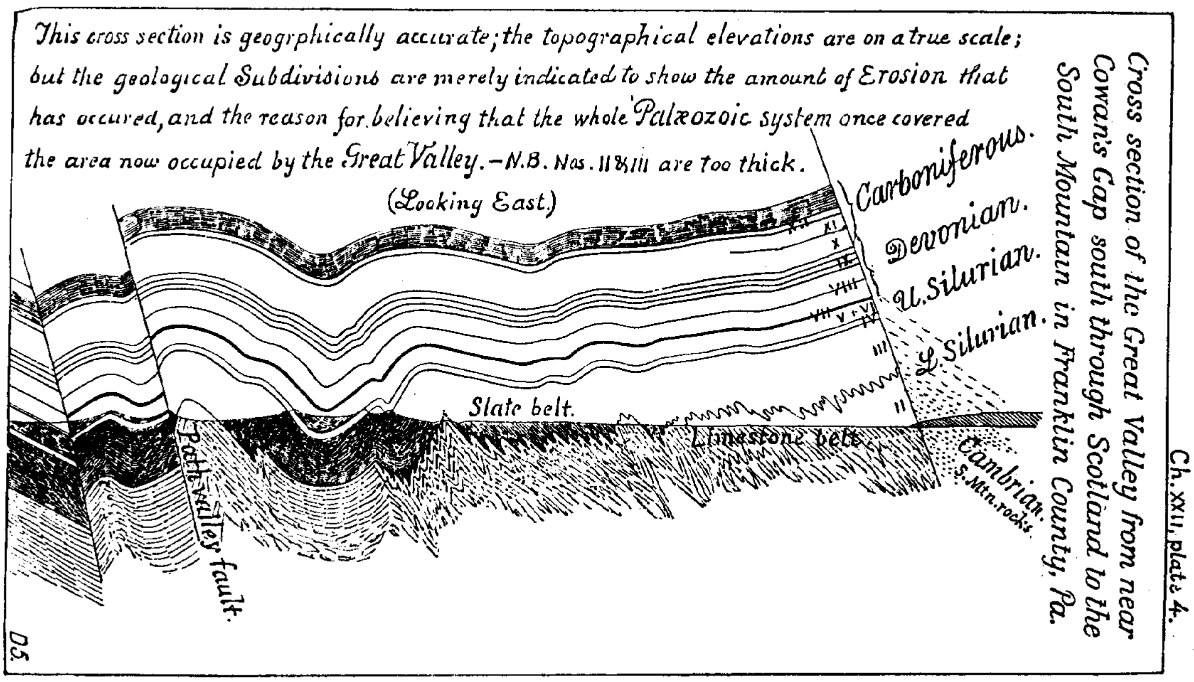
The first day of the trip is devoted to the portion of South Mountain from just south of the Maryland line to just north of Caledonia. At the stops we will examine: (1) a blockfield of presumed periglacial origin, (2) a tor and flat upland surface of possible periglacial origin, (3) metavolcanics, (4) metavolcanics and the Loudoun Formation, (5) a quarry in the Antietam quartzite with some spectacular megaripples and some colluvium, and (6) a 40-foot deep pit near the head of the apron of colluvial and alluvial gravels that are so ubiquitous on the northwest flank of South Mountain.

The second day will be devoted to the area from just south of Carlisle to the vicinity of Pine Grove Furnace. At the stops we will examine: (7) the hydrogeology of water that has its origin on South Mountain and supplies large springs such as those at Boiling Springs in the Great Valley, (8) Hammond's Rocks, an exposure of interbedded Weverton sandstone and conglomerate that has a variety of sedimentary structures, deformed pebbles that yield strain information, and some geomorphology, (9) an exposure of the Loudoun Formation, (10) the history of mining at Pine Grove Furnace State Park, and (11) a large quarry in the Antietam quartzite near Mount Holly Springs.

A listing of published geological information pertaining to the area is presented on page 143. The names of the 7.5-minute topographic quadrangles for the area are given on page 144. The field trip route map is on page 145.

Acknowledgements

A larger cast of characters is involved in a Field Conference than appears on the title page. I have appreciated the alacrity with which all of the authors and stop leaders responded to our call for help. Colleagues Bill Vernon, Henry Hanson, Jeff Niemitz, and Marcus Key at Dickinson College shared freely their knowledge of the South Mountain area and were helpful in selecting several stops. Mark Kauffman, our Department Technician, and several students will serve as road guards and bring you refresh-refreshments to keep you happy. We appreciate the hospitality of Valley Quarries, Inc. and Pennsy Supply, Inc., who have been generous with their hospitality both as we prepared stop descriptions and in letting Field Conference participants visit their operations. Bill Sevon has been a full partner in the preparation of this guidebook, and I have appreciated his hard work, wise counsel, and editorial skills. He wields the sharpest editorial pencil that I know in Pennsylvania.



from Lesley, 1892, p. 281.

PENNSYLVANIA'S VERSION OF THE CATOCTIN METABASALT STORY

by
R. C. Smith, II, S. W. Berkheiser, Jr., and J. H. Barnes
Pennsylvania Geological Survey, Harrisburg, PA

INTRODUCTION

An as-yet incomplete reconnaissance study of rocks of basaltic composition in southeastern Pennsylvania begun in late 1989, quickly revealed several chemically distinct groupings. Data for the elements that are relatively incompatible (i.e., that do not fit in major rock-forming minerals) and immobile (resistant to deuteric, hydrothermal, metamorphic, and weathering effects) were interpreted using published diagrams of analyzed basalts from known tectono-magmatic environments. Many samples fell into the environments that would be expected from other evidence. Examples of these include the Jonestown Volcanic Suite, the James Run volcanics, the Wilmington Complex, the Sams Creek Metabasalt, and basalts in the Sykesville Formation. A few others, such as the "Older Diabase" of the West Chester 15-minute quadrangle and samples from the Reading Prong Hexenkopf Complex, yielded somewhat unexpected results. However, when tectono-magmatic interpretation diagrams were used for samples from the three Mesozoic diabase type localities, derivation from an incompatible element-depleted mantle like that associated with island arcs was suggested. This is in direct conflict with the known continental rift association of the Mesozoic basins and will be discussed later.

A large population of within plate basalts was identified that includes metabasalt flows from the South Mountain anticlinorium, the Pigeon Hills, and Accomac; metadiabase dikes in the Reading Prong and Womelsdorf outlier, the Honeybrook Upland, and the Trenton Prong, as well as probable flows near the Avondale Prong and possibly the Holtwood area. As these outlying basaltic rocks are not readily distinguished from those in the South Mountain anticlinorium on a geochemical basis, it now appears that the Catoctin basalt magma was not restricted to South Mountain. Indeed, the Catoctin event appears to be the most underrated geologic event in the Appalachians. This interpretation appears to be further enhanced by the fact that the youngest Catoctin and oldest Mesozoic basaltic rocks appear, in some ways, to be complimentary geochemical fractions.

PRIMARY IGNEOUS FEATURES

Greenschist grade metamorphism and Alleghanian plus possibly Mesozoic deformation have generally been credited with wholesale destruction of igneous textures and features in metabasalts throughout much of the crystalline core of the South Mountain anticlinorium of Pennsylvania. However, irregularly shaped but recognizable 2 to 5 mm amygdules filled with epidote, chlorite, quartz, and albite are conspicuous in metabasalts in many areas. Less obvious, but common in sheared metabasalts are extremely flattened, thin amygdules, almost universally filled with dark green chlorite. Both types of amygdule occurrences appear to have a large geographic and stratigraphic range.

Less commonly recognized, but present in both the Snowy Mountain (as at Stop 4) and Green Ridge belts (belt terminology of Fauth, 1978, Figure 2), are relatively undeformed pyroclastic-bearing flows. Megascopic pyroclast fragments (Plate I) include black, highly vesicular lapilli <1 cm in diameter; isolated vesicular-cored, sub-rounded bombs >10 cm in diameter, with either crystalline or previously glassy rims; masses of welded cow-pie bombs and spatter (agglomerate-agglutinate) with greenish, highly vesicular cores; and angular, epidotized blocks (autoliths?) that can be >10 cm in diameter. Of the megascopic pyroclastic phases, most appear to be in the upper 50 m of preserved section and typically bear ~5 cm bombs on ~30 cm centers. The less common, agglomerate-agglutinate phase appears to occur within these pyroclastic units.

Rarely preserved in the Snowy Mountain and Green Ridge belts are pahoehoe toes. At a locality on Culp Ridge (Plate II), they have well-developed undulatory to bulbous tops with 3 to 4 cm-thick upper chilled zones (Table 1, sample CRC). These toes tend to be about 0.3 m thick and up to nearly 2 m long. Except for the upper, chilled margin, they are epidotized. At several localities there are less well-preserved epidotized toes(?) with rounded tops and flattened bases that contain sparse spherical amygdules. In the Jacks Mountain belt, possible toes occur along Friends Creek in a highly vesicular metabasalt portion of the section (Table 1, sample FRNDCK).

Possibly related, typically epidotized bodies with elliptical cross sections that range from 10 cm to >10 m (largest at the Pine Mountain quarry) appear to be pahoehoe flow feeder tubes. None, however, contains sufficient internal detail to verify this. Neither toes nor possible feeder tubes have been observed occurring with any sort of breccia, reducing the likelihood that they are misidentified pillows.

Ropey pahoehoe was observed in bedding plane plan view at only one definitive locality in the Green Ridge belt. Three possible examples exist in the Snowy Mountain belt (Plate III). In both areas, the pahoehoe is in overturned flows and the long

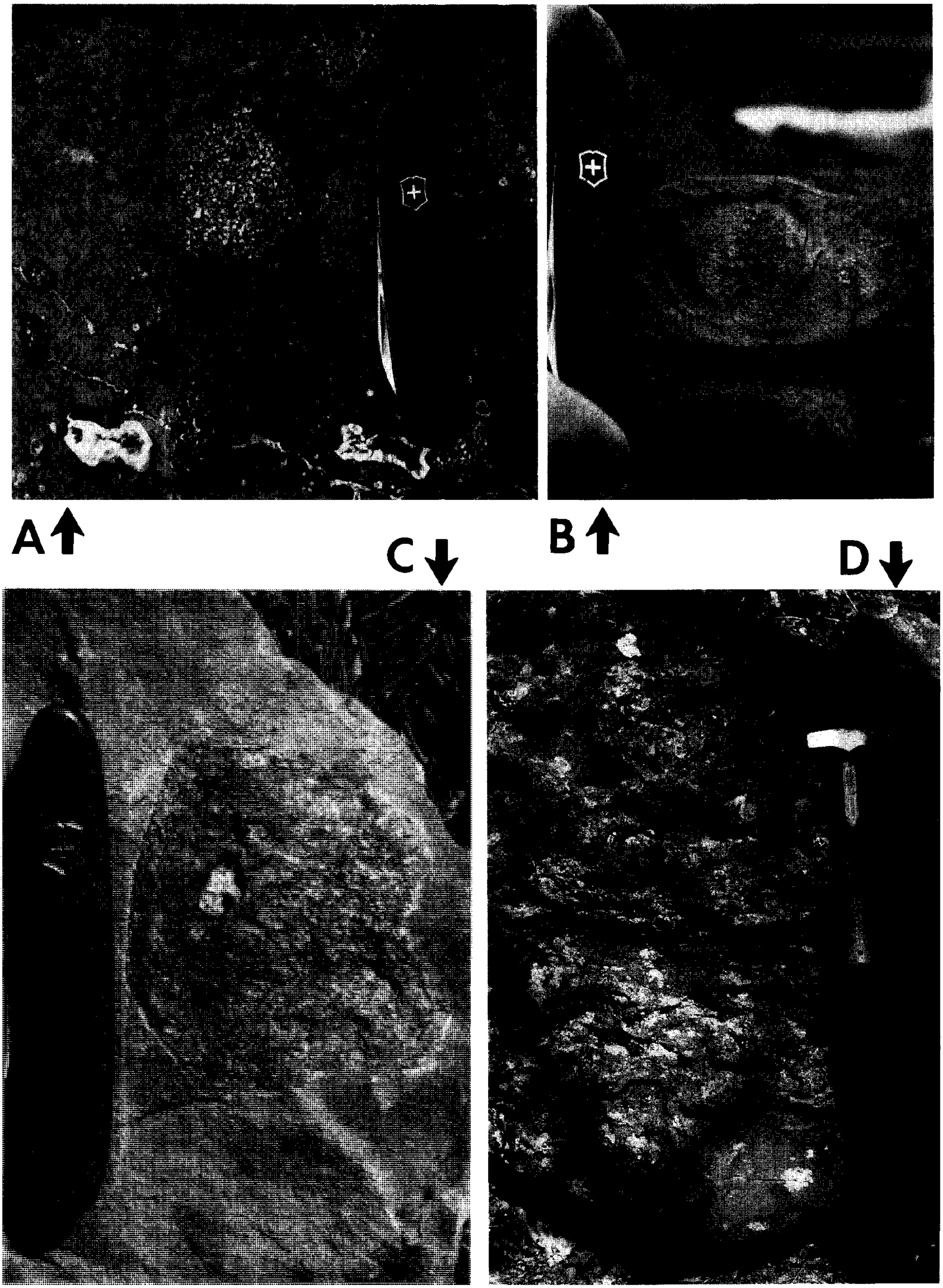
Plate I. Opposite page.

A. Albite-filled amygdules in a pyroclastic bomb in metabasalt containing black, microvesicular lapilli; 0.2 to 2 cm spherical and irregular amygdules with albite rims and epidote or chloritic cores. Float block A, Stop 4, east of Snowy Mountain.

B. Cross section through pyroclast in metabasalt. The bomb has a vesicular core and non-vesicular, now epidotized rim. Southwest of Mt. Hope.

C. Vesicular pyroclast bomb in massive metabasalt. Stop 4, east of Snowy Mountain.

D. Cross section through an agglomerate-agglutinate pile (welded cow-pie bombs and spatter) within a bomb-rich pyroclastic unit. Southwest of Mt. Hope.


A
B
C
D

axis of the ropes is variable but typically plunges to the southwest at about 10 to 20°. The concentric rope pattern suggests that locally the flow direction was from the southeast. At the definitive ropey pahoehoe locality in the Green Ridge belt south of Mount Hope, pipe vesicles, typically with elliptical, 4 x 8 mm cross-sections and lengths up to about 8 cm, are also present (Plates III and IV). Occurring in intimate association with ropy pahoehoe, they confirm So, and that the sequence is overturned (they occur in the base of a bed), and they verify subaerial eruption onto a wet surface. The median bedding defined by the possibly slumped ropey pahoehoe surface is N37°E, overturned 59°SE, that defined by the base of the pipe vesicles in a nearby outcrop is N24°E, overturned 73°SE; and that defined by a plane normal to the long axis of the pipe vesicles N27°E, overturned 67°SE. The pipe vesicles curve downward toward the southeast near their bases. This appears to be due to rotation into the predominant regional cleavage which here trends N28°E, 54°SE, rather than magma flow generation of pipe bowls (Bowls appear to have been largely removed by erosion, but vestiges suggest flow from the southeast). Because the dip of cleavage is less steep than bedding and is in the same quadrant, overturning is confirmed.

GEOCHEMICAL DATA

After locating an outcrop, sampling for geochemical analysis typically consisted of using a sledge hammer and chisel to remove a few kg block of metabasalt. Slab selection was based on identifying blocks containing the least fractured or mineralized areas. Although field dressed to remove weathered surfaces, mineral-filled veinlets, and mineralized fractures, further trimming as needed was done in the lab using a hydraulic splitter and an electric grinder. Sample preparation consisted of jaw crushing, pulverizing, hand-grinding (-200 mesh), sample homogenization, and sample splitting. Commercial analyses were obtained for a suite of major, minor, and trace elements.

Based on previous experience with basaltic rocks, TiO_2 was chosen as being the simplest and most useful parameter for geochemical grouping. Accordingly, Table 1, which presents data for 10 of the most useful, relatively incompatible and immobile elements, is organized by decreasing order of TiO_2 content. The

Plate II. Opposite page.

A. Cross section through a 14 x 46 cm epidotized body with 1 to 5 mm vesicles and having a flat bottom. It is interpreted to be a pahoehoe toe or large bomb and occurs in a pyroclastic unit. Southwest of Mt. Hope.

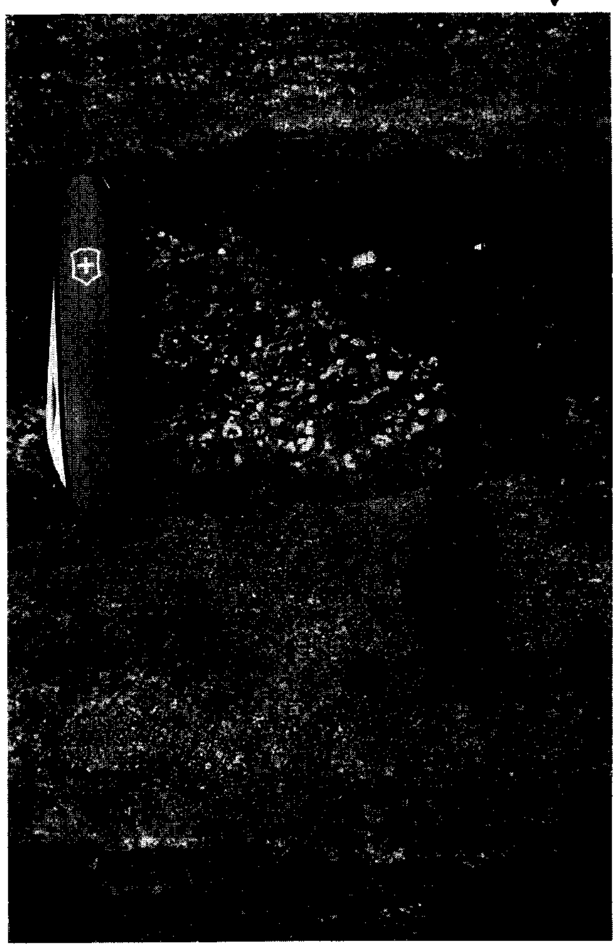
B. Looking down on pahoehoe toes that now protrude through the remnant of the overlying basalt flow, Culp Ridge.

C. Cross section through thin basalt flows. The amygdules in the darker layers, such as the one by the pocket knife, contain more epidote than the lighter flows with smaller amygdules. The predominant mineral filling in both is albite. Stop 3, west of Waynesboro Reservoir.

A↑

B↓

C↓



results of analyses of samples from the South Mountain anticlinorium proper, the Pigeon Hills, Accomac, and chemically similar dikes and possible flows in Grenvillian terrain and portions of the Piedmont are included. For purposes of discussion, the data have been divided into five geochemical groups, each of which may represent a new pulse of magma into the crust. This simplistic division is done on the basis of TiO_2 frequency plot distributions. Whether these groupings can stand the test of additional data is unknown, but in view of the variety of textures sampled (for lack of uniformity from one outcrop to the next), it is remarkable that even such apparent differences exist.

Despite the fact that each group contains samples from not only the South Mountain anticlinorium but also environs beyond, there don't appear to be any obvious chemical differences on the basis of location. This could be interpreted to suggest that each pulse of Catoctin basaltic magma was present over a wide geographic area. In fact, with the exception of Mine Ridge, all of the Precambrian terrains in Pennsylvania appear to have been affected. This near ubiquity appears to provide a regional tool for distinguishing Precambrian from Paleozoic "Wissahickon," with those areas containing Catoctin dikes or flows clearly being Precambrian. Using this reasoning, the Wissahickon in the area of Landenberg, Chester County, and just north of Holtwood, York and Lancaster counties, would be at least as old as latest Precambrian if both sets of basalts have been correctly interpreted with respect to the Catoctin event on the basis of their composition.

Figure 1 contains field boundaries that were derived by Pearce and Cann (1973) from analyses of basaltic rocks of known tectono-magmatic affinities. The data points in Figure 1 represent analyses of samples from the present study that are believed to be derived from the Catoctin event in Pennsylvania. Although it has been popular in recent years to bash the use of such diagrams, we find that their discriminant (pun intended) use can be very powerful. At the minimum they tend to group similar basaltic rocks. At best, they provide one tool to interpret tectonomagmatic environments.

For the Catoctin event, the majority of analyses plot as "within plate basalts" (WPB), suggesting that the present South Mountain anticlinorium was not an active margin, such as a locus of ocean floor basalts, or an island arc setting, etc. Together with the observed sub-aerial igneous textures, such diagrams suggest that, in Pennsylvania, we are dealing with basalts from within a continental plate.

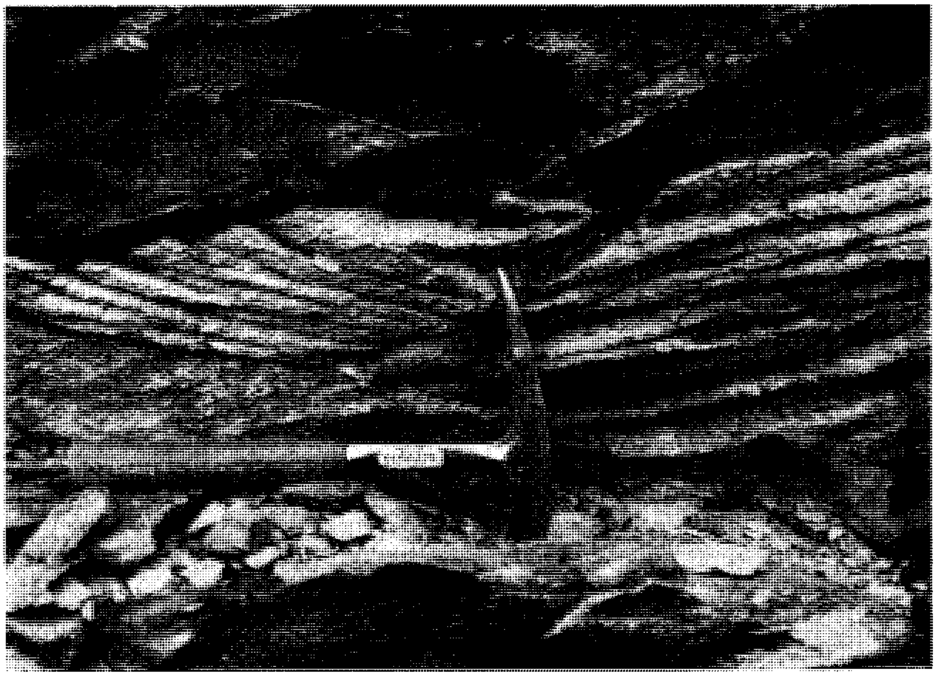
One-sixth of the samples plot outside the WPB field. Most are from the Pigeon Hills and the portion of the South Mountain

Plate III. Opposite page.

A. Ropy pahoehoe at top of a basalt flow, south of Mt. Hope.

B. More distant view of ropy pahoehoe in A.

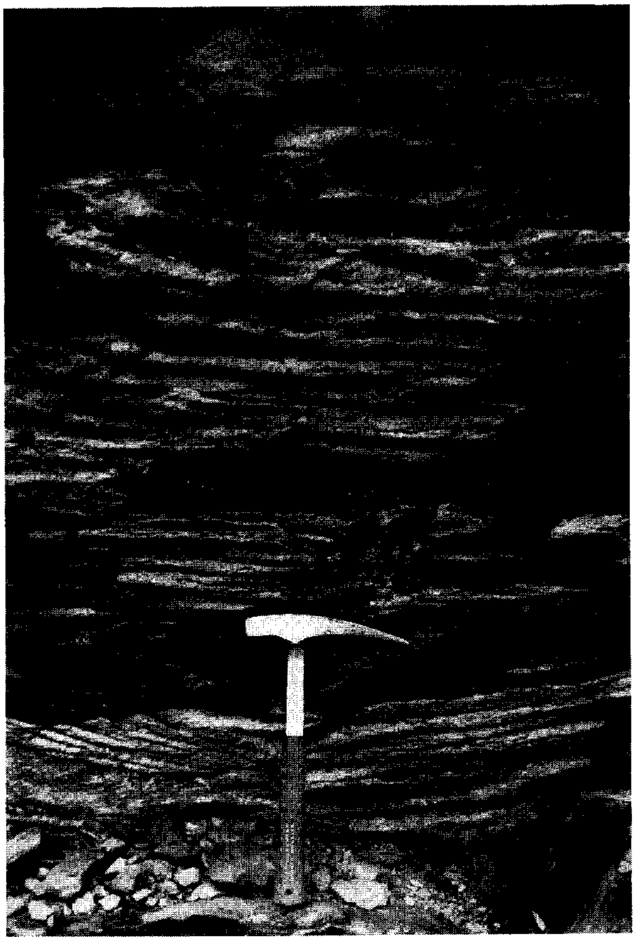
C. End view of pipe vesicles near the base of a basalt flow, south of Mt. Hope.



A ↑

B ↓

C ↓



anticlinorium southeast of the Tunnel Hill-Jacks Mountain fault system. This is the area of Catoctin metabasalt in Pennsylvania that has the greatest affinity with the Catoctin of Maryland and northern Virginia. This geochemically "true Catoctin" continuation of the Blue Ridge is where one might expect a greater oceanic or deeper basin influence in both basalts and clastic cover rocks. Also, we would not wish to rule out the possibility of additional, parallel Catoctin rift basins, with the deeper water basins located farther toward the present southeast.

Several elements tend to remain relatively constant within each of the five geochemical groups which have been defined on the basis of TiO_2 content, as opposed to simply varying directly or inversely with the TiO_2. This suggests that each group represents a distinct pulse of igneous activity, each of which might represent a large, new, crustal magma chamber with a somewhat distinct geochemical signature.

As to the relative ages of the Catoctin event pulses, we suspect a trend toward higher TiO_2 in successively younger pulses. This would make the 3.71 to 4.57% TiO_2 group, herein informally called the "Conewago Narrows," the youngest. However, overturning, several generations and scales of faulting, metamorphism, and scarcity of good outcrop have conspired to make this an uncertain proposition. Even our efforts to sample the apparent stratigraphically highest metabasalts have been vexed by structure. Thus, some of the available clues have come from beyond the South Mountain anticlinorium. First, a chilled metadiabase dike with 3.22% TiO_2 cuts metadiabase with 2.57% TiO_2 near Huffs Church in the Berks County portion of the Reading Prong. Second, three amphibole gneiss bodies near the Avondale Prong, Chester County, the map patterns of which suggest a right-side-up series of basalt flows contain 4.50% (upper), 3.04%, and 1.94% (lower) TiO_2. Third, in the Holtwood Dam area of York County, TiO_2 contents of 2.26%, 2.24%, and 1.67% are encountered from upper to lower flows. Within the South Mountain anticlinorium, where basaltic rocks containing 4.06 and 1.76% TiO_2 occur within 100 m of each other, the latter is cut by as yet unanalyzed dikelets. Together, these observations suggest a trend to higher TiO_2 between pulses.

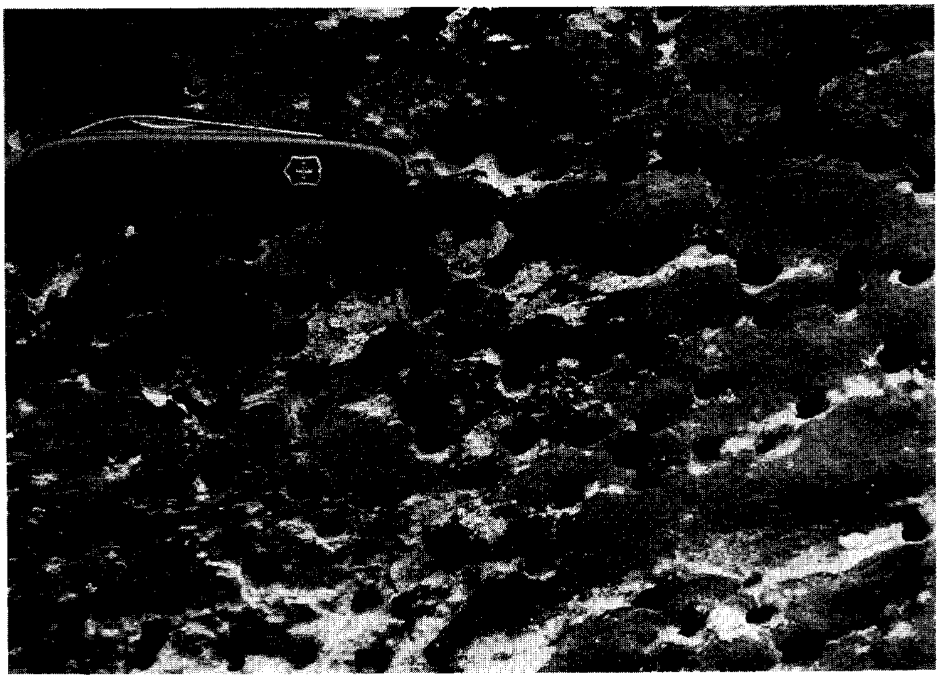
The variation for some elements that does exist within a pulse is seen as a result of normal igneous differentiation within a crustal magma chamber. Within a pulse, we now have samples from several partial sections through series of basalt flows with known facing directions (such as at Stop 3) and lack

Plate IV. Opposite page.

A. Close-up end view of pipe vesicles looking down section, south of Mt. Hope.

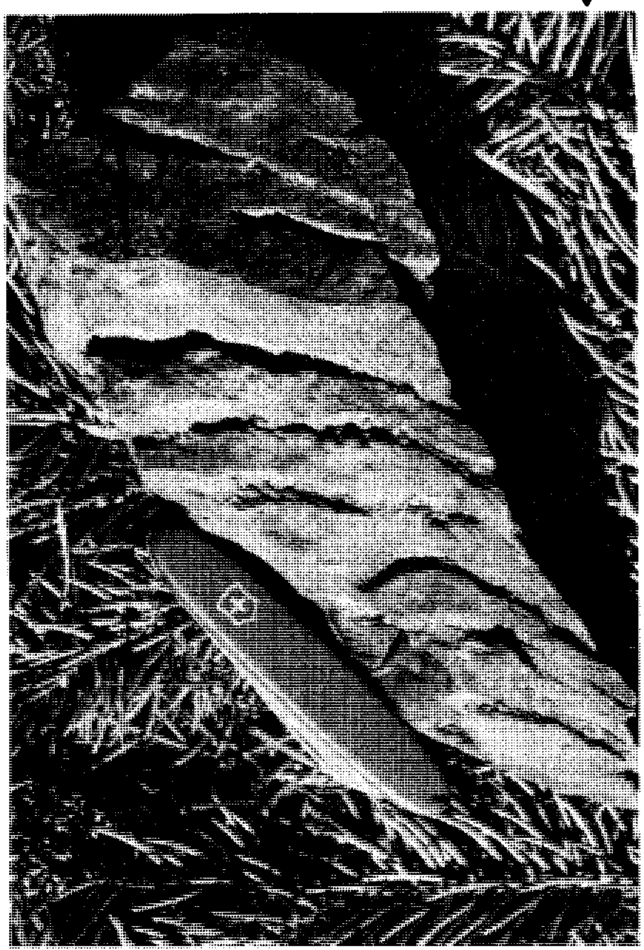
B. Cross section through pipe vesicles in a hand sample of metabasalt. Vesicles curve downward on the right into cleavage. Sample orientation strike normal with east to right. South of Mt. Hope.

C. Oblique view of possible upper chilled surface of a now overturned basalt flow above hammer head. Southwest of Mt. Hope.



A ↑

B ↓



C ↓

Table 1. Catoctin metabasalt geochemical and locality data, listed in decreasing order of $\%TiO_2$. A=Accomac. P=Piedmont. PH=Pigeon Hills. RP=Reading Prong and Womelsdorf outlier. SM=South Mountain. D=dike. F=flow. S=sheet. Analytical data expressed as parts per million (ppm) except where indicated otherwise.

Sample	Cr	Hf	La	Nb	$P_2O_5\%$	Ta	Th	$TiO_2\%$	Y	Zr	Location and Form	Latitude-N	Longitude-W
$4.60 > TiO_2\% > 3.60$													
SHELQ	16	5.1	25	25	.86	1.1	1.1	4.57	48	270	SM-F	39°46'32"	77°24'36"
CONNAR	17	7.3	32	29	.72	1.3	2.0	4.54	54	370	SM-F	39°55'56"	77°19'37"
LANS	2	6.6	22	27	.42	1.1	1.5	4.50	50	280	P-F(?)	39°46'21"	75°46'03"
FIKHL	49	7.4	30	29	.56	1.3	2.1	4.22	56	340	SM-F	40°03'15"	77°08'46"
11050TC	29	7.6	34	29	.60	1.4	2.3	4.18	64	310	SM-F	40°02'44"	77°08'56"
ACC330E	3	9.4	40	37	1.32	1.8	2.4	4.06	66	160	P-F	40°02'39"	76°33'51"
PATPK	14	7.1	30	29	.52	1.3	3.3	3.88	52	320	P-D	40°05'46"	75°44'11"
ANTRES	11	5.6	24	24	.68	1.1	1.5	3.84	42	290	RP-D	40°21'17"	75°52'10"
CONNAR2	43	6.3	25	27	.38	1.6	1.7	3.71	50	250	SM-F	39°59'00"	77°15'48"
BROOKM	39	6.1	23	25	.40	1.2	1.7	3.71	50	230	P-D	40°06'02"	75°31'16"
Median	16	6.9	28	28	.58	1.3	1.8	4.12	51	285	---		
$3.60 > TiO_2\% > 3.10$													
LS3	33	5.5	24	25	.44	1.1	1.5	3.34	44	280	RP-D	40°26'29"	75°40'37"
TOPFND	71	4.1	19	18	.59	1.0	1.2	3.29	36	180	RP-D	40°27'55"	75°41'07"
LS2	29	5.2	23	21	.50	1.1	1.3	3.29	42	230	RP-D	40°26'26"	75°40'14"
ACCOM	22	5.8	14	24	.37	1.0	2.0	3.22	36	250	A-F	40°02'38"	76°33'57"
Median	31	5.4	22	24	.47	1.0	1.4	3.29	39	240	---		
$3.10 > TiO_2\% > 2.60$													
LANM	5	4.3	17	18	.22	.9	1.1	3.04	38	160	P-F(?)	39°46'36"	75°45'40"
ISHMTN	53	3.8	17	17	.46	.7	1.1	3.02	36	190	RP-D	40°24'45"	75°52'43"
BUCKV	170	5.0	16	16	.40	.6	.4	2.95	54	230	SM-F	39°56'22"	77°22'57"
PA94	23	5.2	21	21	.30	1.1	1.5	2.86	52	240	SM-D	40°01'48"	77°08'12"
UWC	46	4.3	15	18	.30	.9	1.1	2.78	38	210	P-D	40°03'53"	75°38'20"
D-266	39	5.2	20	20	.35	.8	1.1	2.78	36	225	RP-F(?)	40°17'51"	76°08'21"
DV	46	4.1	13	19	.30	.7	1.0	2.76	38	190	P-D	40°05'14"	75°32'27"
LYDRY	42	4.6	16	19	.34	.8	1.2	2.74	36	210	RP-D	40°28'16"	75°45'13"
ACCPU	120	4.6	16	19	.28	1.0	1.3	2.72	34	180	A-F	40°02'40"	76°33'50"
CRC	49	3.9	17	16	.40	.7	1.0	2.64	42	250	SM-F	39°46'27"	77°25'50"
Median	46	4.5	16	18	.32	.8	1.1	2.78	38	210			
$2.60 > TiO_2\% > 2.00$													
ACC48E	96	3.9	16.0	17	.24	.7	1.1	2.51	32	190	A-F	40°02'38"	76°33'26"
PIGHL	405	3.8	7.4	11	.25	1.0	.7	2.48	34	155	PH-F(?)	39°51'40"	76°58'16"
LUDCOR	16	3.7	17.0	19	.28	.8	1.6	2.45	28	155	P-D	39°45'59"	76°16'29"
FR1400	110	3.6	9.7	14	.30	.7	.7	2.37	36	150	SM-F	39°52'13"	77°23'50"
ADVQ	120	2.2	6.2	11	.24	<.1	.3	2.36	32	200	SM-F	39°44'52"	77°27'01"
STPFA	89	3.7	12.0	17	.26	.7	1.1	2.31	36	175	P-D	40°08'57"	75°01'39"

Table 1. Continued.

Sample	Cr	Hf	La	Nb	P_2O_5%	Ta	Th	TiO_2%	Y	Zr	Location and Form	Latitude-N	Longitude-W
2.60 > TiO_2% > 2.00 (continued)													
HLTWCOL	190	2.6	7.5	10	.27	<.1	.3	2.26	30	110	P-F(?)	39°49'29"	76°20'28"
HLTWM	230	2.5	7.1	9	.20	.3	.4	2.24	30	100	P-F(?)	39°49'30"	76°20'26"
SNOMTN	120	2.3	7.3	11	.28	.3	.4	2.24	34	160	SM-F	39°50'02"	77°28'55"
CRPC	120	3.2	11.0	13	.22	.4	.7	2.23	36	230	SM-F	39°46'26"	77°25'49"
ELBNW	18	3.4	13.0	16	.28	.7	1.3	2.21	30	200	P-D	40°05'26"	75°36'14"
MVF	140	3.0	6.5	11	.20	.6	.7	2.05	30	150	SM-F	39°57'50"	77°22'05"
GRNRDG	130	2.6	7.8	8	.30	.4	.2	2.04	30	110	SM-F	39°52'20"	77°24'07"
CAT	110	2.5	6.7	10	.25	.3	.4	2.02	30	140	SM-F	39°45'24"	77°27'36"
Median	120	2.8	7.6	11	.26	.5	.7	2.25	31	155			
2.00 > TiO_2% > 1.20													
WMM	180	3.6	11.0	13	.26	.6	.7	1.99	40	240	SM-F(?)	39°44'57"	77°26'59"
HNYNW	33	2.8	9.2	15	.20	.5	.7	1.97	26	130	P-D	40°03'32"	75°56'47"
LANN	240	2.7	8.5	9	.24	.3	.7	1.94	26	110	P-F(?)	39°46'26"	75°46'11"
FURNCK	85	4.0	17.0	16	.32	.8	.9	1.93	32	180	RP-S	40°19'22"	76°10'11"
MP87	170	2.2	6.9	9	.22	.4	.3	1.92	28	100	SM-F	39°43'14"	77°22'20"
PIGHL2	230	2.5	7.0	7	.28	<.1	.3	1.85	36	160	PH-F	39°52'11"	76°57'47"
1105NW	260	2.6	9.9	29	.16	.6	.7	1.76	24	110	SM-S(?)	40°02'52"	77°08'50"
HNY82	90	3.1	11.0	12	.16	.6	1.0	1.75	28	130	P-D	40°01'48"	77°08'12"
ACC180E	180	2.0	5.1	11	.14	.4	.4	1.73	22	170	A-F	40°02'08"	76°33'56"
HL888	240	2.2	7.4	8	.16	.2	.2	1.73	24	160	PH-F(?)	39°52'10"	76°56'52"
HLTWBASE	110	1.8	5.6	9	.21	<.1	.3	1.67	26	115	P-F(?)	39°49'32"	76°20'27"
RTE16	220	2.5	6.2	9	.16	.3	.4	1.64	30	100	SM-F	39°44'07"	77°27'35"
FRNDCK	110	2.3	6.9	9	.14	.5	.4	1.61	28	100	SM-F	39°43'28"	77°22'52"
MRCK	170	2.4	8.5	11	.16	.4	.5	1.59	22	110	P-D	40°04'01"	75°44'00"
PIGHL3	130	1.9	5.6	9	.12	.3	.3	1.35	28	140	PH-F	39°51'34"	77°58'10"
Median	170	2.4	7.4	9	.16	.4	.4	1.75	28	130			

only analyses to prove this differentiation trend. If subtle color change is any guide, within a pulse we expect the normal trend from more mafic to less mafic, i.e., decreasing TiO_2. If so, this geochemical criterion may also give us a facing direction where primary igneous features have been erased.

As to the source of the magmas, little can be said with any degree of certainty except that the more alkali Conewago Narrows basalts (3.71 to 4.57% TiO_2) are probably the result of deeper (or a lesser percent of) melting in the mantle. If there really is a trend toward higher TiO_2, more alkali pulses with time, then one must somehow postulate a system in which each magma pulse is generated from deeper or related mantle regions, but in successively shorter time spans. Such might be the case in a rift environment, where progressive dilation of the crustal cover is in progress.

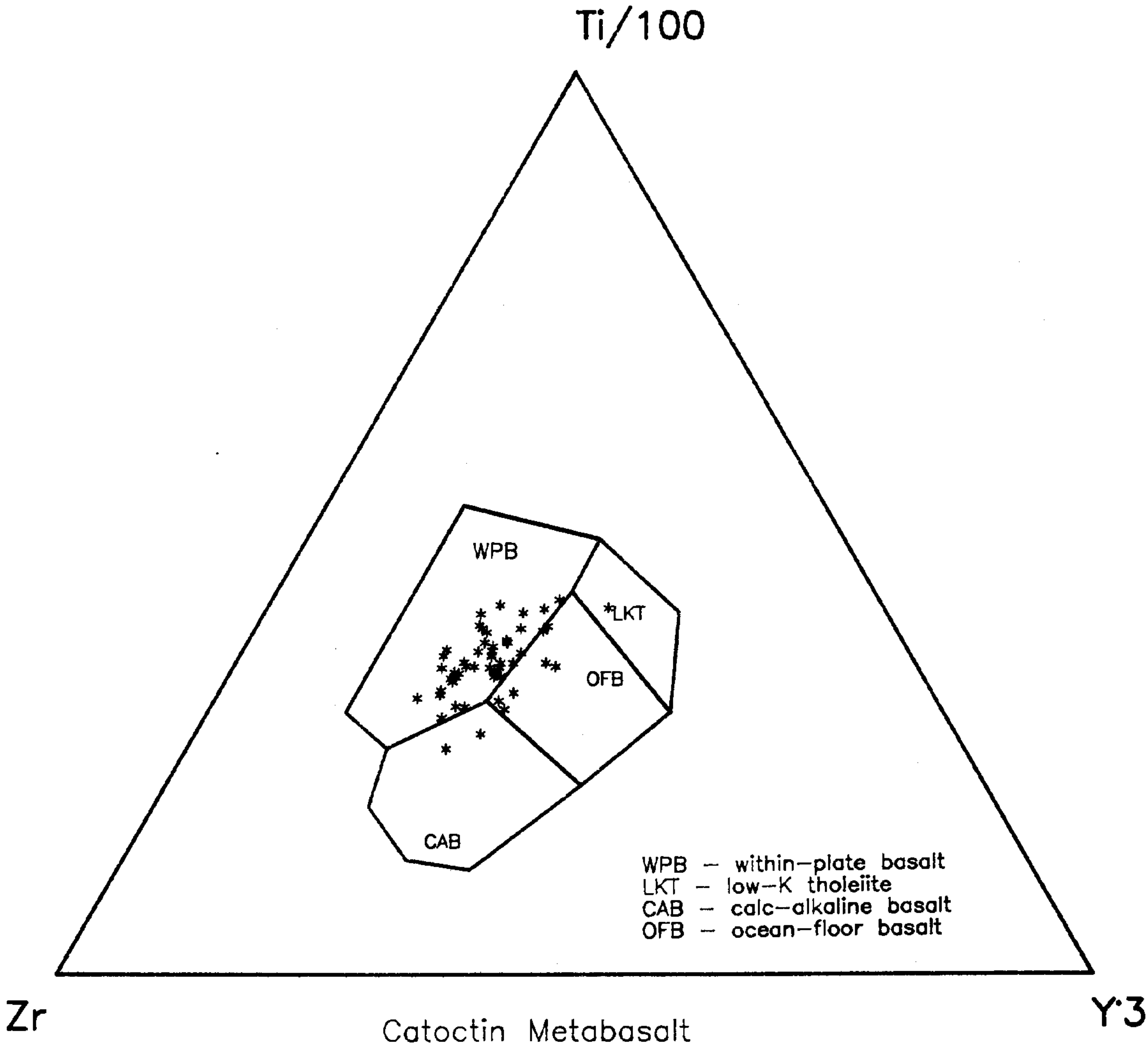


Figure 1. Ternary plot of three relatively incompatible and immobile trace elements for available samples believed to be of Catoctin affinity from Pennsylvania. Field boundaries after Pearce and Cann (1973).

As noted earlier, the adjacent Early Mesozoic basaltic rocks in Pennsylvania yield depleted trace element signatures in conflict with their rift association. This discrepancy would, in large measure, be resolved if the voluminous, somewhat alkali Conewago Narrows basalts of the Catoctin event were derived from the same region of the mantle as those in the Early Mesozoic basins. That this is a possibility is suggested by the apparent complimentary nature of the TiO_2, Nb, Ta, etc. trends of the apparent youngest Catoctin and oldest Early Mesozoic basalts. If the complimentary nature of the geochemical trends relative to a primitive mantle (Hofmann, 1988) is real, then the mantle beneath southeastern Pennsylvania has remained attached to the continental crust, as a keel is to a boat, from ~600 Ma to ~200 Ma. Acknowledging the fact that interpretations of the original map locations of both the Catoctin and Mesozoic are uncertain, nevertheless, it appears that related crust and mantle processes were involved in both late Precambrian and Mesozoic continental rifting.

STRUCTURAL IMPLICATIONS

Root and others (this volume) summarize the regional geology, age, and structure of the South Mountain anticlinorium. Here, we need only to mention a few aspects that relate more directly to metabasalt: fault-bounded plates and overturning.

Plates

For our purposes, the South Mountain anticlinorium can be divided into 4 distinct "thrust" plates (Figure 2). The main plate boundaries are defined by two east-west trending faults, the Shippensburg fault and the Carbaugh-Marsh Creek fault, and the northeast-trending Tunnel Hill-Jacks Mountain fault system.

Metabasalts in the overriding, exposed plate to the southeast of Jacks Mountain ("Catoctin" plate) are believed to correspond directly with those of the Blue Ridge of Virginia and the Catoctin Mountains of Maryland. The metabasalts in the Pennsylvania portion of this block have TiO_2 contents that typically range from about 1.5 to 2.0%. Here, near-surface metabasalts predominate over metarhyolites in the approximate ratio of 10 to 1. The "Catoctin" plate appears to project across the Mesozoic Basin toward the Pigeon Hills and Accomac metabasalts based on an aeromagnetic high trending about N60^{o}E (Zeitz and others, 1980 and D. B. MacLachlan, personal communication, 1990). Furthermore, metabasalts in the "Catoctin" plate are somewhat similar chemically to those in the Pigeon Hills. These latter typically contain 1.4 to 2.0% TiO_2. The Accomac metabasalts are also on the aeromagnetic trend, but chemically seem to contain nearly the full range of TiO_2 contents observed in Pennsylvania in the five geochemical groupings. Besides the typical "Catoctin" plate TiO_2 contents of 1.5 to 2.0%, the chemical equivalents of metabasalts found in other plates are probably also present at Accomac. (See Table 1, ACC series samples which contain from 1.7 to 4.1% TiO_2.)

The plate between Jacks Mountain on the southeast and the Carbaugh-Marsh Creek fault includes both Fauth's (1978, Figure 2) Green Ridge and Snowy Mountain metabasalt belts. This lumping of metabasalt belts is done herein on the basis of similar textures (pyroclasts; agglomerate; albite-rimmed amygdules; and thin, highly fluid flows) and chemistry (TiO_2 typically in the 2.0 to 3.0% range for several metabasalt bodies). Unlike the "Catoctin" block, the volcanics in this more northwesterly block consist of perhaps 80% metarhyolite and 20% metabasalt. Interpretation of geologic maps suggests that locally the "last gasp" volcanics consist of metarhyolite. The metabasalt body observed farthest to the southeast in this plate is in fault contact with the Mesozoic Basin (Fauth, 1978) and contains 4.6% TiO_2. This very high TiO_2, "Conewago Narrows" metabasalt series has been found in the southeastern portion of all but the "Catoctin" plate.

The middle plate, bounded on the south by the Carbaugh-Marsh

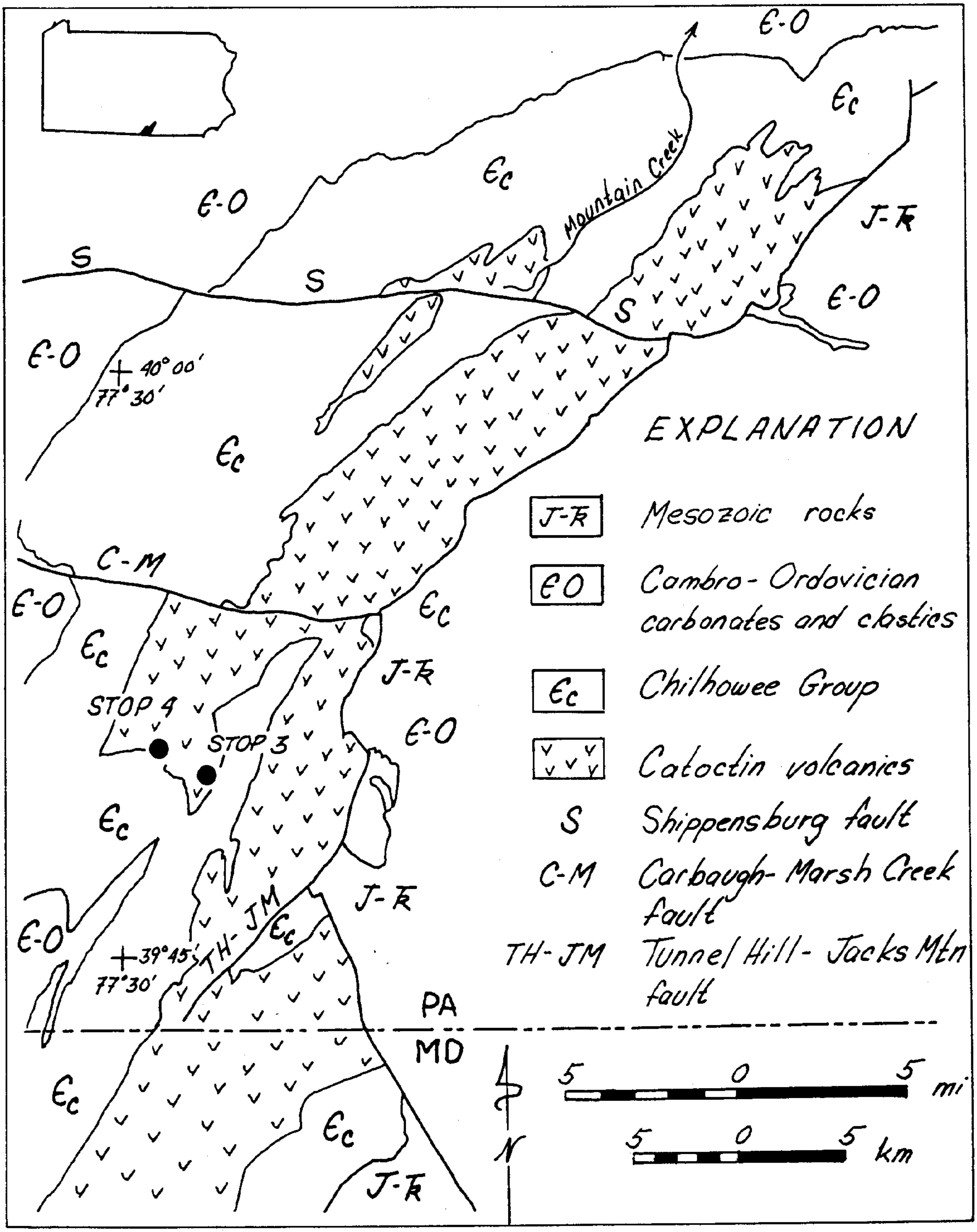


Figure 2. Geologic map of the South Mountain anticlinorium generalized from Berg and others (1980) and Cleaves and others (1980). The geologic hammer is 32.8 cm long and the pocket knife is 8.4 cm long.

Creek fault and on the north by the Shippensburg fault contains less than 10% metabasalt at the surface. Within this plate, there appear to be two geochemical basalt series. One, herein informally called the "Buchanan Valley" series metabasalts, contains (based on only 2 samples) 2.0 and 3.0% TiO_2. The second, the "Conewago Narrows" series, contains (based on 2 samples in this plate) 3.7 and 4.6% TiO_2. Both series presumably correspond to their TiO_2 content analogs in the block to the south of the Carbaugh-Marsh Creek fault. A third series might occur to the west in the headwaters of Mountain Creek Valley (Stose and Bascom, 1929), but no analyses are available. This metabasalt is in places interlayered with metarhyolite and appears somewhat felsic. Outcrops of metabasalt observed in the Dickinson quadrangle portion occur within a few meters of the Loudoun Formation.

Finally, the plate north of the Shippensburg fault lacks any known exposed metabasalt to the northwest of Mountain Creek, but basaltic material with a possible intrusive texture has been encountered there only in the subsurface. To the southeast of Mountain Creek, a few metabasalt bodies of the "Conewago Narrows" series containing about 4.2% TiO_2 occur.

Together, these data might be interpreted to suggest that the previously identified faults separate related volcanics which are now observed at different structural levels. The most profound break occurs across the Tunnel Hill-Jacks Mountain fault zone, which appears to separate metabasalts of the lithologic continuation of the Blue Ridge of Maryland and northern Virginia from Pennsylvania's more diverse volcanics.

Overturning

Root and others (this volume) cite several classical studies that document overturning of the South Mountain anticlinorium and some of the defining clastics. Reconnaissance for this Field Conference adds to the number of overturned outcrops of metabasalt reported by Fauth (1978), including the BUZZE metabasalt at Stop 3. Because the BUZZE metabasalt outcrop pattern has a linear trace, the assumption that most overturned outcrops are a local phenomenon due to high order folding is hard to accept. As this goes to press, we still have not imagined a satisfactory cartoon, but we now suspect that: (1) overturning of the metabasalts is the rule rather than the exception; (2) right-side-up volcanics and superimposed clastics in the study area may occur at higher elevations, where they are above low-angle thrust faults, as veneers developed on Loudoun and similar lithology detachment zones (based in part on multiple benches of the latter); and (3) except for the "Cactoctin" plate, the eastern limb of the anticlinorium is incomplete with respect to metabasalt.

If overturning is as pervasive as suspected, some of the areas mapped as Loudoun Formation should be reexamined. As noted by Fauth (1968, p. 27), certain lithologies originally included in the Loudoun (i.e., above the Catoctin volcanics) "were, in fact, part of the older (sub-Catoctin) but lithologically similar Swift Run Formation." The laminated lithology on the southeast side of Piney Mountain is particularly suspect because of its resemblance to the varved, lacustrine(?) portion of the Faquier Formation of the Marshall Quadrangle, Virginia (Espenshade, 1986), a possibly correlative eastern equivalent of the pre-volcanic Swift Run Formation.

THE LOWER CAMBRIAN CLASTICS OF SOUTH MOUNTAIN, PENNSYLVANIA

by
Marcus M. Key, Jr.
Department of Geology, Dickinson College, Carlisle, PA

INTRODUCTION

The northeast/southwest trending linear ridges of South Mountain are composed of the Lower Cambrian clastic metasedimentary rocks of the Chilhowee Group (Figure 3). The group comprises four formations (from oldest to youngest): Loudoun, Weverton, Harpers, and Antietam. In the South Mountain area, the thickness of the group ranges from roughly 3,000 feet in the Mt. Holly area to 5,000 feet in the Caledonia area. The group consists of lithogenic sediments which originally were shales, litharenites, sublitharenites, quartz arenites, and conglomerates that have been metamorphosed into phyllites, quartzites, and squashed-pebble conglomerates. Currently only the Montalto Member of the Harpers Formation and the Antietam Quartzite are economically important as a source of coarse and fine aggregate, and colored sands for speciality uses.

AGE

The age of the Chilhowee Group has historically been a troublesome issue (Nickelsen, 1956). The problem existed because of a paucity of datable materials in these rocks. The age of the group is now rather certain to be Early Cambrian based on the radiometrically dated, stratigraphically older Catoctin Formation and two biostratigraphically dated formations in the upper and lower units of the Chilhowee Group.

Until recently, the age of the Catoctin Formation was widely disputed with radiometric dates ranging from 420+4 Ma (Silurian) (Nagel and Mose, 1984) to 820 Ma (Precambrian) (Rankin and others, 1969). Recent radiometric ages have been much more consistent. The metarhyolites in the Catoctin Formation from South Mountain have been dated at 597+18 Ma (Aleinikoff and others, 1991). Badger and Sinha (1988) determined the age of the Catoctin in west-central Virginia to be 570+36 Ma. The stratigraphically older Lynchburg Formation in Virginia has been dated at 600 Ma (Mose and others, 1985). All of these recent dates suggest the Catoctin Formation is latest Precambrian in age.

Biostratigraphic contraints on the age of the Chilhowee Group come from fossils in the upper and lower parts of the group. The Early Cambrian age fossils first reported by Walcott (1891) from the Antietam Quartzite have been found throughout the Appalachians in this formation and its stratigraphic equivalents to the south (See Yochelson, this guidebook). The age of the upper Chilhowee rocks has been repeatedly confirmed with different index fossils and in different areas along the Appalachians. The Antietam Quartzite and its stratigraphic equivalents contain a diverse fauna indicative of the Early Cambrian. This fauna contains acritarchs, the worm tube ichnofossil *Skolithos linear-*

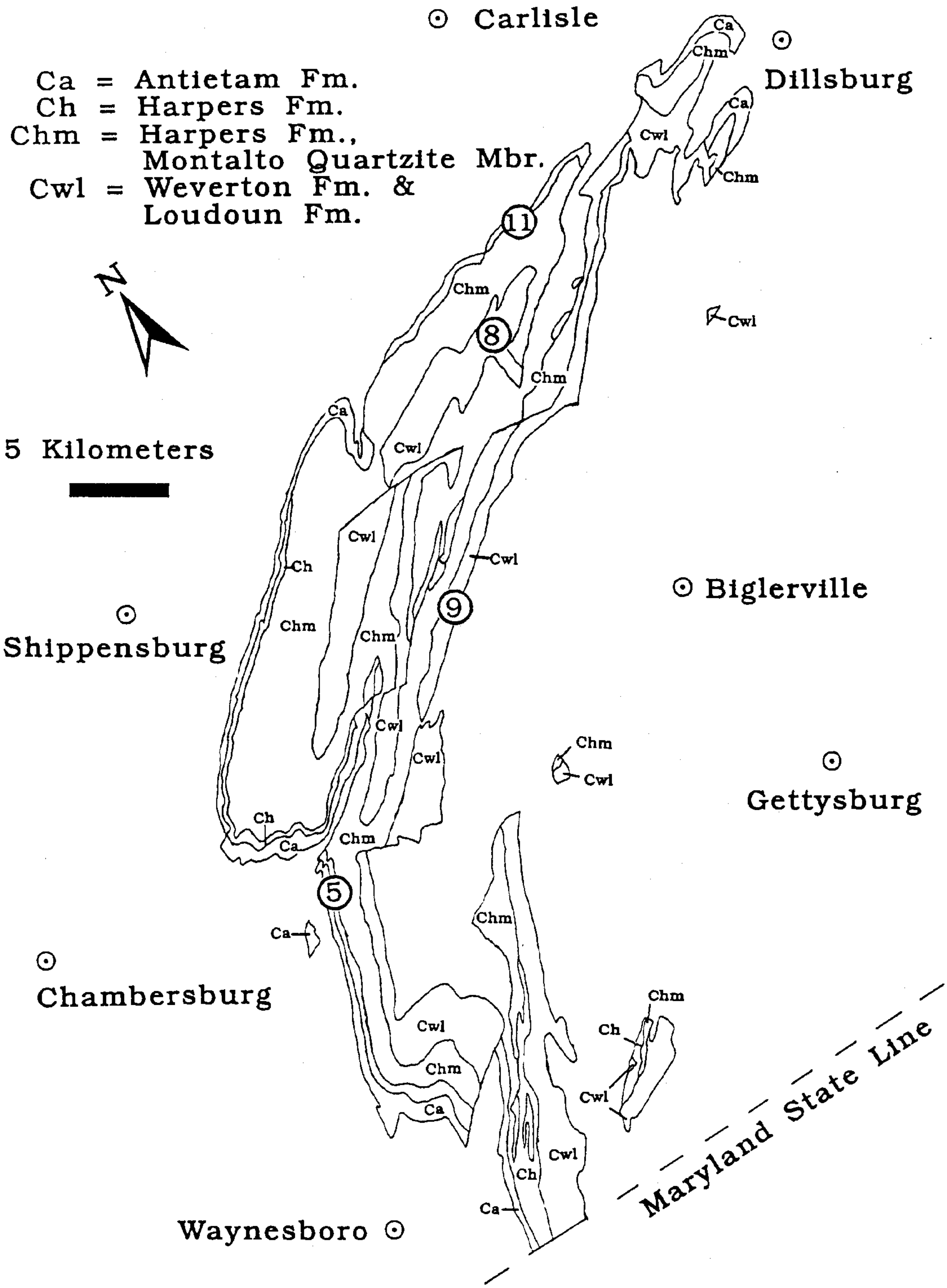


Figure 3. Geologic map of South Mountain area showing distribution distribution of Chilhowee Group formations (Modified from Berg and others, 1980). Numbers refer to Chilhowee Group field trip stops.

is, the inarticulate brachiopod *Obolella*, the articulate brachiopod *Camarella minor*, the trilobite *Olenellus*, the primitive mollusc *Hyolithus communis*, and the ostracod *Indiana tennesseensis*. This fauna has been reported from Pennsylvania to Tennessee (Walcott, 1896; Stose, 1909; Bassler, 1919; Barrell, 1925; Stose and Bascom, 1929; Stose, 1932; Resser and Howell, 1938; Amsden, 1951; Stose, 1953; Stose and Stose, 1957; Laurence and Palmer, 1963; Wood and Clendening, 1982). The occurrence of *Olenellus* constrains the Antietam Quartzite to the *Bonnia-Olenellus* Biozone which existed from 550-540 Ma (Fritz, 1972; Palmer, 1981, 1983). This dates the Antietam Quartzite as latest Early Cambrian.

The presence of Skolithos linearis in the Montalto Member of the Harpers Formation does not provide any biostratigraphic control because this trace fossil occurs in both Cambrian and Precambrian deposits (Fritz and Crimes, 1985). Recently an age diagnostic fauna has been found in the stratigraphic equivalent of the Weverton Quartzite in southwestern Virginia (Simpson and Sundberg, 1987). This depauperate fauna contains many small shelly fossils and the trace fossil Rusophycus which is indicative of earliest Early Cambrian (Fedonkin, 1981). This constrains the age of the lower Chilhowee Group to 570-560 Ma (Sepkoski and Knoll, 1983).

The radiometric dates of the Catoctin Formation place a maximum age on the Chilhowee Group of roughly 570 Ma. The biostratigraphic data suggest an age range of 570-540 Ma. Based on the stratigraphic conformity of all of the Chilhowee Group formations (Nickelsen, 1956; Fauth, 1968), the age of the entire group is considered to be Early Cambrian.

CATOCTIN FORMATION - CHILHOWEE GROUP CONTACT

Underlying the Chilhowee Group is the Catoctin Formation (See the paper by Smith and others, this guidebook). The Catoctin consists of aphanitic, extrusive volcanics (dominantly basalts and rhyolites) that have been subsequently metamorphosed. These represent late Precambrian rift-related volcanism (Badger and Sinha, 1991) which occurred during the opening of the Proto-Atlantic (Iapetus) Ocean (Rankin, 1975). They are inferred to have been generally extruded subaerially based on the absence of pillow lavas and the presence of columnar jointing (Reed, 1955; Fauth, 1968). Petrographic, rare earth element, and major element analyses have confirmed this (Rankin and others, 1969; Dockter, 1990). However, some local subaqueous extrusion has been documented by the presence of pillow lavas in Virginia (Bowring and Spencer, 1987; Kline and others, 1987).

The Catoctin-Chilhowee Group (i.e., Catoctin-Loudoun) contact is not exposed in the South Mountain area (Stose, 1909; Stose and Bascom, 1929; Fauth, 1968). Historically, the contact has been considered both conformable (Bloomer and Bloomer, 1947; Bloomer, 1950; Cloos, 1951; Rankin, 1967) and unconformable (Stose, 1906, 1909, 1932; Stose and Bascom, 1929; King, 1950; Stose and Stose, 1957; King and Fergusen, 1960; Rankin and others, 1969). The absence of a structural discordance between the Catoctin and Loudoun Formations suggests the contact is conformable (Bloomer, 1950; Bloomer and Werner, 1955; Reed, 1955; Fauth, 1968). Both formations exhibit similar deformation patterns (i.e., orientation of cleavage and lineations) (Stose, 1932; King, 1949; Cloos, 1951; Freedman, 1967; Fauth 1968). If there was a period of deformation prior to the deposition of the Loudoun Formation, it would indicate unconformity. Bloomer (1950), Cloos (1951), and Bowring and Spencer (1987) argue that the contact is also conformable due to the interfingering of the Catoctin volcanics with the Loudoun clastics in Virginia.

The presence of Catoctin fragments in the basal clastic sediments of the Loudoun Formation suggests the contact is unconformable [But see Bloomer (1950) for an opposing view]. This indicates the Catoctin was subaerially exposed and eroded prior to deposition of the Loudoun (Freedman, 1967). According to Reed (1955) this period of erosion is preserved by a metamorphosed saprolite at the contact. Another argument for unconformity is that the contact represents a change from terrestrial extrusion of the Catoctin to submarine deposition of the Loudoun (Nickelsen, 1956; Fauth, 1968).

The contact may be unconformable, but it probably represents only a brief time hiatus (Badger and Sinha, 1988). The previously discussed radiometric ages of the Catoctin and the biostratigraphic ages of the Chilhowee suggest little if any time is missing at this contact.

LITHOSTRATIGRAPHY

The basal Chilhowee unit is the Loudoun Formation. The sediments of this formation contain fragments of the underlying Catoctin metarhyolites (Stose, 1906, 1909, 1932; Stose and Bascom, 1929; Freedman, 1967; Fauth, 1968). Thickness in the South Mountain area ranges from 0 to 550 feet (Stose and Bascom 1929, Stose 1932, Freedman 1967, Fauth 1968). In the Caledonia and Mt. Holly areas, this unit crops out along the southeast slope of Piney Mountain. The lithology ranges from a purplish phyllite to a grayish, quartz/phyllite/rhyolite-pebble metaconglomerate (Stop 9). The grains are subrounded to subangular grains with fair sorting (Fauth, 1968).

Overlying the Loudoun Formation is the Weverton Quartzite whose thickness in the South Mountain area ranges from 600 to 1,400 feet (Stose and Bascom, 1929; Stose, 1932; Freedman, 1967; Fauth 1968). In the Caledonia area, this unit crops out along the crests of Piney Mountain and Big Pine Flat Ridge and along the southeast slope of East Big Flat Ridge. In the Mt. Holly area, it crops out along the crest of Mt. Holly (including Hammond's Rocks, Stop 8) and along the southeast slope of Piney Mountain. The lithology ranges from grayish-green and purplish quartzites and quartzose graywackes with occasional thin zones of graywacke conglomerate, quartzite, and phyllite. The grains are subrounded to subangular with moderate sorting (Fauth, 1968).

Overlying the Weverton Quartzite is the Harpers Formation whose thickness in the South Mountain area ranges from 1,500 to 3,000 feet (Stose and Bascom, 1929; Stose, 1932; Freedman, 1967; Fauth, 1968). In the Caledonia area, this unit crops out along the crests of Piney Mountain and Big Pine Flat Ridge. In the Mt. Holly area, it crops out along the crest of Piney Mountain and around the nose of Mt. Holly. There are two distinct lithologies present. The lower light-gray quartzite has good sorting with subrounded grains (Fauth, 1968). This unit is called the Montalto Quartzite Member, and it is 1,200 to 2,500 feet thick. The Montalto is crossbedded and has numerous *Skolithos linearis* worm tubes (Figure 2). In the upper part of this member is a distinctive bluish quartzite unit that ranges from 15 to 25 feet thick and is useful in mapping (Fauth, 1968). The upper greenish graywacke/phyllite is moderately sorted with subrounded grains. It is 300 to 500 feet thick (Fauth, 1968; Root, 1978).

Figure 2. *Skolithos linearis* worm tubes from the Antietam Formation.

Overlying the Harpers Formation is the youngest Chilhowee unit, the Antietam Quartzite whose thickness in the South Mountain area ranges from 440 to 900 feet (Stose, 1909; Stose and Bascom, 1929; Stose, 1932; Freedman, 1967; Fauth, 1968; Root, 1978). The Pennsylvania Geological Survey is currently studying core from an angle hole that penetrated the entire thickness of the Antietam north of Caledonia Park. In the Caledonia area, this unit crops out around the nose of Big Pine Flat Ridge. In the Mt. Holly area, it crops out along the northwest slope and the nose of Mt. Holly. The lithology is dominated by white to grayish quartzite which is medium- to coarse-grained, subrounded to rounded, with good sorting (Fauth, 1968). *Skolithos linearis* worm tubes (Figure 2) are abundant with some up to 15 inches long (Shirk, 1980; Wilshusen and Sevon, 1981, 1982). Megaripples are present in the Mt. Cydonia area (Wilshusen and Sevon, 1981, 1982) (Stop 5). There are numerous abandoned aggregate pits and quarries developed in the Antietam in the South Mountain area. Significant active aggregate operations in the Antietam occur at Mt. Cydonia (Stop 5) and Mt. Holly Springs (Stop 11) as well as Toland.

CHILHOWEE GROUP - TOMSTOWN DOLOMITE CONTACT

Overlying the Chilhowee Group is a thick sequence of Cambrian through Ordovician carbonates. The oldest of these carbonates is the Tomstown Dolomite which rests upon the Antietam Quartzite. The Tomstown is dominantly dolomitic with some minor limestone. The contact between the Antietam Quartzite and the Tomstown Dolomite is apparently gradational as evidenced by a calcareous basal shale in the Tomstown (Brezinski, 1991). In the South Mountain region, this contact is marked by the presence of kaolinite-rich white clays (Berkheiser and others, 1982). This clay formed either from dissolution of the Tomstown which resulted in the concentration of its contained lithogenic sediment (Stose, 1907) or from hydrothermal alteration of the Tomstown (Hosterman, 1968, 1969).

The contact is often assumed to be conformable even though it is not exposed in the South Mountain area (Stose, 1909; Stose and Bascom, 1929; Fauth, 1968; Freedman, 1968; Brezinski, 1991). Besides the lithologic argument for conformity mentioned above, there is also biostratigraphic support for conformity. The Tomstown Dolomite has been dated as latest Early Cambrian due to the presence of the trilobite Olenellus (Fauth, 1968). As the Tomstown and Antietam are of the same age based on the available biostratigraphic control, the contact is probably conformable (Fauth, 1968).

PALEOGEOGRAPHY

The Chilhowee Group was deposited in rift-induced sedimenttary basins that formed as a result of the opening of the Iapetus Ocean in the late Precambrian (Bond and others, 1984). In the Early Cambrian, clastic sediments (including the Chilhowee Group) were deposited along the newly-formed southern and eastern edges of the North American craton (Thomas, 1977). These sediments represent the basal transgressive sediments of the Sauk Sequence (Sloss, 1963; Brown, 1970; Schwab, 1972). The sediments were eroded from the craton, transported to the southeast (Kay, 1951; Whitaker, 1955; Brown, 1970; Schwab, 1970; Whisonant, 1970), and were deposited in shallow marine environments of the Iapetus Ocean along the coast of North America. During this time, the South Mountain area was located at roughly 15^{o} south latitude (Scotese and others, 1979).

Deposition of the Chilhowee Group began with the transgression of the Cambrian sea which onlapped from the southeast and inundated the subaerially exposed Precambrian Catoctin Formation (Nickelsen, 1956; Fauth, 1968). The first sediments resulted in the Loudoun Formation which was partially derived from and deposited on the Catoctin Formation. This is evidenced by the clasts of Catoctin metarhyolites which were incorporated into the basal Loudoun sediments.

Evidence of a northwestern cratonic source area for the Chilhowee Group comes from stratigraphic thicknesses, paleocurrent indicators, and facies distributions. The Loudoun Formation and the Weverton Quartzite thicken to the west (Swartz, 1948). The Montalto Quartzite Member of the Harpers Formation thickens to the north (Fauth, 1968). Stose and Bascom (1929) reported that the entire Chilhowee Group thickens to the northwest. Crossbedding in the Weverton Quartzite indicates a western source area (Whitaker, 1955), while ripple marks in the Antietam Quartzite suggest a northwestern source area (Wilshusen and Sevon, 1981, 1982). The upper Chilhowee formation grades eastwardly from the Antietam Quartzite in the Blue Ridge Mountains to the Chickies quartz-mica schist in the Piedmont (Freedman, 1968). This facies change suggests deeper water to the east and a westward source area.

The depositional environments represented by the Chilhowee Group in the South Mountain area are dominantly shallow marine on continental shelf environments (Nickelsen, 1956). This is evidenced by the presence of trilobites, brachiopods, ostracods, and worm tubes. Water depth increased to the southeast toward the deep-marine Wissahicken basin (Thomas, 1977) where the Chilhowee Group was deposited below storm-wave base (Freedman, 1968; Simpson and others, 1991).

The depositional environment of the Antietam Quartzite can be further refined as a shallow marine beach setting. The Antietam's clean, well-rounded quartz sand and the presence of megaripples led Wilshusen and Sevon (1981, 1982) to interpret this formation as reflecting a near-shore to offshore, below normal wave-base setting. Based on the outcrop pattern and the morphology of the sand bodies, Kauffman and Frey (1979) inferred that the Antietam was deposited in a barrier island environment. The occurrence of *Skolithos linearis* worm tubes in the Antietam places this formation in the *Skolithos* ichnofacies. This facies is indicative of the intertidal to subtidal zone of a shallow, high energy, coastal, beach environment (Seilacher, 1967; Crimes, 1970). The Montalto Quartzite Member of the Harper Formation has identical worm tubes, and it is interpreted to have a similar depositional environment.

Fluctuations in local relief, rate of uplift, and rate of erosion in the source area relative to the rate of subsidence and deposition in the depositional basins resulted in local fining and coarsening trends within the Chilhowee Group. On a regional scale over the course of Chilhowee deposition, sediments generally became finer as well as texturally and mineralogically more mature in response to rising sea level and increasing tectonic stability of the east coast of the North American craton (Fauth, 1968; Patterson and Simpson, 1991). The maturing upward trend is evidenced by a general increase in the degree of sorting and roundness as well as an increase in the amount of quartz clasts relative to those of feldspar (Fauth, 1968). In the South Mountain area, this can be grossly seen by comparing the immature Loudoun conglomerates (Stop 9) with the mature Antietam quartz arenites (Stops 5 and 11). The fining upward trend continued after Chilhowee deposition into the Tomstown Dolomite. Before the end of the Early Cambrian, the sediment became finer until little or no lithogenic sediment was entering the sea and carbonate biogenic sedimentation was initiated (Freedman, 1967, 1968).

System	Series	Formation name	Symbol	Columnar section	Thickness in feet	Character of rocks	Character of topography
ORDOVICIAN	Lower (?) Ordovician	Beckmantown (?) limestone Concealed	Ob		300+	Blue limestone, in part banded and earthy, weathering to tripoli.	Lowlands with few outcrops.
		Tomstown dolomite	Ct		1000±	Coarse glistening gray dolomite, blue limestone, and some shale and sericite schist.	Valleys.
CAMBRIAN	Lower Cambrian	Antietam sandstone	Ca		800	Coarse-grained quartzose sandstone with numerous scolithus tubes and hard vitreous quartzite.	Mountain slopes and small ridges and knobs.
		Harpers schist with Montalto quartzite member	Ch (Cma)		3000	Dark hackly sandy slate or schist and white quartzite (Montalto quartzite member) the latter in thin beds at the south which expand northeastward to represent most of the formation. The upper part of quartzite is filled with scolithus tubes.	High ridges and mountains with sharp valleys on the schist beds.
		Weverton sandstone	Cw		750	Gray and purplish feldspathic sandstone and quartz conglomerate.	Crests of high ridges and steep mountain slopes.
ALGONKIAN		Loudoun formation	Cl		550	Soft purplish arkosic conglomerate, and fine sericite slate.	Steep mountain slopes.
		Unconformity					
		Metabasalt (greenstone)	mb		1000±	Massive to schistose greenstone or altered basalt lava, massive bluish gray rock in places, with green epidote and white quartz masses.	Mountains with ridges, rounded knobs, and open intermountain valleys.
		Aporhyolite	ar		1000±	Bluish gray to pink and purple felsite with white phenocrysts, altered rhyolite lava, and red breccia and tuff probably at top, and bands of white to red sericite schist.	

Fig. 3. Columnar section of South Mountain area.

From Stose, 1932.

FOSSILS AT MOUNT HOLLY SPRINGS, PENNSYLVANIA - A BRIEF HISTORY

by
Ellis L. Yochelson, U. S. Geological Survey (Retired) and
Research Associate, National Museum of Natural History,
Washington, D. C. 10560

Washington, D. C. is a terrible place in the summer, with unremitting high heat and equally high humidity. Before air conditioning it was "impossible" and anyone who was anybody, the socially elite, the President, and the Congress, went elsewhere during the summer. If possible, those in the middle and lower levels of government, sent their families away and stole an occasional few days to visit them. Thus it was that Mrs. Walcott and young Charlie moved to Mount Holly Springs, Pennsylvania in mid-July, 1892.

Washington was worse than normal that July for this was the year that Congress slashed the U. S. Geological Survey appropriation, partly in an effort to get John Wesley Powell to resign. The budget of Charles Doolittle Walcott, Chief Paleontologist, dropped from $51,700 to $14,000, requiring terrible lay-offs, retrenchment, and destruction of ongoing programs (there really is very little new in history of geology). Simultaneously, Walcott was propelled toward the position of Chief Geologist, because Powell was leaning heavily on him for assistance. Walcott did not get to see his family until the first week of August.

There is nothing like being in the field to clear out cobwebs. Add to that, Walcott was a workaholic and nearly a compulsive hammerer on rocks. After arriving at Mount Holly Springs late on August 5th, the following day Walcott recorded in his pocket diary that, with Helena, he was "Out on the quartzite rock in Mt. Holly Gap. Found fragments of the Olenellus fauna." A few more fragments turned up the next day and Walcott then went back into the maelstrom of government.

These few fragments and the brief mention of their discovery were fundamental to the geology of Pennsylvania and Maryland. In brief,during the mid-1880's, the Norwegian geologist W. C. Brøgger had insisted that the sequence of trilobite zones generally accepted in North America was incorrect, thereby causing some impressive stratigraphic tangles, if he was correct. In 1887, Walcott found the fossil evidence to destroy the troublesome "Taconic System", allowing him to concentrate on the problem of series in the Cambrian. On his honeymoon in Newfoundland in 1888, Walcott found the evidence to support Brøgger's insistence that the trilobite *Olenellus* was older than *Paradoxides*.

Armed with this new biostratigraphic knowledge, Walcott moved through the southern Appalachians. In 1889 he had surmised the position of the Olenellus fauna on Chilhowee Mountain in Tennessee on the basis of a few obscure fossils. Various other chores kept him from more field work in the region until after the 1891 International Congress of Geologists in Washington, D.C. Once into the field, he was able to confirm the presence of *Olenellus* when he studied the section at Balcony Falls on the James River. Having found this key fossil, he was then able to move down the Appalachian chain identifying the base of the Cambrian and immediately he published his results (Walcott, 1892a).

A point to note is that these trilobites in what came to be the Antietam Sandstone were not museum quality specimens, but bits and pieces of disarticulated specimens. If this is difficult to visualize, simply think of Chesapeake Bay blue crabs: Walcott was looking for and identifying the equivalent of what remains after a crab feast.

By mid-August 1892 the political situation was no worse and the Washington climate no better. Walcott went into the York Valley and one result of this trip was *Olenellus wanneri*, named for one superintendent of schools in York, Pennsylvania. Arthur Keith of the Geological Survey joined Walcott and Wanner and they moved to the Susquehanna and then back to Emigsville. There Walcott "Discovered Olennelus fauna beneath limestone. Also trace to New Holland from Emigsville & also found it above the Hellam quartzite." This certainly had bearing on the geology of the Blue Ridge and the age of the limestone immediately above what were now the basal sandstones of the Cambrian. After a quick trip to Chiques Rock - the Chickies of today - they headed south.

"With Keith drove across South Mountain <u>via</u> Monteray Station W. M. R.R. & thence north along west of Mtn. & at night to Smithsburg. We found the Olenellus fauna in quartzite & limestones at the western side of the mountain and to 4 <u>mi</u> west of Waynesboro." From here they recrossed the mountains and doubled back, gradually working south and finding *Olenellus* fragments where they now anticipated them. They eventually worked down to Harper's Ferry and before the month was out, Walcott was back in Washington. He had cracked the structure of the Blue Ridge (Walcott, 1892b). Walcott also prepared a more popular account of his work and that of others along South Mountain (Walcott, 1893).

Subsequently, Walcott pursued the Cambrian northward to the Delaware River (Walcott, 1894a) and later into New Jersey (Walcott, 1894b). Every good investigation leads to unexpected results. The fragments in the quartzite at Mount Holly Gap led to the York Valley. This in turn led to the then novel notion of intraformational conglomerates (Walcott, 1894c). The original find which helped unravel the structure of the Blue Ridge and the subsequent discoveries in Pennsylvania were then summarized and published in that noteworthy but seldom consulted series, a U. S. Geological Survey Bulletin (Walcott, 1896). If Walcott had not been so concerned about his wife and small son slowly stewing in the Washington heat of 1892 and if he had been richer and able to afford a more luxurious locale for them to visit, our understanding of South Mountain might have been delayed for years.

GEOLOGY OF THE BLUE RIDGE MOUNTAINS, PENNSYLVANIA.

by
Samuel I. Root, The College of Wooster, Wooster, OH and R. C. Smith, II, Pennsylvania Geological Survey, Harrisburg, PA

INTRODUCTION

The Blue Ridge Mountains extend as a prominent morpho-tectonic belt from Alabama to the Susquehanna River. Geologically it is composed of Precambrian crystalline rocks overlain by latest Proterozoic(?) to earliest Cambrian siliciclastics, which are dominantly coarse grained. Topographically, it rises above the Mesozoic rocks of the Gettysburg-Newark Lowlands that border it on the east along much of its length in Pennsylvania. Where the Mesozoic rocks are absent, as to the south, it is bordered by the lowlands of the Piedmont. The Blue Ridge also rises above the less resistant Cambro-Ordovician carbonates and shales underlying the Great Valley that forms its western border. Structurally, the Gettysburg-Newark Lowlands represent a much younger, Early Jurassic rifting and deformation in which Mesozoic strata are tilted to the northwest at 15^{o}-25^{o}. Intrusions of diabase, both as large sheets and extensive dikes and small flows, accompanied this regional tilting. The Piedmont is a complex terrane of multiply deformed metamorphic and plutonic rocks of Precambrian and lower Paleozoic age. Both the Mesozoic and Piedmont rocks are separated from the Blue Ridge by a major fault which forms the eastern margin of the Blue Ridge. The rocks of the Great Valley, which are conformable upon the Blue Ridge siliciclastics, were deformed together with the Blue Ridge during the terminal Paleozoic Alleghanian deformation and are thus related in style and process (Figure 5).

The oldest rocks exposed in the core of the Blue Ridge occur in Virginia and to the south. Here, structurally complex Grenville-age (±~1Ga), high-grade metamorphic rocks such as layered granulite gneiss are intruded by crystalline plutons including charnockites. Large scale ductile deformation locally imposes a mylonitic fabric on these rocks. Overlying the Grenville basement in places is a thin sequence of metagraywackes, meta-arkoses, and phyllites referred to as the Swift Run Formation, which may be the equivalent of the Faquier Formation to the east. These sedimentary rocks are in turn are overlain by a thick sequence of Late Proterozoic volcanics of the Catoctin Formation. Overlying these volcanics is a thick succession of latest Proterozoic(?) to Early Cambrian coarse siliciclastics of the Chilhowee Group. Only the Catoctin and Chihowee units are known in any substantial amounts within the Blue Ridge from the Potomac north to the Susquehanna River. This paper is a review of the pertinent literature and ideas on the bedrock geology of the northern Blue Ridge.

REGIONAL GEOLOGY

From the Potomac River northward, the Blue Ridge may be

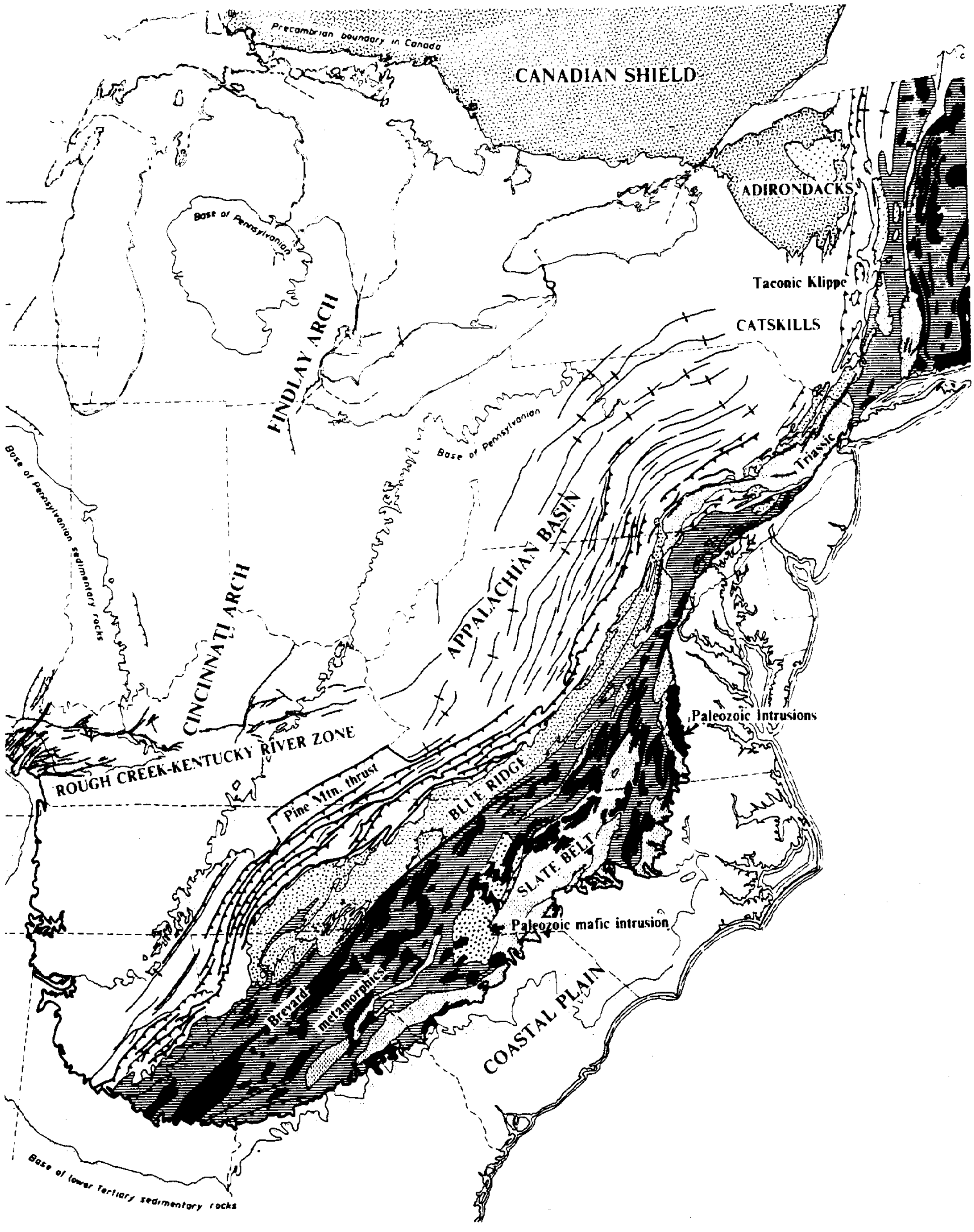


Figure 5. Tectonic sketch map of the southern and central parts of the Appalachian Orogen. In the Blue Ridge the stippled pattern indicates Grenville age crystalline rocks and the dotted pattern on the flanks represents younger Precambrian Catoctin volcanics and Cambrian (?) quartzites and graywackes (from Spencer and others, 1989, Figure 1-IA, p. T157: 2).

considered structurally as a northeast-trending anticlinorium cored by Catoctin volcanics. The limbs are defined by the Chilhowee siliciclastics which form a moderately southeast dipping upright but incomplete limb on the east and a vertical to overturned limb on the west. Regionally, the anticlinorium plunges to the northeast and disappears near Dillsburg, both by plunging out of the Chilhowee strata as well as being overridden by the Late Alleghanian Yellow Breeches thrust sheet (Figure 6) which postdates folding of the Blue Ridge. From Shippensburg, south to a few miles into Maryland, the structure reverses plunge to the southwest. Major deformation of the Blue Ridge is considered to be an Early Alleghanian event (terminal Paleozoic). The unambiguous Taconic deformation (Middle to Upper Ordovician) that affects a considerable area elsewhere in eastern Pennsylvania has not been recognized.

In Pennsylvania, the eastern flank of the Blue Ridge is cut out by a large displacement, Jurassic-age normal fault that juxtaposes Mesozoic-age northwest dipping Mesozoic strata of the Gettysburg Basin against the Blue Ridge anticlinorium. The western flank of the Blue Ridge passes into Cambro-Ordovician carbonate rocks without a structural break. These rocks have been deformed together and are structurally related. It is their differing resistance to weathering and erosion that causes a geomorphic distinction into the Blue Ridge and Great Valley physiographic regions.

Rocks referable to the Catoctin and Chilhowee, such as the Pigeon Hills and Accomac volcanics and the Chickies and Hellam siliciclastics, occur to the east in some of the complex fault slices in Adams and York Counties and extend across the Susquehanna River to Chickies Rock in Lancaster County. These are not part of the Blue Ridge morpho-tectonic feature, but are clearly related in terms of genesis and composition and ultimately must be considered in any geologic synthesis of these rocks. It has been suggested (D.B. MacLachlan, personal communication, 1990) that these units may have accumulated in a second rift arm that extended northeast from the major rift in which volcanics and siliciclastics of the proto-Blue Ridge accumulated. These Late Precambrian rifts may have established the site of the later Mesozoic rift basins (Root, 1989).

Catoctin Formation

The lithology of the rocks in the Blue Ridge and related Great Valley are described briefly in the stratigraphic column (Table 2). The thickness of the Catoctin Formation in Pennsylvania is not known because the base has not yet been recognized in the state. Any thickness calculated must account for internal distortion produced by flowage and development of a tectonite fabric in the volcanics as well as external distortion resulting from unrecognized folding or faulting. In Pennsylvania, the Catoctin volcanics are complex, but perhaps can be generalized as a series of metamorphosed rhyolites (metarhyolites), phyllites (tectonites?), and a metamorphosed basalt (metabasalt) unit (Table 3). The presence of metarhyolites in Pennsylvania is un-

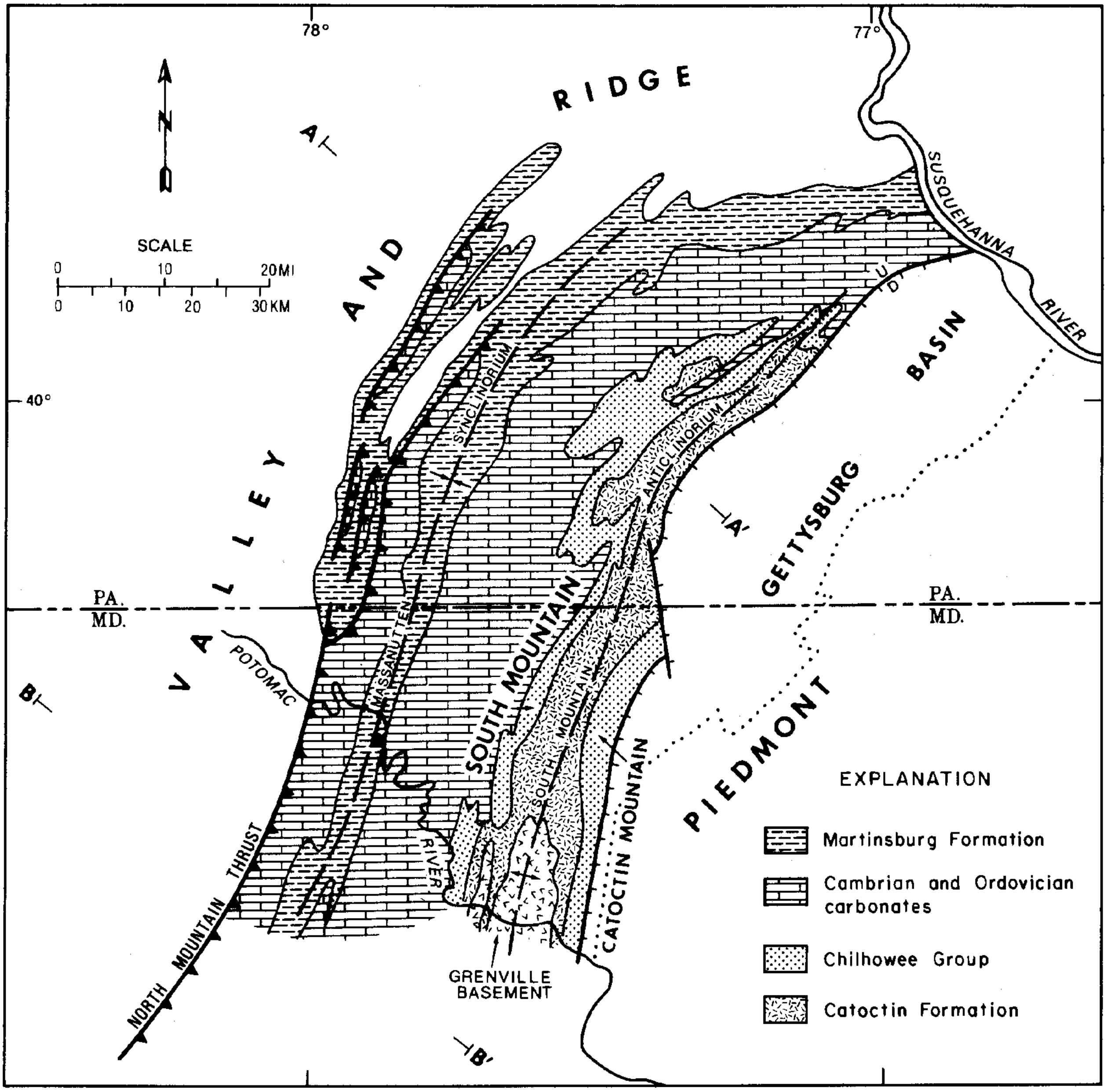


Figure 6. Generalized geologic map Blue Ridge, Potomac River to Susquehanna River.

usual because the Catoctin is principally a metabasalt in northern Virginia and Maryland. Fauth (1978) calculated a Catoctin thickness of >2,500 feet (>750 m) on the eastern flank of the Blue Ridge. Anecdotal evidence suggests that more than 8,000 feet (>2,500 m) of Catoctin volcanics have been penetrated in drill holes in southern Pennsylvania. If this figure is reliable it may indicate structural repetition or steep dips. To the south, near the Potomac River, Nickelsen (1956) calculated a maximum thickness of 1,000 feet (300 m), however, in places the Catoctin was absent, suggesting either extremely irregular accumulation or erosion prior to deposition of the Chilhowee siliciclastics.

Table 2. Geologic section, Blue Ridge, Pennsylvania (from Fauth, 1978).

System	Series	Group	Formation	Thickness (in feet)	Description
Cambrian	Middle Cambrian		Waynesboro Formation	1,000	Thin buff basal and upper red ridge-forming sandy units. Middle portion is blue limestone.
	Lower Cambrian		Tomstown Formation	1,350	Dolomitic limestone in upper part. Mottled silty dolomite in middle part. No exposures at base.
		Chilhowee Group	Antietam Formation	700	White, light-gray, and yellowish-gray, medium-grained protoquartzite and quartzite; locally friable sandstone. Contains *Skolithus* tubes.
	Lower(?) Cambrian		Harpers Formation	2,500 +	Olive fine-grained graywacke with dark bands and laminations, dark phyllites, and massive gray quartzite (Montalto Member).
			Weverton Formation	900 +	Gray and purplish-gray, laminated and crossbedded, quartzose graywacke. Some gray quartzites and purplish phyllites. Conglomeratic at base.
			Loudoun Formation	200	Blue, purple, and dark-gray phyllite intercalated with laminated, very fine grained graywacke. Polymictic conglomerate at base locally.
Precambrian	Upper Precambrian		Catoctin Formation	2,500 +	Altered green basalt containing chlorite and epidote; commonly amygdaloidal, rarely porphyritic. Altered blue, gray, and red porphyritic rhyolite; locally banded, finely laminated, or dense. Some green, yellowish-gray, and dark-gray, blebby and fragmental phyllites.

Table 3. Lithologic units in the Catoctin Formation (from Fauth, 1978).

Fauth (1977) Catoctin Furnace quadrangle area, Maryland[1]	*Fauth (this report)* Iron Springs quadrangle area, Pennsylvania[1]	*Fauth (1968)* Caledonia Park quadrangle area, Pennsylvania[1]	*Freedman (1967)* Mt. Holly Springs quadrangle, Pennsylvania[1]
Porphyritic metabasalt Metabasalt	Metabasalt	Metabasalt Epidosite Greenstone	Metabasalt
Gray phyllite	Green phyllite Gray phyllite	Phyllite	Phyllite members, Loudoun Formation (in part)?
Red porphyritic metarhyolite	Red porphyritic metarhyolite	Red porphyritic metarhyolite	Lavender metarhyolite
Blue metarhyolite	Blue metarhyolite Orange porphyritic metarhyolite	Bluish metarhyolite Metarhyolite breccia	? Aporhyolite porphyry Aporhyolite
	Mottled metarhyolite	Mottled metarhyolite	
	Vitreous metarhyolite Hematitic metarhyolite	No equivalents recognized	

Age of the Catoctin volcanics

Badger and Sinha (1988) obtained a Latest Precambrian age of 570±30 Ma for Catoctin volcanics based on analyses of metabasalts from the west limb of the Blue Ridge near Waynesboro, Virginia. For their $^{87}Sr/^{86}Sr$-$^{87}Rb/^{86}Sr$ isocron they chose the least altered portions of flows (containing 2.7±0.3% TiO_2) and a clinopyroxene separate. This same isocron yielded an initial $^{87}Sr/^{86}Sr$ ratio of 0.7035 which for a 570 Ma suite suggests little if any crustal contamination of a magma derived from a depleted mantle.

Based on U-Pb isotopic analyses of optically sorted zircons from three metarhyolite samples of Catoctin metarhyolite in Pennsylvania, Aleinikoff and others (1991) report an age of 597±18 Ma, confirming the latest Precambrian age of Badger and Sinha. Aleinikoff and others (1991) also distinguished an older, 759±2 Ma volcanic event at Mount Roders in southwestern Virginia from the Catoctin event in the Blue Ridge of northern Virginia and South Mountain, Pennsylvania. Badger and Sinha (1991) further extend the younger Catoctin-related event northward to mafic dikes in eastern New York, the Tibbit Hill Volcanics in northern Vermont and southern Quebec, and even to the Hughes Lake igneous complex in Newfoundland.

Chilhowee Group

Overlying the Catoctin is the Chilhowee Group. The contact between the two sequences represents a change from extrusive igneous activity to sedimentation. Of interest is the amount of deformation that occurred between cessation of volcanism and dedeposition of siliciclastics. In Pennsylvania, it has been generally considered to be slight, although there is little direct evidence. The Chilhowee Group consists of four formations (Table 2). In ascending order they are the Loudoun, Weverton, Harpers, and Antietam Formations. Root (1970) suggested that in Pennsylvania, these formations aggregate to a thickness of about 5,200 feet (1,600 m). They are composed mainly of quartzites and metagraywackes with subordinate conglomeratic and phyllitic units. The age of the Chilhowee is still a subject of debate and is discussed by Key (this volume). Only the uppermost formation, the Antietam, contains fossils; a trilobite fauna representative of the base of the Cambrian. The thick sequence of strata below the Antietam may then still be of earliest Cambrian age or they may be as old as latest Precambrian (Proterozoic) age. They are usually assigned the designation of the Lower (?) Cambrian Series. The Antietam is overlain conformably by carbonates and shale of the Tomstown Formation which mark the beginning of deposition of a thick sequence of Cambro-Ordovician shallow water marine carbonates that attain a thickness of about 14,000 feet (4250 m) (Root, 1970).

PREVIOUS STUDIES

Much of the foundation for contemporary understanding of Blue Ridge geology in Pennsylvania was established by the mapping of G.W. Stose. This pioneer did much of his work in the early part of this century although some of his material was not published until much later (Stose, 1909, 1932, 1953; Stose and Bascom, 1929). Since his early mapping, a number of quadrangles in the Blue Ridge have been remapped in detail (Fauth, 1968, 1978; Freedman, 1967; Root, 1968, 1978).

A number of topical papers have been published over the past 45 years that examine the geology of the Blue Ridge, principally the structure and volcanics. The classic paper by Ernst Cloos (1947) on oolite deformation in the Blue Ridge established the deformation style in this region and demonstrated the utility of petrofabrics in understanding structural processes in this area; particularly the use of deformed oolites in the Cambro-Ordovician carbonate rocks, chlorite blebs and amygdules in the volcanic rocks, and stretched pebbles in the siliciclastic rocks as strain markers as well as transport indicators. His students continued this type of research, mostly south of Pennsylvania, and the work of Nickelsen (1956) and Whitaker (1955) are major contributions to this body of knowledge. Papers by Root (1970) and Root and Hoskins (1978) discuss faulting within the Pennsylvania segment of the Blue Ridge and a recent paper by Root (1989) discusses basement control of structure in the Blue Ridge and particularly at the faulted western margin of the Mesozoic Gettysburg Basin. The stratigraphy and petrology of the siliciclastics and volcanics are described in the various state and federal survey quadrangles and earlier reports cited previously. A paper by Rankin (1975) discusses the regional aspects of the Catoctin volcanics and relates them to the Late Proterozoic evolution of the eastern margin of North America. The Geological Society of America publication Volume F-2 on The Appalachian-Ouachita Orogen in the United States (Hatcher and others, 1989) is a fine summary of recent work in the Blue Ridge.

STRUCTURE OF THE BLUE RIDGE

Folding

In Pennsylvania the Blue Ridge is dominated by folding. To the south, commencing in Virginia, the Blue Ridge is involved in large scale, low angle thrusting (Figure 5) so that the contact between the Great Valley and Blue Ridge is faulted; the northwest limb is cut out by the thrusts on which displacements ranging from a few tens of kilometers to 125 km have been calculated from surface mapping, but on which displacements of perhaps 260 km are suggested from seismic reflection profiles (Hatcher and others, 1989).

Displacement diminishes on the thrusts as the Blue Ridge is traced to the north from Virginia; faults are absent near the Potomac. Here, shortening displacement is transferred from faulting to folding and at the Potomac River the Blue Ridge is an anticlinorium, significantly overturned, verging to the northwest with axial planar cleavage dipping about 22^{o} southeast (Nickelsen, 1956). The anticlinorial structure is well expressed by the geomorphology of the region. In Maryland and northern Virginia, the volcanic core forms an area of rolling hills, whereas the resistant Chilhowee siliciclastics form bounding mountain ridges; Catoctin Mountain on the east is the upright east-dipping fold limb, which is faulted against the Mesozoic Gettysburg and Culpeper Basins. South Mountain, nine miles (14 km) to the west, is the markedly overturned west-facing limb which passes conformably into the overlying Tomstown Formation of the Great Valley. This relatively simple pattern persists to the Pennsylvania border (Figure 6).

In Pennsylvania, the Great Valley and Blue Ridge are dominated by two major folds: the South Mountain anticlinorium whose major axis lies along the volcanic core of the Blue Ridge and is the structural extension of the anticlinorium that begins in northern Virginia, and the complementary Massanutten synclinorium on the west, whose major axis passes through the Middle and Upper Ordovician Martinsburg shale belt in the center of the Great Valley and extends from Massanutten Mountain, Virginia, northeast into Cumberland County, Pennsylvania, and possibly beyond to the anthracite region. In Pennsylvania, the eastern flank of the Blue Ridge is truncated by the large normal fault, parallel to the anticlinorium trend, that bounds the Mesozoic Gettysburg Basin cutting out the ridge of Chilhowee strata that define this upright flank so well in Maryland as Catoctin Mountain. Thus, north of the Jacks Mountain belt of Fauth (1978, Figure 2) most of this east flank consists of Catoctin volcanic rocks juxtaposed against Mesozoic beds. The west flank of the fold, as expressed by the Chihowee Group ridge (equivalent to the South Mountain ridge in Maryland) is present along most of anticlinorium. It is transected in a number of places by faults at large angles to the anticlinorium trend, which only offset the ridges.

Although the geometry of the Blue Ridge folds is readily described, the processes of formation are far less apparent. The geometry is summed up by Cloos (1947, p. 845) for the Virginia-Pennsylvania area as follows:

> "The South Mountain fold is a large asymmetrical overturned anticline. Its axial plane dips to the southeast, and its crest is the western slope of South Mountain. Cleavage dips steeper on the upper than lower limb thus forming a fan which opens to the northwest. All parts of the fold participate and reveal an identical deformation plan: fold axes are nearly horizontal, cleavage dips southeast, lineation is in the cleavage plane normal to the fold axes, also dipping east. All formations including volcanics participate. . . .
>
> Intensity of deformation varies greatly within the fold depending on (1) physical properties of materials, (2) location within the fold, and (3) geographical location. . .
>
> Approaching South Mountain from the west, intensity grows gradually, is strongest in the lower limb, decreases toward the crest and upper limb. Abrupt changes as would be expected in large-scale thrusting were not observed, the deformation seems to be absorbed within a much wider complex. . . .
>
> The fold is interpreted as a large "shear" fold as distinct from flexures. Deformation is thought to be due to laminar flow on subparallel planes. The presence of flow planes and the fact that ooids are extended at large angles to bedding show that stratigraphic thicknesses as now seen are not equivalent to thickness deposition."

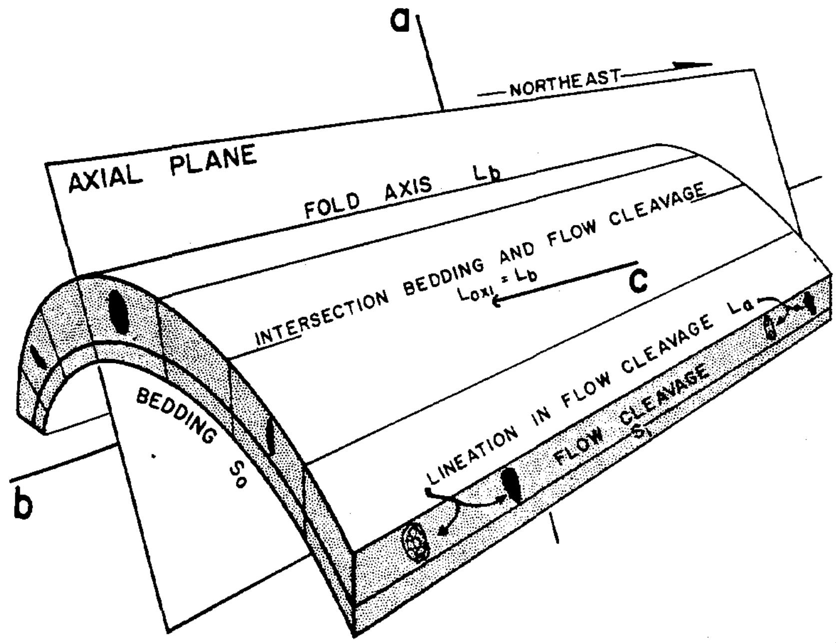
It is important to note that in Pennsylvania the west limb of the anticlinorium, as defined by the Antietam quartzite, is generally subvertical rather than overturned at a large angle as in Maryland and Virginia. Cleavage is commensurately steeper. The description of the fold in Pennsylvania should be modified as a grossly asymmetric but only slightly overturned structure, with a considerable component of plunge; regionally to the northeast north of the Shippensburg fault and to the southwest south of this fault. The relationship among the various structural elements of the South Mountain deformation plan relative to a tectonic coordinate system of three mutually orthogonal axes a, b, and c are shown in Figure 7. In addition to the principal South Mountain cleavage, a minor, later slip or crenulation cleavage is imposed on this fabric; particularly where the rocks are fine grained. The significance of this will be discussed later.

According to Ramsay and Huber (1983) cleavage is related to the finite strain state. The fabric forms perpendicular to the shortest axis (Z-direction) of the finite strain ellipsoid and increases in intensity with the strain ratio. Cleavage represents the traces of the XY planes of adjacent strain ellipsoids. This would correspond to the ac plane of Cloos shown in Figure 7. They cite the early work of Sorby who suggested that cleavage formation is the result of the following processes:

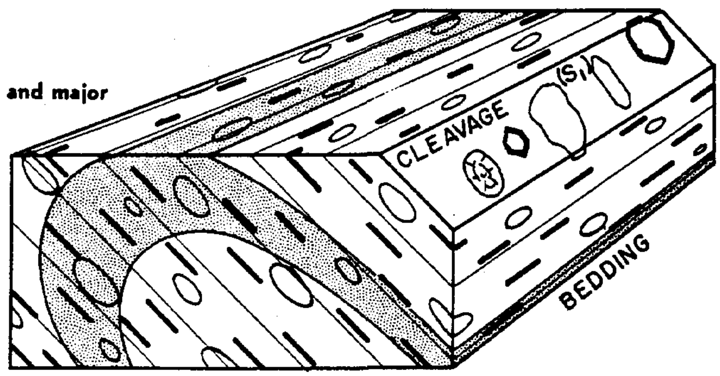
1. Mechanical reorientation of initial platy and acicular minerals.
2. Mechanical reorientation of minerals crystallizing during metamorphism.
3. Preferred growth of new minerals in directions controlled by stresses at time of growth or by deformational anisotropy already present at time of growth.
4. Plastic flow of individual crystals inducing shape anisotropy related to principal strain directions.
5. Change of shape of crystals by pressure solution and redeposition.

Beyond this, Cloos (1947, 1971) considered the cleavage in the Catoctin to have formed when at least parts of the volcanics were in a plastic state and that there is displacement or shearing on the cleavage surfaces. Furthermore, he suggested (1971) that the minor, later slip low angle cleavage in the South Mountain fold is the result of post-cleavage external rotation from an initially more upright orientation. This must be considered if the cleavage is attributed to a subhorizontal layer parallel shortening, not a vertical flattening. Many details on the South Mountain cleavage await resolution.

In some areas of the Catoctin volcanics and especially the pelitic portions of the Harpers and Loudoun Formations, cleavage is pervasive and developed at the millimeter and sub-millimeter scale. It obliterates or renders obscure primary structures in many areas and imposes the tectonite fabric of the South Mountain deformation plan on almost all elements of the rock. Exceptions in the metabasalts are discussed by Smith and others elsewhere in this volume. The quartzites and graywackes are more resistant to cleavage development and cleavage development is less prominent, although in thin section mineral grains usually show effects of strain and cataclasis. Quartz is typically elongated in



Schematic diagram showing the tectonic coordinates and major structural elements in an anticline.



LINEATION IN PLANE OF FLOW CLEAVAGE
MICAS (DASHES), OOLITES, AND ROCK FRAGMENTS (ELLIPSES) ORIENTED DOWN THE DIP OF CLEAVAGE PLANES

Figure 7. South Mountain deformation plan (from Fauth, 1968).

the a and shortened in the c direction (Nickelsen, 1956; Fauth, 1968). The volcanics behaved as a ductile layer (passive to flexure flow mechanics) during deformation, whereas the more competent Chilhowee Group acted as a relatively stiff layer (flexural slip to flexure flow mechanics) during deformation. Folds as expressed by plunging Chilhowee strata show geometry of chevron folds; the southwest plunging Grier Hollow anticline in the Caledonia Park area (Fauth, 1968) is a box fold as are some of the smaller northeast plunging anticlines at the northern terminus of the Blue Ridge (Friedman, 1967; Root, 1978). Such fold geometry is associated with dominantly stiff layer behavior.

On a regional scale, the South Mountain Anticlinorium is a fault bend fold (Figure 8) developed above the ramp segment of a regionally subhorizontal blind thrust which is postulated to be in the Ordovician Martinsburg shales by Kulander and Dean (1986), but considered to be within the Lower Cambrian shales by Woodward (1985) (Figure 9). The anticlinorium is part of the major Massanutten-Blue Ridge thrust sheet which carries the Massanutten Synclinorium as well (Kulander and Dean, 1986). The thrust underlying this sheet emerges as the North Mountain fault on the west

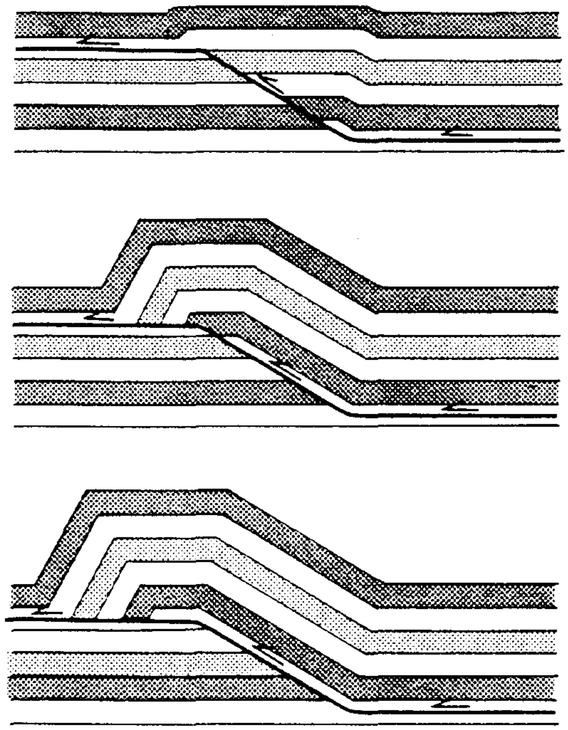


Figure 8. Development of fault-bend folds (from Spencer, 1988).

side of the Great Valley (Figure 6). South of Pennsylvania it is a single major fault, but in Pennsylvania it splits into a number of strands which form several major highly faulted anticlines cored by Cambro-Ordovican carbonates in Franklin and Fulton Counties. The thrust sheet is 27 miles (43 km) wide in Maryland where it is well defined. The eastern end of the sheet, however, is cut off by the Mesozoic normal fault bounding the Gettysburg Basin.

Faulting

Large scale, non-outcropping, subhorizontal thrusts dominate the regional structure of the region. Because deep wells have not been drilled or detailed seismic profiles released in this area, the position of these thrusts is speculative. It was mentioned that the Martinsburg shales probably serve as the decollement horizon underlying the Massanutten Blue Ridge thrust sheet (Kulander and Dean, 1986) (Figure 9) and, considering our present state of knowledge, this is a most reasonable interpretation. Also, Kulander and Dean (1986) suggest that the Blue Ridge has been displaced to the west a minimum of 40-45 miles (64-72 km) on the major thrust (Figure 9). However, the weak shales in the Lower Cambrian carbonates have traditionally been considered to be a major decollement horizon and a cross section by Shumaker in Woodward (1985) shows the thrust sheet to be transported on a decollement in the Lower Cambrian (Figure 9). Clearly, the Lower Cambrian shales serve as the decollement horizon on the east side of the South Mountain Anticlinorium. Considering mechanics of

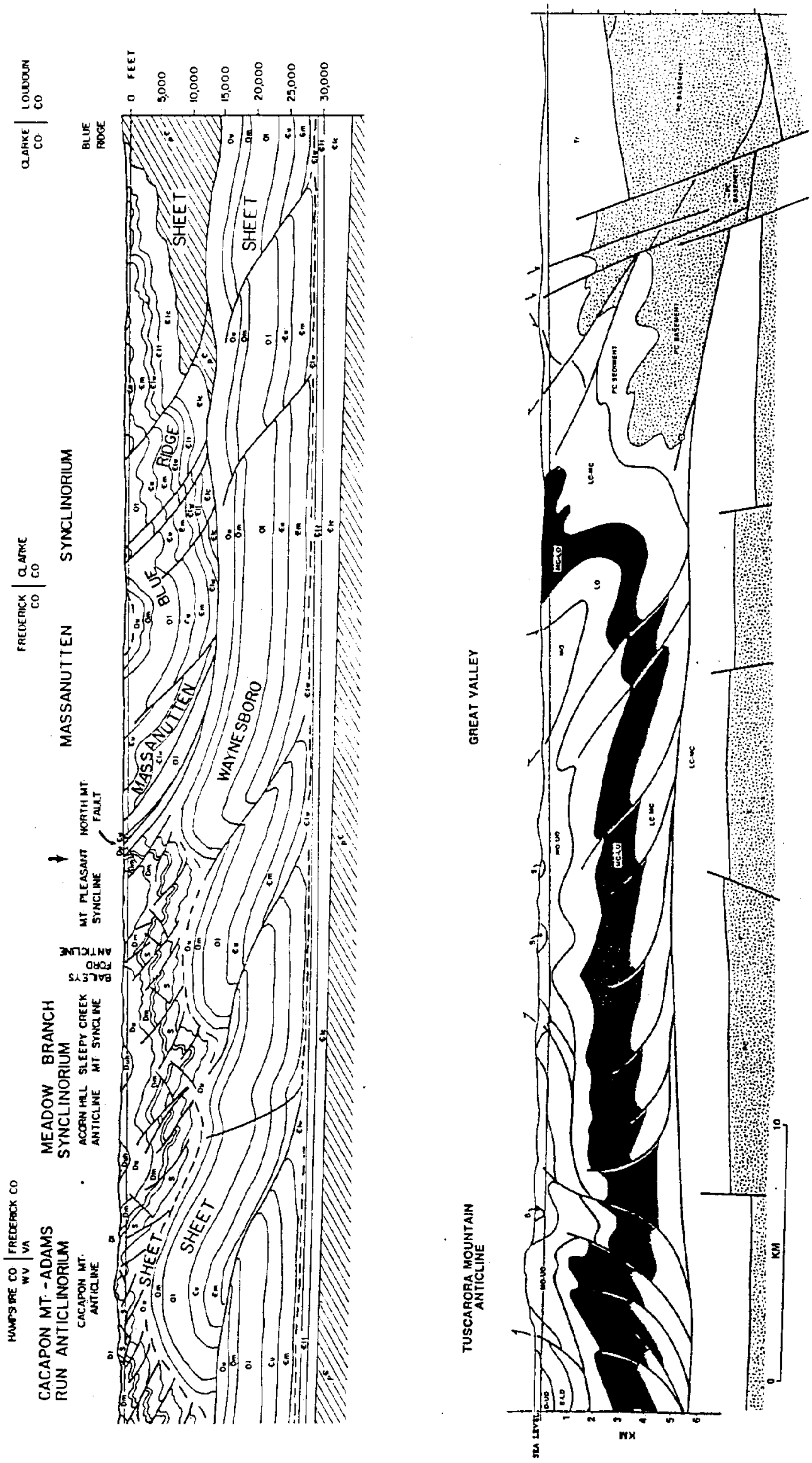


Figure 9. Regional geologic cross sections. Line of cross sections shown on Figure 6. A-A' from Woodward (1985) and B-B' from Kulander and Dean (1986).

fault-bend folding (Figure 8), the upright, east-dipping limb of Chilhowee Group strata in Maryland certainly must parallel a ramp fault cutting up from a lower decollement horizon. As traced into Pennsylvania, where the ridge of Chilhowee Group siliciclastics is cut out by faulting associated with the Lower Mesozoic rifting, this relation is less assured. The deeper decollement horizon could be either in the Lower Cambrian strata, in which case the older units are emplaced from a deeper thrust even farther to the east, or it could be a detachment deep in the Catoctin. The thrusts are considered to extend further east under the Piedmont. It must be noted that at these depths highly ductile shearing will prevail and layer anisotropy may not be as significant as it is in traditional fault-bend folds. The position of this eastern ramp undoubtedly served to locate the position of the later Mesozoic normal faults that border the Gettysburg Basin.

In terms of orientation, two classes of faults are recognized in the South Mountain anticlinorium: those parallel to the regional northeast-trending structural grain and a few faults oriented east-west. The east-west faults, which include the Carbaugh-Marsh Creek and Shippensburg faults (Figure 10), have been described by Root and Hoskins (1977) who believe that these faults are part of a major fracture (the Transylvania Fault) extending deep into the continental crust. This may be reflected in the composition of the volcanics which contain more metabasalt south of the fault than to the north. Within the Blue Ridge they observed that semi-independent shortening occurs in strata on either side of the Carbaugh-Marsh Creek Fault and lesser semi-independent shortening occurs across the Shippensburg Fault. From this they concluded that these faults predate Alleghanian deformation and are possibly of Taconic or, more likely, Late Proterozoic origin. The east-west faults were reactivated during the Alleghanian Orogeny and in places link with northeast-trending steep Alleghanian thrusts dipping 45-60° southeast. They suggested that basement north of these faults was structurally lower, which would account for the northeastern plunge of the South Mountain structure. During Mesozoic rifting these east-west faults were reactivated with development of the normal fault at the northwest margin of the Gettysburg Basin and are the site, in the basin, of both offset of the border fault and wrench structures (Root, 1989).

Those faults parallel to the South Mountain structural grain are of two differing types (Figure 10), although they both dip southeast. Faults in which older strata occur on the southeast side of the fault are Alleghanian age steep thrusts and related to development of the South Mountain fold. Where younger beds are on the southeast side of the fault and the faults show attributes of brittle deformation those faults are considered to be normal faults related to Mesozoic rifting in which younger beds have been downdropped to the southeast (Root, 1989). These younger normal faults are superimposed on the South Mountain fold and decrease in amount away from the margin of the rift basin.

Fauth (1978) notes the importance of the Tunnel Hill-Jacks Mountain fault system and its possible reactivation during the

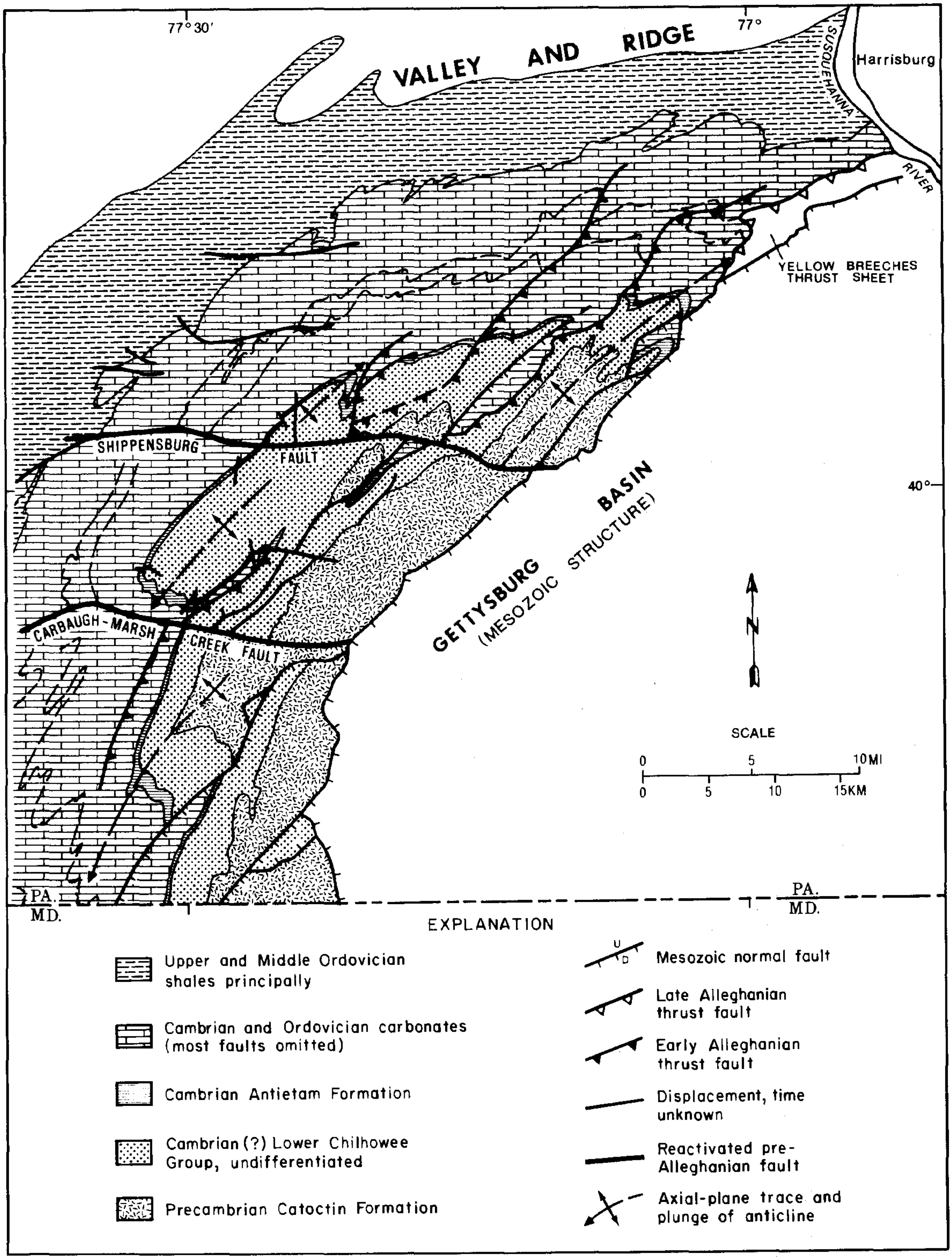


Figure 10. Generalized geologic map of the Blue Ridge area in Pennsylvania (after Berg and others, 1980).

Jurassic. It is herein suggested that the 20° east-dipping thrust observed near the west portal of the tunnel is a splay of a more intensely deformed and cribbed zone in the mid portions of the tunnel which, based on deep water-well data, continues southeast at about 30° and exposes overturned Weverton Formation near the east portal. These details, however, in no way discredit Fauth's recognition of fault reactivation.

Origin of the Blue Ridge

A number of individuals have considered (see discussions in Root, 1989; Hatcher and others, 1989) that parallelism of the structural grain of both the South Mountain fold and the Mesozoic rift basin derives from a fundamental structural grain generated during the Late Proterozoic development of the east margin of North America (Laurentia). This structural grain seems to be related to Iapetion rifting that preceeded the development of a passive continental margin bordering the Iapetus Ocean. Review, in Hatcher and others (1989), of the post-Grenville geology of the Blue Ridge suggests that the Catoctin and Chilhowee accumulated within a rift on the continental plate. This model has merit because the Catoctin is a continental tholeiitic basalt with restricted subaerial rhyolites of peralkaline affinity (Rankin, 1975) and the overlying siliciclastics have many rift-related attributes. Furthermore, South Mountain, which is at a major bend in the Appalachians, may represent a triple junction in which rhyolites are abundant and Chilhowee deposition is exceptionally thick. A failed arm of this triple junction, an aulacogen, that points to the continent might account for the east-west oriented Transylvania Fault (Root and Hoskins, 1978). Subsequently, the rift phase developed into a passive margin with miogeoclinal Cambrian and Ordovician strata upon the Chilhowee rift strata.

CHRONOLOGY OF DEFORMATION

Timing of the thrusting and folding that produced the Catoctin volcanics-cored South Mountain Anticlinorium can only be inferred. Regional stratigraphic relations in the central Appalachians indicate that the sequence from the Chilhowee to the Pennsylvanian is essentially conformable. Analysis at the basin scale indicates stratigraphic breaks are few and minor. The break between the Catoctin volcanism and sedimentary sequence is minor, at least in terms of structural discordance. Therefore, the sequence from Catoctin to the Pennsylvanian may be considered as an enormous westward-tapering prism that has been deformed together. This is well illustrated on the Massanutten-Blue Ridge thrust sheet where the core of the Massanutten Synclinorium in Virginia contains infolded Devonian strata with the same deformation plan as the South Mountain fold. These relations suggest that the deformation which produced the South Mountain fold was an Alleghanian event; the same event that folded strata in the Valley and Ridge as well as on the Appalachian Plateaus.

As observed earlier, the principal cleavage is associated with South Mountain deformation and is therefore an Alleghanian cleavage. It is deformed in places by a minor slip cleavage. This minor deformation is not indicative of a later orogeny, but rather, indicates that Alleghanian deformation was an extended event as has been confirmed by a number of workers. For example, Nickelsen (1979) demonstrated several distinct stages in the development of Alleghanian folds in the anthracite area; Geiser and Engelder (1983) recognized two non-coaxial phases in the Alleghanian orogeny based on joint orientation on the Appalachian foreland; and Root (1970) and Root and MacLachlan (1978) described Late Alleghanian emplacement of the planar, subhorizontal Yellow Breeches thrust sheet over the folded, plunging nose of the South Mountain fold (Figure 10). Most recently, the junior author and D. B. MacLachlan have observed a subhorizontal, brittle, extensional fabric that appears to cut both Alleghanian cleavage and isoclinal folding. This extensional fabric may be an element of Mesozoic deformation. Further study is warranted.

A TALE OF TWO MOUNTAINS - BOTH SOUTH

by
David B. MacLachlan
Pennsylvania Geologic Survey, Harrisburg, PA

INTRODUCTION

The Blue Ridge *physiographic* Province, an essentially continuous upland underlain by Precambrian and Lower Cambrian crystalline, volcanic, and siliciclastic rocks extending from northern Alabama to south central Pennsylvania, terminates on the Cumberland-York County line near Brandtsville by north-eastward plunge of the highest quartzite bed. The use of the term *morpho-tectonic* belt (if that hybrid bastard should not be squelched in its infancy) by Root and Smith in the preceding introduction is more suspect, and uncritical use of this concept may lead to generalizations which obscure significant structural features. The Blue Ridge is clearly divisible into two major segments, with a distinct boundary in southern Virginia, which face on the Southern Appalachian thrust belt and Central Appalachian fold belt respectively. Each segment has the same general tectonic style and deformation plan as the belt on which it faces, and it is, at best, problematical that they should be assigned to the same tectonic province. With the understanding that only the northern segment is directly involved, the provincial description given above by Root and Smith is an adequate summary of most of the relevant literature. The main structure commonly called South Mountain in Pennsylvania, a doubly plunging compound-complex anticlinal, is also adequately described in most particulars, but the structural analysis suffers somewhat from excess reliance on the assumption of tectonic continuity with structures to the south. A plausible case may be made that this fold forms a (minor) third structurally distinct segment of the Blue Ridge. Practitioners of regional tectonic synthesis have taken a different perspective as reflected in Figure 5 of this guidebook, which is derived from the 1969 American Association of Petroleum Geologists tectonic map, and perhaps more clearly by Hatcher and others, 1989, in their Appalachian tectonic map (Plate 1, dated 1990), although they differ only in minor detail in this area. They terminate the Blue Ridge in northern Maryland and extreme southern Pennsylvania against the Gettysburg basin. They include the main Pennsylvania South Mountain (Blue Ridge ?) segment in the Valley and Ridge Province. By this criterion, inclusion of the latter in an *anything-tectonic* belt is in doubt. The justification for including (PA) South Mountain in the Blue Ridge is precisely because it was originally defined as a physiographic not a tectonic province.

The use of the name South Mountain in several contexts more-or-less united by topographic continuity in this region inevitably produces confusion among students attempting to understand its structure, and, it is to be feared, sometimes to their teachers. The South Mountain *fold* of Maryland is in fact the direct continuation of the Blue Ridge structure of northern Virginia. As detailed Root and Smith above, this structure has been intensively studied by E. Cloos and his students. North of exposure of middle Precambrian crystalline rocks of the Blue Ridge core near the Potomac River this fold embraces three topographically distinct elements. Catoctin Mountain, the easternmost, is supported by Cambrian(?) to Lower Cambrian, predominantly quartzose, clastic rocks (Chilhowee Group) on the eastern limb of the fold. Though somewhat impacted by higher order folds, these Catoctin Mountain rocks generally dip moderately to gently eastward. A medial rolling upland of the axial zone is underlain by Late Precambrian Catoctin metavolcanics. Internal structures of this terrain are not fully resolved owing to substantial obliteration of primary structures by metamorphism and cleavage. The western limb of the fold is again marked by a Chilhowee supported ridge which is the only topographic feature in Maryland to properly bear the name South Mountain. This is a marked contrast to the broad, structurally and topographically complex terrain bearing the same name in Pennsylvania, but the nomenclatural differences are historic rather than geologic, and it would be fatuous to claim that either is more correct. As far as can be determined, the rocks of the Maryland ridge are generally overturned including the limited area of its extension into southern Pennsylvania. It provides tenuous topographic continuity between the between the general uplift of the northern Blue Ridge proper and the South Mountain folds of Pennsylvania which clearly lie somewhat northwest of the major trend. The Maryland South Mountain is, in fact, clearly structurally discontinuous with its eponymous Pennsylvania cousin. The interpretation and magnitude of that discontinuity is debatable, but it may well be greater than is superficially apparent.

The general description of the Blue Ridge structure of Maryland given by Root and Smith above is quite complete as consistent with the present state of knowledge. It invites comment only in three particulars that seem to bear on its relationship to Pennsylvania structures: the plunge of fold, its apparent northeastward extension, and the probability of thrust faulting on its western flank.

DISCUSSION

The South Mountain (MD) fold is described as north plunging, as is most conspicuous in the northward disappearance of the gneissic core. Limb convergence is actually slight (Figure 6) despite the relatively low dip of the eastern flank, and a low plunge is implied for the axial trace projected to the (eroded) crest in the Chilhowee. A steeper plunge might be deduced from the crystalline nose, but the discrepancy is resolved by an estimated 150 percent stratigraphic thickening of the Catoctin Formation from the Potomac to southern Pennsylvania (see Root and Smith) plus any (possibly considerable) allowance to be made for structural thickening and repetition of the relatively incompetent metavolcanics in the axial zone. I will subsequently argue that this low plunge persists northeastward without (genetically) significant interruption to at least western Lancaster County, and that, with respect to the regional structure of the Appalachian Valley and Ridge folds, it is more anomalous than the steeper reversed plunges terminating the (PA) South Mountain folds.

The strike of the (MD) Blue Ridge units is completely truncated by faults at the western margin of the Gettysburg basin, as highlighted by the tectonic maps cited above. Catoctin Mountain ends south of the Pennsylvania line. Its projected trend, however, lies 2-3 km northwest of the Pigeon Hills on the eastern side of the Gettysburg basin. This figure is commensurate with the lower range of contemporary estimates of Mesozoic extension of the basin, and I have no doubt both are essentially the same feature. The axis of the (MD) South Mountain fold is less precisely definable, but it apparently extends slightly into Pennsylvania. In my efforts to resolve stratigraphic and tectonic problems in the so-called northwestern "Piedmont" of Pennsylvania (North of the Chickies-Oregon thrust it is essentially identical stratigraphically and tectonically to the Lebanon Valley section of the Great Valley. One must not be overimpressed by the geologic significance of physiographic provinces.), I have concluded that this axis must be represented east of the basin in the Hellam Hills and Chickies anticline. The evidence of the structural alignment is suggestive, but the association of both with the Lower to Middle Cambrian carbonate bank margin is compelling. The Hellam-Chickies ridge is bounded on the north by an imbricated thrust zone with a minimum displacement of over a kilometer (more likely at least 5 km) if all movement is assigned to the Alleghanian. Differences in tectonic fabric in the upper and lower plates point to the possibility of much larger displacement. This leaves us with interesting situation that a substantial and possibly major thrust passes into the Gettysburg basin between the (MD) South Mountain fold (extended) and (PA) South Mountain. With this happy thought it is appropriate to proceed circumspectly to the point where South Mountain meets South Mountain.

The (MD) South Mountain ridge extends north-northeast into Pennsylvania with the local name Mt. Dunlop to the valley of Red Run, thence as Green Ridge nearly to Marsh Creek. The Antietam Cove fault is first mapped at the Maryland line by Root (1968) and Stose (1909) at the contact between the Waynesboro and Elbrook formations (about Lower - Middle Cambrian boundary). It undoubtedly extends southward at this horizon for a few miles, but any greater prolongation is speculative. For about 7 miles to the northeastward the Waynesboro Formation, structurally on the west flank of (MD) South Mountain and east of the fault, transects about 2 km of strata on the plunging nose at the southwestern termination of (PA) South Mountain to the west. In Antietam Cove the upper member of the Harpers Formation and younger rocks of the east block are pinched out against the fault, which continues north along the western flank of Green Ridge as the boundary between metarhyolites of (PA) South Mountain and the Montalto Member in the western limb of the Maryland structure. The northern end of Green Ridge lies about a mile north of the limit of modern detailed mapping (Fauth, 1978), and the 1981 State

Geologic Map compilation suggests that this area has not been critically re-examined since Stose (1932 - field completion probably by 1920). If this mapping is correct (Stose has a mixed record on faults), the Antietam Cove fault swings eastward along the southern slope of the Marsh Creek valley toward the Gettysburg basin. A much lower fault dip is implied than is figured in Fauth's cross-sections to the south. There is a fully fault-bounded outlying block of the Chilhowee Group right at the west margin of the Gettysburg basin which obscures the relation of the Antietam Cove fault to the basin margin. Root and Smith's Figure 10 ignores this complicating detail, but their interpretation that the primary fault trace bends northeastward to merge with the Mesozoic border fault is congruent with my intuition. In keeping with the increasing body of data that links many extensional faults bounding Lower Mesozoic basins in eastern North America to pre-existing Alleghanian thrusts, Faill and MacLachlan (1989) projected the Chickies thrust along the Alleghanian grain to merge with the west margin of the basin tangent to the defection about 6 miles northeast of Marsh Creek. By this reasoning, a direct connection between the Antietam Cove fault and the Chickies thrust is implied.

Published accounts of the Antietam Cove fault are conflicting. None, oddly, seems to make much recognition of the Alleghanian thrusts parallel to it in this area, as mapped by Fauth (1968; 1978) and Root (1968; 1971; 1978), which are more abundant in the adjacent carbonate valley but also appear in South Mountain. Root (1968) recognized problems interpreting this fault which he resolved by postulating a sinistral strike-slip fault parallel to the regional tectonic grain. His preference in this matter seems informed by his apparently unaltered conviction that the Pennsylvania Blue Ridge (?) is identical with that in Maryland as proclaimed by Cloos (1951). [Note his axial trace of the South Mountain anticlinorium in Figure 6]. There is in fact no way that the Maryland axis can cross the steeply dipping (overturned) beds of the linear ridge extending from Marsh Creek well into Maryland. Root's (1968) model allows slip consistent alignment of the strike of contacts on the west flanks of the (PA) and (MD) South Mountain folds from the top of the Waynesboro to the base of the Antietam, but apparently fails to accommodate the substantial differences in dip shown by his structural data. A vertical component must, in any case, be introduced to consistently explain the long segment at an approximately constant stratigraphic level north of Antietam Cove (Fauth, 1978). The rather abrupt, though not severe, change of trend there militates somewhat against significant strike-slip. In the face of comling field eveidence, a standard more severe than I perceive as met here, a local anomaly must weigh more heavily than any perceptions deriving from regional structure. I simply note that to the best of my knowledge, as an Alleghanian strike-slip structure parallel to the tectonic grain, the Antietam Cove fault is absolutely unique in the entire folded Appalachians.

Fauth's (1978) text simply cites Root without comment or explication. His cross-sections, however, show the Antietam Cove as a normal fault (complete with displacement arrows) dipping steeply eastward. In the absence of the contrary assertion in the text, one would assume that his sections as drawn show only Lower Mesozoic extension. I accept this movement as implicit in acknowledging Root's linkage of the northeast end of the fault with the basin margin. Indeed, I can see no other explanation for the block supporting The Knob, which projects west of the main fault trend. Fauth's sections are unsatisfactory in some particulars near the fault. The fault dip is much too steep to be consistent the dip implied near Marsh Creek. A value comparable to that he provides for the putative Mesozoic extension fault at Tunnel Hill (about 20°) would be more appropriate. R. C. Smith has recently made extensive investigations of the basalts in Fauth's area. In the process he became convinced of fairly pervasive overturning east of the Antietam Cove fault despite reversals attributable to higher order folds. In two long field days he gave me a sampling of his evidence. After reviewing Fauth's orientation data I conclude Smith is correct. A major overturned limb is inconsistent with Fauth's sections and genetically incompatible with simple extensional faulting if the sections are essentially correct west of the fault, as they appear to be. If the fault has an Alleghanian thrust precursor, which is increasingly recognized as a normal case - especially where the dip of the extension fault is moderate, a major overturned limb confined to the upper plate is easily explicable with significant initial shortening. I surmise that Fauth forced his data somewhat to preserve essential continuity between the two South Mountains, but his sections are ultimately more credible than Root's.

In this guidebook Root and Smith take Fauth's cross-sections at face value and abandon Root's former model. They map (Figure 10) the Antietam Cove fault from Marsh Creek to the Maryland line as simple Mesozoic extension. Any clear evidence supporting this interpretation is wanting from Antietam Cove southward, and I conclude that the structural relations Root mapped across the fault cannot be so generated. While I am not fully aware of the perplexities that initially led Root to reject a thrust model, with concurrent folding it appears to me to be sufficient. Subsequent extension in this area cannot be excluded, it merely tends to diminish the original disjuncture between the two blocks, but it is not required here.

The western margin of the northern Blue Ridge segment in Virginia is marked by a series of thrust faults which have been recognized in Maryland approximately as far north as the gneissic core. These are generally interpreted as relatively steep splays from a deep decollement. It should be apparent that I strongly suspect one such fault projecting into northern Maryland from Pennsylvania. I would add to this another mapped by Root (1968) in the southern Chambersburg quadrangle between the Antietam Cove fault and (MD) South Mountain, which probably merges northward with it in the colluvially covered subcrop of the Waynesboro in Antietam Cove. If the Antietam Cove fault dies out southward along the Waynesboro-Elbrook contact, the other may well pick up the displacement. At the level of detail apparent in the 1968 Geologic Map of Maryland it appears feasible to link the northern and southern occurrences. Mindful of the concise field description of the Tomstown Formation as the concealed interval between the Waynesboro and the basal Cambrian clastics, it is unremarkable that such faulting has not been mapped, as it would extend in large part through that interval. A different kind of evidence to suggest faulting is found in the discontinuity of the Antietam Formation along the west limb in Maryland. This tough quartzite of fairly uniform thickness is elsewhere along the western flank of both the Maryland and Pennsylvania folds everywhere that faulting is not established. (The finer grained, argillaceous arenites of the east flank the Blue Ridge in Maryland and its York County extension are not always so persistent.) These are obviously less than compelling arguments for continuity of thrusting along the base of the Blue Ridge. They are significant to me because I am evolving hypotheses anent the Upper Precambrian evolution of the Blue Ridge, which I lack the space and energy to explicate here. These imply more subsequent displacement north of central Maryland than the minimum required to accomodate the observed disjuncture between the (PA) and (MD) folds. The latter could amount to no more than thrusting the (MD) sufficiently westward to suppress an intervening syncline.

The approach taken by the regional tectonic maps above cited to provincially separate (PA) South Mountain from the Blue Ridge is not without merit for emphasizing some structural features. As Root (1970; 1978) has previously asserted the essential deformational continuity of the folded Appalachians from (PA) South Mountain to the west and he affirms that view in his final paragraphs in this guidebook, I think I do him no injustice to approach (PA) South Mountain from the west. We concur that the progressive changes in metamorphism, cleavage aspect, and other features of the meso- and micro-structure are only those appropriate to the exposed stratigraphic level and the former thickness of cover implied. More frequent exposure of thrust faults east of First Mountain, as are generally considered essential to the structure to the west but are presumed to be frequently blind, may also have the same cause. Neither of us, however, I presume, considers the primary basal decollement presently shallower to the east, although repetition of section in mega-duplexes may provide substantially higher surfaces of near horizontal transport. The most apparently anomalous feature of (PA) South Mountain as gleaned from cursory inspection of the 1980 State Geologic Map is that it appears as a curved brachyanticlinal that seems conspicuously less elongate than typical Ridge and Valley folds. If one notes, however, that it lies athwart the axial culmination developed radially to the maximum curvature of the Pennsylvania reentrant, it becomes apparent that one may draw a linear traverse northward from South Mountain to the toe of the Allegheny Front crossing no rocks younger than a few narrow tracts of Lower Middle Devonian. About 4.8 km of strata on either side of this traverse, missing on this transverse arch, may be observed when ascending stratigraphically to the intermontain coal basins. This transversed structural relief is at least comparable to that apparent along the (PA) South Mountain axis. Transverse arching in the later may be reasonably supposed to be somewhat more laterally compressed owing to its interior position in on a smaller radius of the Appalachian curvature. In short, the (PA) South Mountain folds are exactly where and what they ought to be if considered as a component of the Ridge and Valley system.

In contrast the (MD) South Mountain fold is exposed nearly to the axis of the transverse arch, as identified by southward projection of the previously proposed traverse, and plunges gently toward (not away) from the axial culmination. To a first approximation, the axis projects across the axial culmination and the Gettysburg basin with little or no change of plunge. Certainly there is nothing that remotely approaches the amount of differential uplift of the (PA) South Mountain axis. While the genetic significance of this difference is not clear, it is egregious to claim a strictly homogenous tectonic origin for the two. For some time I have had an uncomfortable sense that the Chickies thrust conceals a considerable volume of rock I would like to know more about. Recently I have come to extend some measure of that sense to Blue Ridge relationships in southern Pennsylvania.

I hope that I have at least persuaded most readers that the Maryland-York fold and the Pennsylvania South Mountain folds are at least distinct anticlinals. If I succeed in infecting one who has more youthful vigor than I with my concern that the distinction may be greater than superficially apparent, I will consider my efforts fully repaid.

plain as residuals and now form some of the highest mountains of the region. Where the surface passed beneath the sea it was planed off by marine erosion and received a covering of sediment which was brought into the sea by the streams. In early Cretaceous time the surface of the plain adjacent to the sea sank gently so that marine deposits lapped successively farther inland and a great thickness of beds accumulated, but the wearing down of the surface of the land continued until there was renewed uplift in middle Cretaceous time, which renewed active erosion, and the streams began to cut their channels into the newly formed plain.

This peneplain is now found as the floor beneath the Lower Cretaceous deposits in New Jersey, eastern Pennsylvania, and Maryland. It formerly rose gradually inland, but is now much dissected, the parts composed of softer rocks having been entirely removed and even the portions cut on the hardest rocks have lost their original evenness. A remnant of an old peneplain surface preserved on the upturned bevelled edges of quartzite beds composing the top of Kittatinny Mountain in New Jersey and Pennsylvania now 1700 feet above sea is believed to represent this stage of peneplanation and this surface has been given the name of Kittatinny peneplain.

In Adams County and adjacent region the Kittatinny peneplain is believed to be represented by the higher flat parts of South Mountain at an altitude of 1,800 to 2,000 feet. The nearly even sky line of Green Ridge at 1,800 to 1,900 feet and broad crests of Big Flat Ridge at 1,900 feet apparently represent the surface where it was reduced nearly to a plain; the higher flat tops of Snowy Mountain and of Big Pine Flat Ridge, which are over 2,000 feet in altitude, probably represent portions of the surface that were not completely reduced to the peneplain level because they are composed of broad exposures of massive resistant Cambrian quartzites. The same peneplain is preserved on North Mountain and associated ridges west of Cumberland Valley at about 2,000 feet.

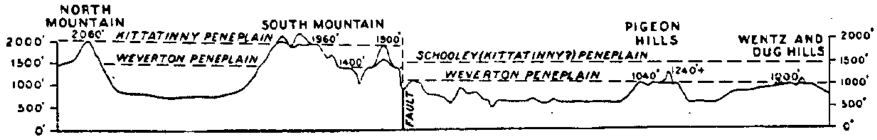


Fig. 10. Profile across Adams County and adjacent region, showing physiographic peneplains and their possible displacement by post-Cretaceous faulting.

The profile of the land surface southeast of Adams County (see fig. 10) shows that the floor on which the Cretaceous sediments were deposited, rises from sea level to 500 feet altitude in the vicinity of Baltimore and apparently continues to rise inland to 1,050 feet on the tops of the highest schist hills in Carroll County, Md. and to over 1,250 feet on the top of the Pigeon Hills in the eastern part of Adams County and in York County. This is considerably lower than the Kittatinny peneplain on the nearby South Mountain and if they are parts of the same peneplain it has apparently been faulted since its formation. Furthermore, the fanglomerate hills, which are Triassic alluvial cones, are believed to have been deposited in about the form that they now have and not to have been materially reduced by erosion.

From Stose, 1932.

SOUTH MOUNTAIN GEOMORPHOLOGY

by
G. Michael Clark
Department of Geology, University of Tennessee, Knoxville, TN

INTRODUCTION

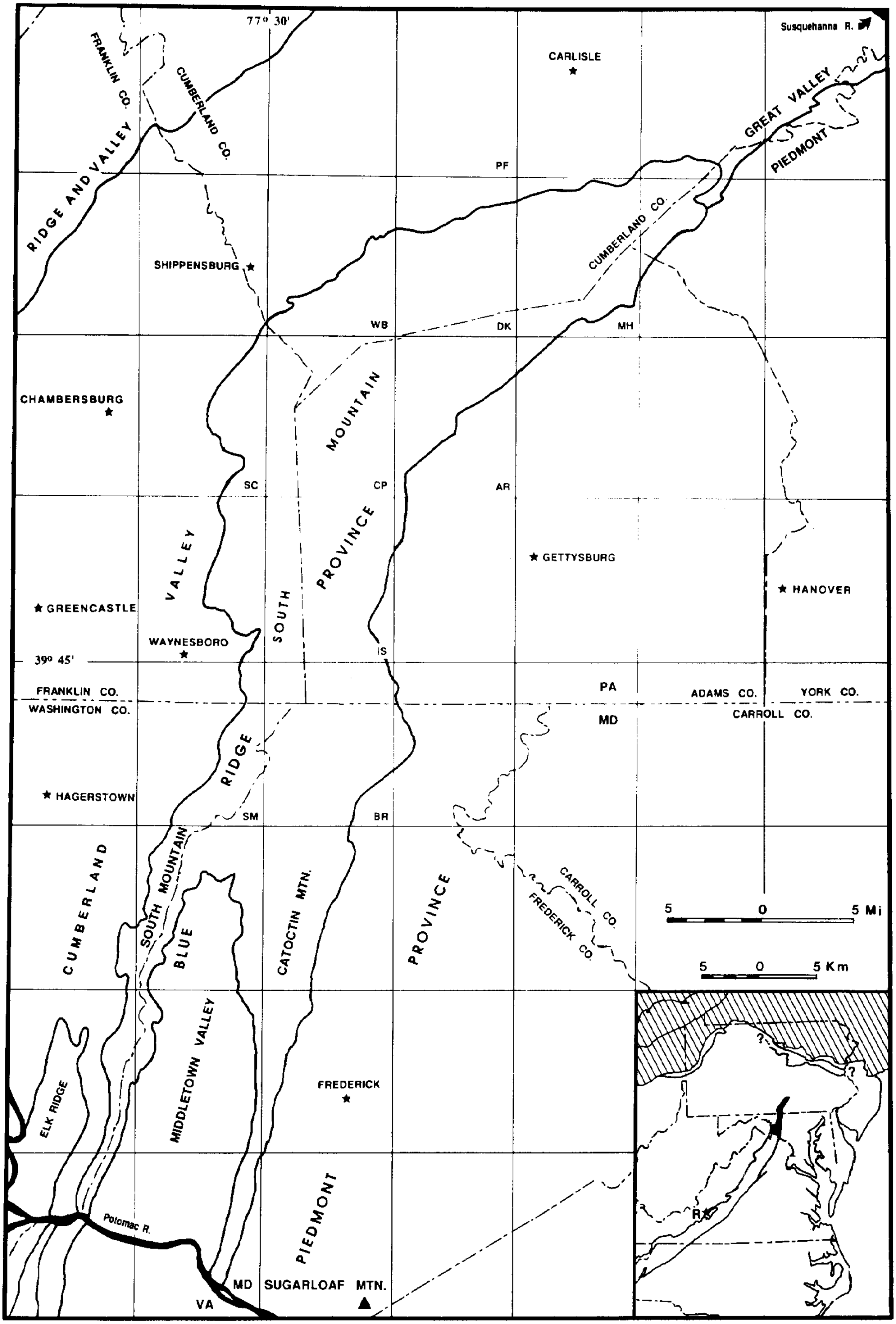
South Mountain is geographically near the midpoint of northeast-southwest Central Appalachian structural trends and comprises the northeastern terminus of the northern section of the Blue Ridge geomorphic province which extends south to the latitude of Roanoke, Virginia (Figure 11). Several modern overviews provide background for the present state of geomorphic research in this part of the Appalachians south of the glacial borders. Mills and others (1987) summarized studies of Appalachian geomorphic research and noted some pressing landscape origin problems in the region. Gardner and Sevon (1989) brought together papers on geomorphic evolution of the Appalachians that focused attention on fundamental questions about the geomorphic history of this mountain system. Mills and Delcourt (in press) describe several aspects of the nonglacial Quaternary geology of the Appalachians, including hillslope, fluvial, and periglacial features. South and Catoctin Mountains are at once both unique geomorphic entities and microcosms of Appalachian mountain geomorphology; they have individualistic landforms and materials, plus they illustrate subsets of problems in mountain geomorphology that are representative of many other parts of the Central Appalachians.

The origins of topography and drainage remain at the core of classical American geomorphology. Among the many unsolved geomorphic problems on South and Catoctin Mountains are the existence of accordant ridge tops that locally truncate lithology and structure in the upland areas, and the origin and evolution of the fluvial drainage from and around this massif. Early workers explained the summit "levels" as remnants of a once-continuous peneplain; subsequent arguments simply were about the number of partial peneplains that existed and about their correlation with surfaces elsewhere in the Appalachians. Fenneman (1938), for example, cited as distinctive the summit level at 366-396 m elevation on South Mountain that extends 8-10 km north of Potomac River and then is abruptly replaced by higher elevations to the north. With respect to the genesis of Central Appalachian rivers, this mountain range does not figure prominently in published drainage evolution schemes that mainly targeted the origin and evolution of transverse drainage. This omission is due, at least in part, to the fact that this upland is not bisected by transverse drainage north of the Potomac River.

As the father of the geographical cycle of erosion concept, Davis (1889) illustrated his notions about the peneplain as an ideal end state of fluvial erosion using the nearby Ridge and Valley province in Pennsylvania. The highest ridge crest levels were interpreted by Davis (1889) as remnants of a once-continuous peneplain, and the transverse paths of master rivers were used as supporting evidence. Davis' promulgation of the geographical cycle of erosion was so effective that for decades many workers concentrated on definition, description, and correlation of numerous summit level accordances, and the most common hypothesis used to explain Appalachian upland surfaces and transverse drainage was the peneplain concept (Sevon and others, 1983). Since the peneplain concept fell from favor (e.g., Flemal, 1971), a common practice has been to deny, ignore, or explain away the fact that ridge top relationships do exist. The ascendance in importance of process geomorphology in the United States might have discouraged researchers from investigation of field relationships that smacked of regional denudational chronology. Mills and others (1987) remarked on the overall shortage of geomorphologists working in the Appalachians, and noted many pressing applied geomorphic problems in the region that require significant input of human resources. These trends in American geomorphology, and perhaps other factors, have resulted in a nearly-complete stasis of geomorphic research on upland surfaces in the region as a whole and in the Northern Blue Ridge section in particular. Whatever the reasons, there is little modern geomorphic research on South and Catoctin Mountains to synthesize, although there are several excellent individual studies of selected features and areas (e.g., Godfrey, 1975; Middlekauff, 1987).

Geomorphic research today is concerned with aspects of process geomorphology, applied geomorphology, and Quaternary geology and geomorphology that are based upon sound theoretical and operational concepts, and that draw from related sciences in an interdisciplinary format. Examples of research that would be highly appropriate to further our understanding of South and Catoctin Mountain geomorphology could include: geomorphic studies of hillslope and fluvial processes; measurement of weathering, erosion, and denudation rates for selected catchments; dating the numerical ages of exposure and erosional histories of geomorphic surfaces using new cosmogenic methods; and surface and subsurface geomorphic and geophysical investigation of the complex diamicton deposits.

Figure 11. Opposite page. Index map of South Mountain-Catoctin Mountain area. Inset map shows northern section of the Blue Ridge geomorphic province, which extends from latitude of Roanoke, Virginia (= R) to northeastern end of South Mountain, Pennsylvania. Lined area = extent of Late Wisconsinan ice at 18 Ka; solid lines beyond 18 Ka border = extent of Pre-Wisconsinan ice; data from northeastern Pennsylvania courtesy of D.D. Braun. Main Map: Rectangular gridding shows location of 7.5-minute quadrangles; AR = Arendtsville; BR = Blue Ridge Summit; CP = Caledonia Park; DK = Dickinson; IS = Iron Springs; MH = Mount Holly Springs; PF = Plainfield; SC = Scotland; SM = Smithsburg; WB = Walnut Bottom. Base map from U.S.Geological Survey, 1:250,000 sheets, Harrisburg and Baltimore sheets.



77° 30'
Susquehanna R.
CARLISLE
FRANKLIN CO.
CUMBERLAND CO.
RIDGE AND VALLEY
GREAT VALLEY
PIEDMONT
PF
CUMBERLAND CO.
SHIPPENSBURG
WB
DK
MH
CHAMBERSBURG
SOUTH MOUNTAIN PROVINCE
SC
CP
AR
VALLEY
GETTYSBURG
HANOVER
GREENCASTLE
WAYNESBORO
IS
39° 45'
FRANKLIN CO.
WASHINGTON CO.
PA
MD
ADAMS CO.
YORK CO.
CARROLL CO.
RIDGE
HAGERSTOWN
SM
BR
CUMBERLAND
SOUTH MOUNTAIN
BLUE
CATOCTIN MTN.
PROVINCE
CARROLL CO.
FREDERICK CO.
5 0 5 Mi
5 0 5 Km
ELK RIDGE
MIDDLETOWN VALLEY
FREDERICK
Potomac R.
PIEDMONT
MD
VA
SUGARLOAF MTN.

REGIONAL GEOMORPHOLOGY

The concept that there are discrete natural regions that can be defined and circumscribed was richly developed by German and French geographers who identified and described areal entities they named Landschaften and pays, respectively. The traditional method of classification has been that of subdivision, with the construction of a descending hierarchy of successively finer subdivisions. The weakness in this approach, of course, is that it assumes we have some understanding of the causes of similarity and variation within and between landscape categories. Nonetheless, the regional concept was eagerly adopted by botanists, climatologists, foresters, geographers, physiographers, and soil scientists in western countries. In the United States the classical criteria for recognition of geomorphic provinces have been similarities or differences in: geologic structure, lithology, topography, and geologic history (Thornbury, 1965). Today in areas where landmasses can be identified as collages of microplates, this tectonic attribute can often be used effectively as a regional geomorphic criterion. Landscapes also bear the stamps of the various formative climatic environments under which they have evolved, although this factor has largely been ignored in American regional geomorphology. Finally, the operation of similar geomorphic process groups should logically be expected to result in similar erosional and depositional landforms, so that landform genesis would also seem to be a highly desirable criterion. Many process geomorphologists, however, would no doubt argue that to implement such a scheme successfully would require data and genetic understanding far in excess of those available at present.

The striking geologic and physiographic similarities within and differences between geographically large land units in the Appalachian Highlands lend themselves well to this type of landscape analysis. Fenneman (1938) and Thornbury (1965) agreed that the best rationale for dividing a landmass into provinces is the one that allows the greatest number of general statements about each subdivision before qualifications and exceptions become necessity. In most areas of the Appalachians, however, subdivision has not progressed much beyond subdividing provinces into sections. One hindrance to the advancement of regional geomorphology has been its lack of a quantitative basis. Godfrey and Cleaves (1991) specifically targeted the quantification of areal magnitude as a topic requiring numerical treatment if effective systems of landscape classification are to evolve, and constructed a ranking of landscape units based upon areal extent. The Godfrey and Cleaves (1991) hierarchy lends itself well to classification of landscape units in the Central Appalachians, and has been modified for use in the South Mountain area (Table 4).

An informal application of landscape classification to the South Mountain-Catoctin Mountain massif is illustrated in Table 5; some of the units are mapped onto Figure 12. Refinement of sectional-level landscape units into subsections, districts, and subdistricts is possible using existing geologic and topographic maps, and these subdivisions can be observed in the field. In

Table 4. Ranking of landscape units used in description of South and Catoctin Mountain landforms and landscapes. Modified from Godfrey and Cleaves (1991).

Rank/Areal	Magnitude (km^2)	Basis (Dominant Entity)	Examples
Realm	10^7	Largest plate-tectonic units	North American Plate
Major Division	10^6	Sub-continental entities	Appalachian Highlands
Province	10^5	Regional similarity	Ridge and Valley; Blue Ridge
Section	10^4	One tectonic-landscape style	Northern Blue Ridge
Subsection	10^3	Structure-landform similarity	South Mountain; Catoctin Mountain
District	10^2	Form-material relationships	Middletown Valley
Subdistrict	10^1	Direct material-form linkage	Ridge Road Upland
Zone	10^0	Few form-relief parameters	Diamicton apron
Locale	10^{-1}	Individual landforms	Stream terrace remnant
Compartment	10^{-2}	Single form/relief units	Lobe or terrace slope break
Feature	10^{-3}	Specific microform	Opferkessel; expanded fracture
Fracture			

Table 5. Informal subdivision of landscape lnits in the South Mountain-Catoctin Mountain area.

Province	Piedmont	Blue Ridge	Ridge and Valley
Province	Piedmont	Northern Blue Ridge	Middle Section
Section	Mesozoic extensional basins	South Mountain-Catoctin Mountain	Appalachian Great Valley
Subsection	Gettysburg Basin	Strong belt of NE-SW trending linear ridges	Cumberland Valley
District	The border fault zone complex	Ridge upland	Shale uplands
Subdistrict	Diabase dike ridge (major individual)	Diamicton footslopes	Carbonate lowlands

general, landscape units follow structural trends and the map patterns of lithologic units, as might be expected in a deeply-eroded old fold-belt mountain system. One striking difference, however, can be seen by contrasting the appearance of the west and northwest sides of South Mountain and the north half of the east side of Catoctin Mountain as shown on Figures 11 and 12. Northwest and west of South Mountain are extensive constructional lobelike aprons and sheets of diamicton deposits that extend out from the mountain fronts from one to several kilometers, achieve known thicknesses ranging up to 137 m, and thus are largely responsible for creating topographic form in these areas. The form and relief of this landscape is further complicated by the destructional effects of subsurface solution of the underlying carbonate bedrock, so that an unknown amount of land surface subsidence has occurred as well. Landscape classification, therefore, need not be restricted to the practical classification of land for planning and management; it can also have theoretical value in delimiting morphogenetic units. European and Canadian physical geographers, for example, have refined the techniques of geomorphological mapping in an attempt to provide a cartographic explanation of all of the mappable relief elements in an area.

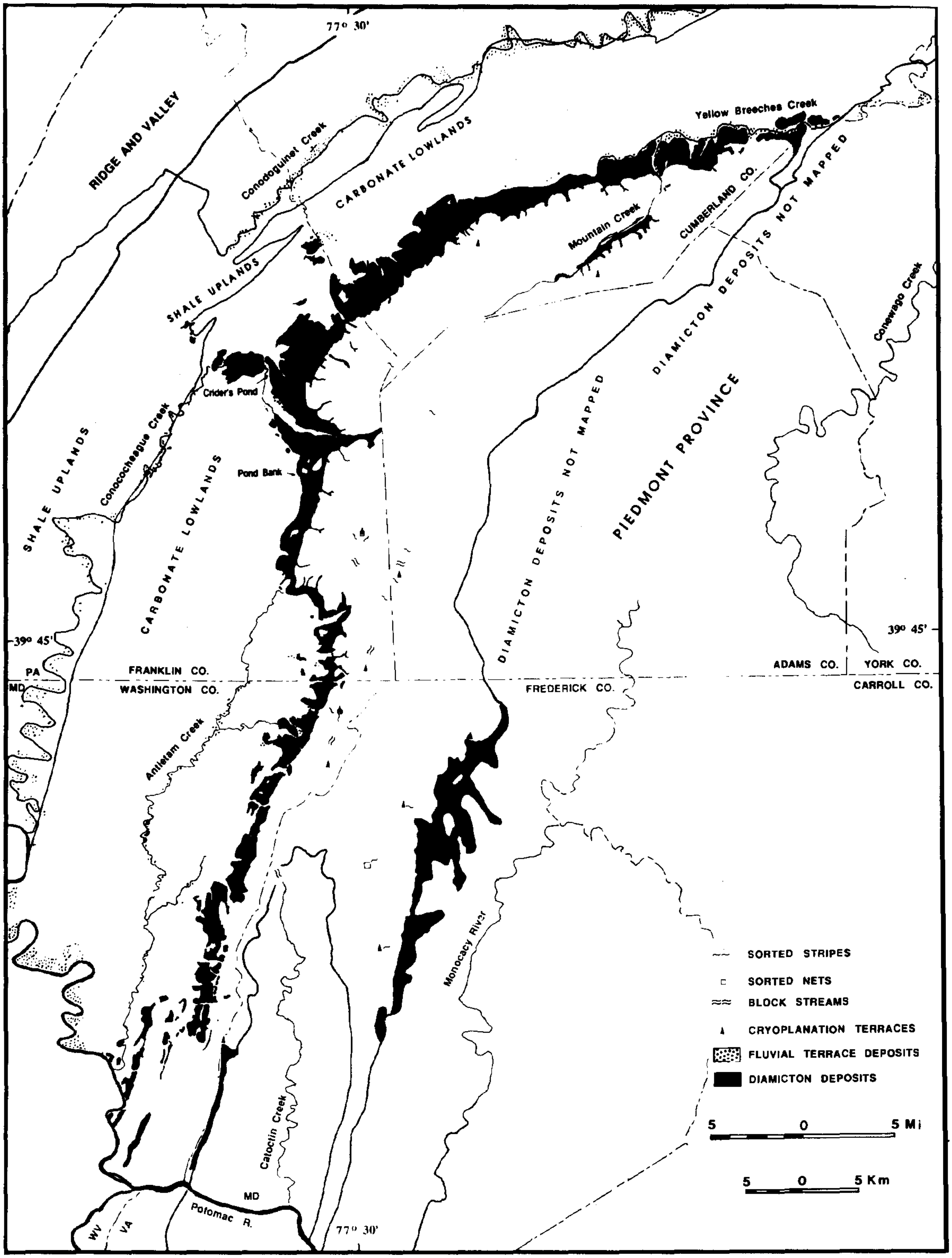
TOPOGRAPHY

The South and Catoctin Mountain massif extends 107 km from about 20 km WSW of the Susquehanna River to the Potomac River; first S45°W, then abruptly S25°W into Maryland, and finally, S10°-20°W to Potomac River (Figure 11). The sharp change in structural and topographic grain of South Mountain is at the latitude of Caledonia Park, and occurs across the Carbaugh-Marsh Creek fault zone that separates different folding intensities. Fauth (1968) reported that folds north of the Carbaugh-Marsh Creek fault zone are more numerous and more intensely developed than those south of the fault zone. In Pennsylvania and northern Maryland, South Mountain varies relatively little in width, averaging about 15 km. In Maryland, about 15 km south of the Mason-Dixon line, the Middletown Valley opens to the south separating South Mountain from Catoctin Mountain, both of which trend S10° to 20°W to Potomac River. South of the opening of Middletown Valley both South and Catoctin Mountains narrow toward the Potomac River, ranging in width from 1-3 km.

In the central part of the range in Pennsylvania and northern Maryland, the gently rounded to flat-crested upland summits reach 525-640 m; their slope angles vary from nearly horizontal to about 15°, except near and in the narrow V-shaped first- to third-order valleys where sideslopes of 20°-30° are common. On the outer mountain flanks away from exiting valleys, slopes increase and then decrease valleyward in a generally upward convexo-concavo form through a wide range of slope angles. Away from stream valleys that exit the mountain, the steeper crest slopes range from about 17°-23°. The lower midslopes range from about 6°-12° with an average around 9°. Footslopes on diamicton sheets and aprons are in the 1°-2° range in undissected areas. Maximum mountain relief is about 375 m and average local relief is approximately 130-200 m.

Where split by the Middletown Valley in Maryland, South and Catoctin Mountains have form differing from the topography in

Figure 12. Opposite page. Map showing locations of selected periglacial features, and locations of large fluvial terrace and diamicton deposits. Diamicton deposits in Cumberland, Franklin and Washington Counties generalized from 1:24,000-scale mapping completed from county soil surveys by N. Potter. Mapped diamicton deposits include bouldery, cobbly, and sandy deposits; some small areas of high alluvial terrace gravels, residual sand from weathering of the Antietam Formation, and other materials may be included at this scale. Areal distribution of diamicton deposits in Frederick County highly generalized from county soil map, and may include areas of fluvially reworked and retransported material, especially bordering streams and in distal areas. Mapped terrace deposits are old alluvium above modern flood levels, except for lowest terrace levels which may be flooded during extreme events. Base map from U.S. Geological Survey, 1:250,000 series, Harrisburg and Baltimore sheets. South Mountain to the north, and from each other.



77° 30'
RIDGE AND VALLEY
Conodoguinet Creek
CARBONATE LOWLANDS
Yellow Breeches Creek
Mountain Creek
CUMBERLAND CO.
DIAMICTON DEPOSITS NOT MAPPED
SHALE UPLANDS
Conewago Creek
Crider's Pond
Pond Bank
Conococheague Creek
SHALE UPLANDS
CARBONATE LOWLANDS
DIAMICTON DEPOSITS NOT MAPPED
PIEDMONT PROVINCE
39° 45'
39° 45'
PA
MD
FRANKLIN CO.
WASHINGTON CO.
FREDERICK CO.
ADAMS CO.
YORK CO.
CARROLL CO.
Antietam Creek
Monocacy River
Catoctin Creek
SORTED STRIPES
SORTED NETS
BLOCK STREAMS
CRYOPLANATION TERRACES
FLUVIAL TERRACE DEPOSITS
DIAMICTON DEPOSITS
5 0 5 Mi
5 0 5 Km
MD
Potomac R.
WV
VA
77° 30'

South Mountain represents the overturned western limb of the Blue Ridge Anticlinorium, with the topographic crest predominantly underlain by the Middle, or Ledgemaker Member of the Weverton Formation. To the east, Catoctin Mountain in the north is underlain by wide outcrop widths of the Weverton Formation that narrows southward, and in the south by a continuous strip of the Loudon Formation and by increasing outcrop widths of the Catoctin Formation.

DRAINAGE

No throughgoing fluvial system bisects the Blue Ridge north of the Potomac River, and surface runoff from South and Catoctin Mountains is therefore longitudinal either to the Potomac or the Susquehanna (Figures 11 and 12). The northern end of South Mountain is drained by Yellow Breeches Creek, tributary to Susquehanna River. Northwest of South Mountain the central and master streams draining Cumberland Valley are Conodoguinet Creek northeast to Susquehanna River, and Conococheague Creek south to the Potomac River. The longitudinal surface divide of contest is about 1.5 km south of the Shippensburg-Scotland 7.5-minute quadrangle boundary. South from the northern part of the Waynesboro quadrangle, however, Antietam Creek drains the mountain front southward to Potomac River. Southeast of South Mountain the main streams receiving flow from South Mountain are Conewago Creek draining northeast to the Susquehanna River, and, southwestward, for South and Catoctin Mountains, the Monocacy River which flows south to the Potomac River.

Within South Mountain map patterns in the low order drainage basins tend to be trellis-shaped on clastic rock units and dendritic on the Catoctin Formation. First-, second-, and third-order streams as delimited on 1:24,000 topographic maps have steep gradients about 55-75 m/km to their exits from the mountain fronts, where gradients drop to an average of 30 m/km on proximal diamicton aprons and decrease rapidly toward distal areas. On many such deposits, especially over carbonate bedrock units, the smaller streams are not perennial, and only the larger streams maintain surface drainage to the major longitudinal valley drainageways.

There is some evidence for structural control for the siting of certain stream reaches and individual "straight" stream segments. On South Mountain, several larger streams tend to parallel the position of mapped faults. For example, the fault trace of the transverse Carbaugh-Marsh Creek Fault is remarkedly parallel to the course of Conococheague Creek over most of its alluvial valley within the Scotland quadrangle (Fauth, 1968, Plate 1). This part of the valley is floored with alluvium, however, so that bedrock is obscured. A segment of the headwater area of Conococheague Creek northeast of the Chambersburg Reservoir is also paralleled with a mapped longitudinal fault (Fauth, 1968, Plate 1). Mountain Creek exits South Mountain in a water gap along a mapped transverse fault (Freedman, 1967). Of course there are many straight stream segments and water gaps on South Mountain that do not coincide with mapped faults or other known bedrock structures. Faults are not easily identified on South Mountain because of: the large thickness of mappable rock units, lithologic similarity in different parts of the stratigraphic section, gradational rock unit contacts, lack of stratigraphic control, and almost continuous soil cover. Moreover, streams may follow shattered and unhealed bedrock along structures that lack offsets such as lineaments, fracture traces, and major joints. Thus, the very limited bedrock exposures on South Mountain preclude any rigorous tests of structural control on stream location.

West of South Mountain in the Hagerstown Valley, Maryland, Nutter (1973) was able to locate sufficient exposures in carbonate bedrock units to map joint patterns. He compared rose diagrams of straight stream reaches with rose diagrams of joint strikes in an area north of the junction of Antietam and Beaver Creeks and concluded that the drainage pattern of Antietam Creek is strongly controlled by strike and cross joints, and that the map pattern of Beaver Creek also showed some, but lesser, joint control.

In the mountains only the larger creeks have floodplains that are continuously developed for appreciable lengths along their courses, and that have widths mappable at a 1:24,000 scale. Examples on South Mountain are Mountain Creek, Antietam Creek, and Conococheague Creek. In the Middletown Valley, Maryland, Catoctin Creek is tributary to Potomac River, and has developed a floodplain on Precambrian crystalline rocks in the core of the South Mountain Anticlinorium.

WEATHERING AND SOIL GEOMORPHOLOGY

Modern research throws increasing doubt on the simple dichotomy of physical versus chemical weathering; much weathering is turning out to be biochemical. Nonetheless, an elementary distinction can be made between clastic rocks that appear, on the outcrop level, to fragment mechanically (Figure 13), and those where loss by solution is evident. The only quantitative study of the effects of weathering in land denudation in the South and Catoctin Mountain area is by Godfrey (1975) who concluded that terranes with different lithologies would undergo differential lowering over time. An apparent paradox in the South Mountain upland is represented by extremes of rock decomposition. For example, there are exposures of firm bedrock, such as Hammond's Rocks (Stop 8), and sites where deeply-weathered, in-place, thoroughly rotted rock, or saprolite, occur such as Mt. Cydonia (Stop 5). Some, but certainly not all, of these weathering extremes can be understood by recourse to the composition and texture of matrix and cement in these rocks. In the case of resistant members within the Weverton Formation that can give rise to exposures of firm rock, the grains and rock fragments in the sandstone and conglomerate beds are held tightly by a dense matrix largely composed of sericite and quartz (Fauth, 1968). In weathered exposures of quartzites in the Antietam Formation, there is little to no primary matrix or cement remaining for study; this rock apparently had sufficient original permeability

Figure 13. Slump block separation from shattered outcrop near summit area of Salamander Rock. Notebook case 23 x 30 cm.

and a soluble matrix and/or cement to allow for removal of these materials. Field relationships at Mt. Cydonia suggest that what is observed there is the lower portion of the iron and clay accumulation zone. If this is so, then by analogy with full weathering profiles in subtropical and tropical regions that have similar iron and clay zones, the weathering profile has probably undergone deep truncation.

Soil maps have been prepared for the six-county area that includes the South Mountain-Catoctin mountain massif. Modern soil maps are valuable for a number of geologic purposes in addition to their obvious primary agricultural and silvicultural uses. In addition to many applied geomorphic uses, detailed soil maps depict soils developed from residual and transported geologic parent materials. On South and Catoctin Mountains many soils are developed on a wide variety of transported, or colluvial, parent materials. Bordering the major mountain creeks (Mountain Creek, Conococheague Creek, Antietam Creek) some soils have also developed on alluvial parent materials on the present floodplains. Soils developed from carbonate rock parent material in

Cumberland Valley also reflect their parent rock material. Soils developed from impure carbonate rock units tend to have a solum with silt loam texture; soils formed from material weathered from relatively pure limestone tend to have a solum with a silty clay or silty clay loam texture. In the Cumberland Valley and on the Piedmont, there are also modern soils developing on the floodplains, and on a number of alluvial terrace levels. Study of the terraces and the soils developed in them would help to unravel the Quaternary fluvial geomorphic history of the area.

The mineral and rock composition, texture, and structure of parent material have profound influence on two very important areas of soil science, and these are well illustrated in the South and Catoctin Mountain area. First, parent material exercises strong controls on the nature and properties of soil that develops in it. Both physical and chemical properties of soils bear stamps of original parent material; these properties are evident both in the field and in the results of laboratory soil characterization analyses. Second, a prime interest area of soil morphogenesis is to understand the genetic pathways along which soils progress as they develop over time. For example, two nearby soils with the same climatic, vegetational, and topographic environments but with significantly differing parent materials will evolve along quite different genetic pathways as the final major control on soil development, time, runs its course.

Soil survey information was used by Noel Potter to map both diamicton deposits and alluvial terrace remnants (Figure 12) on the north and west flanks of South Mountain in counties that have modern soil surveys. An interesting aspect of an exposure in these deposits at the Mainsville quarry (Stop 6) is the soil development at the top of the exposure (E. J. Ciolkosz, personal communication). A red paleosol is developed in the upper part of the diamicton, is truncated, and is overlain by colluvium in which the modern brown soil development is occurring. These soil geomorphic relationships indicate that the Mainsville diamicton deposit beneath the truncated red soil is relatively old, and probably predates Wisconsinan cold-phase events. If extensive mobilization of regolith on South Mountain did occur in Late Wisconsinan time, these sediments: (1) never reached downslope to the site of the present Mainsville operation; (2) bypassed the older deposits (in channels?); or (3) were present in at least some areas but were removed by one or more Latest Wisconsinan erosional event(s). Thus, soil geomorphology reveals glimpses of the complicated Late Cenozoic history of the diamicton deposits, and identifies problems that need study. Unfortunately, South Mountain has not received the detailed soil geomorphic research that has been conducted in colluvium in the Ridge and Valley province to the north (Ciolkosz and others, 1990). Soil geomorphic study of nearby high-level gravels and alluvial terrace remnants in the Cumberland Valley would provide new information about the origin and age of these clastic sediments that overlie carbonate bedrock units, and might also reveal information about their relationships with the mountain footslope and toeslope diamicton deposits.

PALAEOCLIMATOLOGY, WEATHERING, AND EROSION

Several closely-spaced glacial margins separate the Appalachian Quaternary realm into two divisions: regions known to have been glaciated by Laurentide Ice Sheets during Late Cenozoic glacial ages when ice was there, and areas beyond the glacial borders such as South Mountain in Pennsylvania (e.g., Braun 1989a). Present interpretation of the glacial and interglacial history of these unglaciated mountain landscapes is difficult because of the lack of datable geomorphic surfaces and deposits and the dearth of modern research.

Palaeobotanical research, particularly of pond, bog, or marsh sites is one approach that can provide information on vegetational history, and, by extension, information about former climates. Two such deposits in the South Mountain area are the Pond Bank and Crider's Pond sites (Figure 12). Both of these deposits are preserved in depressions on diamicton deposits on the western slopes of South Mountain.

Samples of organic-rich sandy clay from the spoil pile of the Pond Bank deposit discussed by Pierce (1965) were examined for plant fossils by Tschudy (1965) who assigned the deposit a Mid-Late Cretaceous age. Although no paleoclimatic interpretation of the vegetational assemblage was made, it may be assumed that the climate was sufficiently warm and wet to support lush vegetation and that weathering and erosion were at a maximum (Cecil, 1990). Unfortunately, no samples are known of the more than 20 feet (> 6.1 m) of lignite containing seeds and pieces of wood that is reported to occur at depth in the deposit (Pierce, 1965).

New information on vegetational history is quite likely forthcoming from Professor Alfred Traverse (personal communication, August, 1991) from a sample found at the Mainsville site in May, 1991. A clast of unconsolidated sediment containing organic material including vegetation macrofossil fragments was discovered on one of the cut benches in the Mainsville quarry, collected, and submitted for sample preparation and identification.

Crider's Pond, 3.2 km east of Scotland, Pennsylvania is at an elevation of 289 m. Watts (1979) studied the pollen and macrofossil remains from an 8 m organic core from Crider's Pond. From base to top the core was composed of: silt with black bands, banded silt, organic silt, peat, and organic silt. Several ^{14}C age dates were obtained: a near-basal date of 15,210 yr BP, a date in the banded silt of 13,260 yr BP, and a date in the peat of 11,650 yr BP. Watts also identified five pollen zones as follows: Zone Cr-1 (base), dominantly pine and spruce overlain by a small peak of birch dated to about 15,000 yr BP; Zone Cr-2, dominantly spruce with pollen of tall wet-meadow herbs; base of Zone Cr-3, abrupt and large increases in pollen diversity of both aquatic and woodland species, an assemblage similar to that found in the southern part of the present-day Boreal Forest; Top of zone Cr-3 red spruce followed by white pine; Zones Cr-4 and Cr-5, Holocene vegetational assemblages of predominantly pine and oak
The significance of the Crider's Pond fossil record is that it does document vegetational changes in the South Mountain area

during the past 15,000 years. The record does not indicate tundra vegetation such as occurs in the Longswamp sequence studied by Watts (1979) in Berks County, but rather a forest tundra dominated by spruce. Longswamp was within 60 km of the Late Wisconsinan ice border while Crider's Pond was 125 km from the ice and a half a degree farther south. This presumably accounts for the differences if the 14C dates are correct.

Watts (1979) suggests that discontinuous permafrost may have existed in the South Mountain area during and for some time following the Late Wisconsinan glacial maximum (18 Ka). He also suggests that the climate was cold, dry, and windy. It would have been these climatic conditions that controlled weathering and erosion during the period 25-15 Ka at comparable elevations and exposures. What conditions were like in the much higher and much more exposed mountain slope and crestal areas is not known, but they probably were much more severe. The probable products of this weathering and erosion are the subjects of much of this chapter.

An interesting aspect of the core stratigraphy at Crider's Pond is the presence of numerous scattered small rock fragments in zone Cr-3 suspended in the massive (visually unbedded) sediment. Watts (1979) considered solifluction and storm activity as possible agents of delivery for the rock fragments in the center of the pond.

Sparsity of dateable, uninterrupted, long-term terrestrial stratigraphic sequences in the Appalachians has driven researchers to the marine record. Using the marine 18O record as a proxy, Braun (1989b) concluded that, in addition to a major glaciation at about 2.4 Ma, eight out of ten major cold-phase events in the last 0.85 Ma, including the three for which there is stratigraphic evidence in northeastern Pennsylvania, would have brought severe cold-phase environments to the unglaciated Appalachians. An important geomorphic effect of these cold-phase events would have been enhanced erosion. Braun (1989) was able to find several localities where geomorphic effectiveness of mass wasting processes in generating colluvium could be assessed. These localities in east-central Pennsylvania, north-central West Virginia, and northwestern South Carolina bracket the South Mountain area both in latitude and in elevation. Calculated cold-climate hillslope erosion rates are 3-13 times greater than present-day fluvial erosion rates for eastern United States. Repeated for the number of major cold-climate events interpreted from the marine record, Braun (1989b) estimated that tens of meters of material has been stripped from ridge crests and transported valleyward during the last 0.85 Ma alone. Applying these data to South and Catoctin Mountains raises some interesting possibilities with respect to origins of the upland surfaces and diamicton deposits. These considerations will be discussed in subsequent sections.

Thus there are two end members of palaeoclimate for the South Mountain area: the Late Cretaceous which was warm and wet and the Late Pleistocene which was cold and dry. What happened during the 65 million intervening years? Until recently there was not much evidence which could be used for guidance. However, some diverse research is beginning to show a consistent pattern for Cenozoic palaeoclimate in the eastern continental area of North America.

Poag and Sevon (1989) reported on sedimentary deposits of the U.S. Middle Atlantic continental margin and showed a consistent pattern of decreasing siliciclastic deposition and increasing chemical sedimentation from the Late Cretaceous to the Middle Miocene. This indicates a decreasing amount of physical erosion and an increasing amount of chemical denudation in the Appalachian source area which includes South Mountain. This pattern changed significantly in the Middle Miocene when large quantities of sediment were transported to the offshore. What caused the marked change in nature and quantity of transported material in the Middle Miocene? Climate is one possibility.

Barron (1989) indicates that the Appalachians would have been an area of focused precipitation throughout the Cenozoic, but with gradually decreasing rates. Frakes (1979) discusses at length the complexity of the Cenozoic climate changes and their problems. Cenozoic climates in eastern North America apparently varied considerably during the first half of the era, but followed a major trend of increasing warmth and rainfall accompanied by a lack of pronounced seasonality. Frakes indicates that a major change in climate starts in the Middle Miocene with a trend of cooling and rainfall change which culminates in the Pleistocene. Tiffney (1985) discusses the vegetational changes that occurred in northeastern North America during the Cenozoic and notes that the warm temperate to subtropical vegetation which gradually developed to cover much of North America during most of the Eocene was gradually replaced as the climate began world-wide cooling and attained increased seasonality. A similar story is reported by Wolfe (1985).

A speculative scenario which can be created from the above is that during the first part of the Cenozoic (to the Middle Miocene) the climate was sufficiently wet and warm to support a cover of abundant vegetation which inhibited physical erosion but enhanced chemical erosion. These conditions caused deep weathering of rock, but allowed only a minimal amount of this weathered rock to be eroded in clastic form. As both climate and and associated vegetation changed significantly, a critical threshold was reached in the Middle Miocene and large amounts of clastic sediment were eroded in the Appalachians and transported to the Middle Atlantic offshore basin. Erosion slowed during the Pliocene, but was renewed in the Pleistocene.

It is suggested later (Stop 5) that saprolite developed in metaquartzite of the Antietam Formation, along with its associated iron cementation and kaolinite, are the result of the deep Cenozoic weathering. It is further suggested that the major erosion which gave the landscape of today its basic form commenced in the Middle Miocene. The final sculpting of the landscape was accomplished during the Pleistocene, and this is the subject of most of the remainder of this paper. Thus the landscape we see in South Mountain today is polygenetic in origin and owes little of its present form to weathering and erosion under the present climate.

PRESENT CLIMATE, WEATHERING, AND EROSION

South Mountain is within the area of humid continental warm summer climate. This climate occurs in the area of conflict between polar and tropical air masses. During the winter, polar continental air masses dominate with much colder weather interrupted occasionally by surges of tropical maritime air. During the summer, maritime and continental air masses bring higher temperatures and increased rainfall. The climate has a large annual range of temperature, high summer humidity, and about 150 frost-free days. Because of the influence of the nearby Atlantic Ocean, rainfall is distributed fairly uniformly throughout the year. Table 6 presents a summary of climatic data compiled from four stations near South Mountain.
The stations used for compilation of Table 6, Carlisle, Chambersburg, Gettysburg, and Shippensburg occur in low-relief, broad valley positions at elevations of 142,195, 152, and 207 m respectively. South Mountain is characterized by narrow ridges and

Table 6. Summary of climatic data compiled for Carlisle, Chambersburg, Gettysburg, and Shippensburg, Pennsylvania. Data from

Month	Monthly Prec.[1] Mean	Monthly Prec. Mean High	Monthly Prec. Mean Low	Daily[2] Prec. Mean Max.	Snowfall[3] Mean	Annual Mean Temp.[4]	Annual Mean Max. Temp.	Annual Mean Min. Temp.	Record High Mean Temp.	Record Low Mean Temp.	Days Temp.[1] Crosses Freeze Line
Jan	3.02	6.66	1.03	1.95	8.9	29.9	38.3	22.0	72.4	-12.5	18
Feb	2.64	4.88	0.52	2.07	8.7	31.5	40.6	22.6	73.9	-11.3	17
Mar	3.56	6.28	0.86	2.7	7.0	40.3	51.0	30.4	84.7	0.7	18
Apr	3.54	7.42	0.94	2.03	1.9	51.6	63.4	39.9	93.3	17.7	6
May	3.79	7.92	0.96	2.37	0	61.9	74.2	49.8	95.4	28.2	0
Jun	3.87	10.73	1.13	5.18	0	70.6	82.4	58.9	100.6	37.3	0
Jul	3.63	8.02	1.19	3.58	0	74.8	86.6	63.1	103.9	45.0	0
Aug	3.80	9.94	0.91	3.62	0	73.0	84.6	61.5	102.4	41.7	0
Sep	3.50	9.87	0.47	4.49	0	66.2	78.1	54.6	100.7	29.8	0
Oct	3.14	9.34	0.50	3.56	0	54.4	66.3	43.1	93.8	20.5	4
Nov	3.11	7.31	0.61	2.53	1.7	43.2	52.9	34.1	81.2	7.1	13
Dec	3.11	6.02	0.70	1.78	6.2	33.0	41.0	25.1	70.0	-10.0	19
Annual	40.63				33.2	52.5					95

1 - in inches; 2 - 1951-1980 data only; 3 - 1931-1980 data only; 4 - in degrees Farenheit.

	Years of record	Continuous since	Records
Carlisle	99	1916	Highest annual prec.: Shippensburg, 1937, 59.13
Chambersburg	94	1921	Lowest annual prec.: Chambersburg, 1930, 19.60
Gettysburg	112	1904	Highest monthly prec.: Carlisle, 6/1972, 18.51
Shippensburg	54	1933	Lowest monthly prec.: Carlisle, 12/1877, 0.05
			Highest temperature: Chambersburg, July, 107
			Lowest temperature: Gettysburg, January, -20

valleys, generally with local relief of 130 m or more. Low elevations of 152 m at the change from adjacent broad valley to mountain slope at the northeast end near Dillsburg rise to around 275 m near Shippensburg and then decline to about 245 m at Waynesboro. Maximum elevations range up to almost 640 m in the Big Flat area and are frequently greater than 450 m. These topographic conditions produce considerable variations in South Mountain climate from the surrounding recording stations. Unfortunately there are no year-around recording stations within the area of South Mountain to provide even generalities about the differences in microclimate.

When thinking about potential microclimate differences, the following should be kept in mind. The ridge crests will receive more and higher velocity winds than the valleys. During the day the valley bottoms may be cold sinks while the warmest temperatures are occurring on the shoulders of the slope. Slope orientation is very important on clear sunny days when the difference of light between the North and South facing slopes amounts to 46 units (say in g cal/cm2 hr-1). In diffuse light all slopes receive the same amount.

Amounts, types, and rates of weathering in the present humid continental climate of Pennsylvania have received very little attention. Sevon (1984) calculated an in situ atmospheric weathering rate of 0.26 m/Ma for a resistant sandstone in northeastern Pennsylvania. Pavich (1989) indicates that weathering in the Piedmont of Virginia produces upland regolith at rates between 4 and 20 m/Ma. Godfrey (1975) has done the most comprehensive work directly related to South Mountain and its rocks. His work in the Fishing Creek basin of Maryland involved rocks from the Harpers Formation and the Weverton Formation. Godfrey indicates (Godfrey, 1975, Table 4, p. 30) that the rate of soil formation decreases with increasing percent of matrix and varies from 18.3 m/Ma for 10 percent matrix to 3.66 /Ma for 50 percent matrix. These rates are in excellent agreement with those of Pavich (1989) and thus may provide a good indication of the rate of weathering in South Mountain under the present climate.

Erosion rates in the Appalachians under the present climate have been determined by many people and are reviewed by Sevon (1989). The various erosion rate determinations show great variability and a close relationship to disturbance of the surface by human activity. For the South Mountain area Godfrey (1975) determined a rate of land surface lowering of 2.5 m/Ma in a forested and undisturbed watershed.

PERIGLACIAL GEOMORPHOLOGY

Introduction

There is evidence for a strong periglacial influence in the development of landforms and materials on South and Catoctin Mountains (Clark and Ciolkosz, 1988). By the term "periglacial" (Lozinski, 1909) is understood cold climatic environments, with or without permafrost, and their landscape units (Godfrey and Cleaves, 1991) produced directly and indirectly through the process effects of strong frost action, intensive mass wasting, and aeolian activity operating on land that is seasonally snow free (Black, 1966, p. 329; Washburn, 1980, p. 2). Fluvial processes and fluvial landscape units are also of great importance, and differ from their humid temperate region counterparts in both form and process especially as they have been affected by ground ice and surface ice and snow (Clark, 1988; French, 1976; Worsley, 1984). Within a framework of cyclic-time and cold climate denudation, Troll (1948) enumerated six general periglacial process groups. These were: congelifraction, cryoturbation, solifluction, river gravel deposition, gelideflation, and cryoplanation. These periglacial process groups operate on different temporal and spatial scales. There are many specific processes, earth materials, and landforms produced due in part to the almost endless variety of space-time combinations, coupled of course with variations in local geologic materials, and surface and subsurface conditions. The treatment below will be limited to landforms, materials, and inferred processes of development that are relevant to the South Mountain-Catoctin Mountain area.

Some periglacial processes such as snow and slush avalanches operate quickly and produce "instant" landforms and deposits. Snow avalanches, and their stream-channel counterparts, slush avalanches, are important depositional agents, which produce or help to produce landforms such as cones, lobes, and avalanche road-bank tongues. Alpine mudflow events are another example of rapid periglacial processes that are important in presently-active mountain periglacial environments.

Small patterned ground features (less than about 40 cm) produced during diurnal freeze-thaw cycles are an example of periglacial features that can form over short time spans as days to weeks. Other periglacial processes require longer time intervals. The growth of ice wedges is an example of a periglacial process group that requires many years of growth (Mackay, 1990). Processes that operate to produce cryoplanation terraces are an example of even slower process rates. Priesnitz (1988) estimated that even under favorable conditions, cryoplanation terraces require on the order of 10,000 yr to reach full morphological development.

The size scale of periglacial features is important because forms of different sizes may have different origins and chronologies. Karte (1982) found that periglacial features can be grouped into three empirical categories: microforms, mesoforms, and macroforms. Periglacial microforms exist at the feature or compartment scales (Table 4), and thus there can be another size hierarchy nested within this category. Microforms include: both sorted and nonsorted patterned ground, gelifluction lobes, small frost mounds, and tors of periglacial origin (Washburn, 1980). Periglacial mesoforms are on the zone or locale scale of landforms (Table 4); examples are: block fields, block slopes, block streams, Felsenmeer, individual cryoplanation terraces, and snow avalanche tongues. Periglacial mesoforms grade upward in size into periglacial macroforms on the subdistrict scale level (Table 4). Examples of periglacial macroforms are: some types of large diamicton sheets and lobes, dells, certain asymmetric valleys, and the various domal, smooth convexo-concavo debris-mantled slopes, steplike, and planar features that collectively make up periglacial landscapes. Excellent overviews of how various scales of features interact and overlap to produce periglacial landscapes are in French (1976) and Clark (1988). There also may be subsurface periglacial features that do not fit into a surface classification, such as: bedded and oriented rock chip deposits, involutions, and rock and soil wedges.

Terminology is the bane of periglacial geomorphology. Usage problems remain, despite efforts in English language works, as those by French (1976) and Washburn (1980), who have attempted standardization and the separation of descriptive terms from terminology with genetic connotations. The terminology used here is descriptive and objective wherever appropriate and widely accepted standardized geometric nomenclature exists in the literature.

Periglacial Features

Introduction

To date (1991), features of probable periglacial origin reported on South and Catoctin Mountains include: tors, topographic terraces and risers, block streams, sorted stone stripes, and sorted circles and nets. Some of the forms are transitional with each other. Almost all of the forms and materials discovered on South and Catoctin Mountains to date are along or very near access routes as roads, trails, and power lines where both reconnaissance efficiency and visibility can be good. This bias in the inventory needs to be borne in mind. The reported paucity of large-scale sorted stone nets, for example, may be due to the lack of roads and trails over many of the high, nearly flat areas where these forms occur.

Block streams

Block streams are elongate swaths of contiguous blocks that have their long axis generally perpendicular to topographic contour (White, 1976). Smith and Smith (1945) reported the presence of block streams in the northern section of the Blue Ridge province. South and Catoctin Mountain block streams may occur in several different topographic situations, as in ravine heads, on sideslopes, or in valley bottoms. There are both forested and treeless block streams on South and Catoctin Mountains. Tree-covered sites dominate the inventory. Excellent examples of forested ravine-head block stream occurrences are along the access road between High Rock and the crest of Quirauk Mountain in the Smithsburg quadrangle, and below North View, Cat Rock in the Catoctin Furnace quadrangle. For example, one block stream below North View derived from talus breakdown, averaged 23 m in width, had an average gradient of 13.5°, and extended downslope 132 m. Outstanding examples of tree-covered valley bottom block streams are in the Fishing Creek drainage basin, Catoctin Furnace quadrangle, and were studied by Godfrey (1975). The several forest-free block streams are, however, visually the most spectacular, and they permit the study of block fabrics, microtopography, and other features unimpeded by vegetation mats. The following description of individual block streams will be limited to the treeless occurrences; some data for these forest-free block streams are in Table 7.

Table 7. Treeless South Mountain block streams.

Name Quadrangle	Latitude Longitude	Elevation (m)	Gradient (degrees)	Trend	Condition
Green Ridge North Iron Springs	39°49'50" 77°26'40"	415-501	--	--	
Green Ridge Central Iron Springs	39°49'15" 77°26'44"	381-512	5.1-11	S72-87°W	Disturbed (logging)
Green Ridge South Iron Springs	39°48'32" 77°27'15"	347-460	--	--	
T(r)ucker Run Waynesboro	39°48'54" 77°30'01"	463-488	2	S45°E	Excellent
Devils Racecourse Blue Ridge Summit	39°44'55" 77°27'13"	305-335	--	SSW	Treeless area removed
Raven Rock Hollow Smithsburg	39°40'20" 77°31'20"	421-482	3.5-5	S20-50°W	Disturbed in local areas
Black Rock Road Myersville	39°36'12" 77°34'11"	402-421	4-7	S10-65°E	Toeslope removed

Green Ridge Block Streams. The largest of these three block streams was studied in 1969 by students of Noel Potter, Jr., Dickinson College, as a research project. The students identified a main block stream with six tributary block streams and two subsidiary single block streams, one north and one south of the main feature (Table 7). A segment of Green Ridge about 1 km long and underlain by the Weverton Formation served as the source area for the blocks. The middle block stream varies in width from 9-46 m and extended downslope from the ridge crest for a minimum of 1675 m; the lowermost visible portion ends at water level in the Waynesboro reservoir. This block stream lacked trees and surface fine materials for most of its length. Three distinct linear segments of this main block stream were defined. A headward "source" section was characterized in the downslope direction first by a continuous block slope with two topographic terraces having treads inclined 15-20° and about 45 and 85 m wide separated by risers of 35°. These features had angular blocks up to 6 m in a-axis length and occupied the upper 180-215 m of this 610 m long segment. Downslope, the remaining portion of this upper section trended S72°W with a mean slope of 5.5°. The middle "central" segment was also about 610 m long, had a mean gradient of 5.1°, exhibited good edge definition and vertical orientation of tabular blocks, and was fed by a number of tributary block streams entering from the sides. Topographic depressions, elongate in the upslope-downslope direction, are common in the middle section; dimensions range from 0.9 x 2.4 m and 0.6 m deep to 2.4

x 9 m and 1.2 m deep. The lower or "bottom" segment trended downslope S87°W, had a mean gradient of 11°, and varied in width between 30-60 m.

Block fabric analyses showed several statistically-significant correlations. All block axes' lengths decrease with increased distance from the ridge crest in the headward source segment. Correlation coefficients for block size were weaker in the central and bottom segments. Block shape factors were more pronounced in the source and bottom segments, and block shape was more irregular in the central section. Block shape factors, when related to block axis dip, and their relationships with slope direction and slope angle showed weak to not significant correlation coefficients.

The multiple-tributary block stream investigated by Potter's students, and the two bordering single block streams would have been ideal features for additional research, Unfortunately, much of the Green Ridge block stream area has been subjected to logging by heavy mechanized equipment subsequent to this study, and many of the features have been disrupted or destroyed.

Tucker Run or Trucker Run. Earlier topographic maps named the fluvial drainage in this hollow Tucker Run, but the name has been changed to Trucker Run on later map editions. There are several discrete treeless areas along the upslope-downslope trend of this block stream separated by wooded areas. This is the only known treeless block stream on South Mountain that is in undisturbed condition (Figure 14), and it displays a number of pristine features. There is complex microtopography consisting of elongate to nearly circular cone shaped pits and some low mounds. Visual size sorting of large blocks from small boulders has produced sorted nets with the smaller stones in the center. Vertical orientation of tabular blocks is evident in several areas, with the ab-planes trending parallel to the gradient of the field.

Devils Racecourse or Devil's Racecourse. Thornbury (1954, Figure 16.14, p. 414; 1969, Figure 21.1, p. 511) reported this block stream west of Gladhill, Pennsylvania as an example of a palaeo-periglacial, or relict, landform. Source rock for this feature was apparently greenstone metabasalt derived from the Catoctin Formation. Some time prior to 1968, however, all of the treeless area had been removed for use in the manufacture of roofing material (G. M. Clark, field notes, 15 August 1968).

Raven Rock Hollow Block Stream. With a forest-free length of about 1 km, the Raven Rock Hollow block stream is the longest known treeless block stream in the South and Catoctin Mountain areas. This block stream shows several features typical of deposits formed by periglacial movement and is also exceptional because of its accessibility. The features include: obvious upslope source in Weverton Formation outcrops; downslope reduction in block size; some late-stage block orientation; patterned ground (circles) formed after the block field was emplaced; and recent development of Opferkessel. These features are discussed in detail at the description for Stop 1.

Black Rock Road Block Stream. This feature is located on the east flank of South Mountain west of Black Rock Road. An unknown volume of the toeslope of this block stream has been removed, but

Figure 14. View upslope of Tucker (or Trucker) Run block stream. Geopick on top of large on-edge block is 32 cm long.

much remains for study, including the potential for subsurface investigation in the excavated area. Upslope portions of this complex block stream display a number of features, including several feeder streams and highly complex microtopography. Slope gradient changes rapidly over short horizontal distances; slopes of 4 to 6 degrees are common as are small essentially horizontal areas. Microtopography is pronounced, and one feeder stream has a V-shaped depression oriented along gradient. The Black Rock Block Stream is an excellent example of longslope transport of surface blocks. Much of the surface blocks and boulders is derived from the Weverton and the Loudon Formations, although the lower 0.3 km of the features overlies the Catoctin Formation (Fauth, 1981).

Interpretation. Although relationships with mapped bedrock geology at several localities require minimum longslope transport of at least several hundred meters over very low gradients, there is no surface evidence of block stream migration today. Large blocks are weathering and breaking up in place without separation of the constituent fragments (See Stop 1). Neither the forested block streams nor the forested margins of the treeless examples show any disturbance of arboreal vegetation. Organic matter from lichens, mosses, leaves, and woody stem material slowly accumulates on and around blocks and boulders. Eventually, a vegetation mat develops that is capable of supporting rooted plant life. Trails and logging traces shown on maps over half a century old show no disruption in the field. Features within the block streams such as cone-shaped topographic depressions and sorted stone circles are nearly equant in plan view, and most logically formed after significant down-gradient motion ceased. An inactive, or more likely truly fossil, nature is the most rational conclusion that can be drawn from present evidence. Why some block streams are forested and why several are treeless is of interest. Observed relationships at the edge of the treeless features are most easily understood if a scenario of gradual forest encroachment is envisioned (e.g., Sevon, 1987; 1990).

Sorted stripes can be traced downslope into the headward areas of several block streams, and their downslope fringes may terminate as sorted stripes. Sorted circles and rarely sorted stripes as defined by concentrations of finer and more rounded boulders and cobbles are found within the treeless areas of several block streams. Physical continuity, therefore, is one line of evidence that links the origin of the block streams with that of sorted patterned ground, and a probable periglacial origin. Both visual and measured fabrics in South Mountain block streams are similar to those found in active periglacial environments. Thus there seems little doubt that South Mountain block streams are partly to wholly periglacial features.

Sorted Patterned Ground

Introduction. Both large scale (> 2 m dimension in plan view) sorted stone stripes and nets (Washburn, 1980) occur in the South and Catoctin Mountain areas (Tables 8 and 9). Most occurrences are either in areas underlain by the Weverton Formation and/or the blocks and boulders in their stone borders have been derived from this rock unit. Almost all of the finds to date (1991) have been on gently-sloping upland areas. Finds of sorted nets are relatively rare, the observed nets tend to occupy nearly horizontal upland flats and to become transitional to stripes in the slightly higher gradient areas downslope.

Table 8. Large-scale sorted nets on South and Catoctin Mountains.

Quadrangle Location	Latitude Longitude	Elevation (m)	Gradient (degrees)	Trend	Average width of stone border (m)	Mesh (stone-free center) dimensions (across slope x down slope, in m)
Iron Springs Snowy Mountain	39°50'02" 77°29'30"	622	0-3.5	N10°E	1-3	5.5 x 7.3
Smithsburg Quirauk Mountain	39°41'44" 77°30'49"	649	3.5-4	S40°W	3.6	2.5 x 5.0
Catoctin Furnace Salamander Rock	39°34'38" 77°29'19"	555-561	1.5-2.5	S15-30°E	2	4.6 x 6.5

Table 9. Large-scale sorted stripes on South and Catoctin Mountains.

Quadrangle Location	Latitude Longitude	Elevation (m)	Gradient (degrees)	Trend	Average width of stone border (m)	Remarks
Caledonia Park Headwater Valley Long Pine Run	39°57'53" 77°27'45"	530-533	3	N52°W	5.3	SE side of road
Caledonia Park Big Flat Ridge	39°58'02" 77°27'17"	590	1.5	S18°E	2.4	SE side of road
Caledonia Park Piney Mountain	39°55'42" 77°25'19"	558-570	3	N42-72°W	4.1	Both sides of road
Caledonia Park Piney Mountain	39°56'03" 77°24'56"	559	8	N52°W	6.9	SE side of road
Iron Springs Monument Rock	39°48'44" 77°29'02"	436	3-8	N85°E	8.0	N side of road
Iron Springs Snowy Mountain	39°50'02" 77°29'30"	622	2-3.5	N0°E	5.2	On tread surface
Iron Springs Currans Road	39°49'58" 77°29'52"	619	2.5-9	S20°E	5.7	SSE side of road
Iron Springs Three Springs Road	39°49'25" 77°26'27"	509	2-10.5	S80°W	6.5	W side of road
Iron Springs Three Springs Road	39°47'50" 77°27'23"	518	2-4	S50°E	5.9	NW side of road
Iron Springs Three Springs Road	39°47'07" 77°28'02"	436	7-10	N80°W	8.6	W side of road
Iron Springs Three Springs Road	39°46'55" 77°28'20"	415	10.5-12.5	N50°W	5.7	SE side of road
Iron Springs Currans Road	39°49'41" 77°29'21"	543-549	6-8	S50°W	8.1	W side of road
Iron Springs Currans Road	39°49'51" 77°29'17"	573-579	3-8	S40°W	7.7	W side of road
Waynesboro Snowy Mountain	39°49'31" 77°30'16"	594-600	4-7	N70°W	5	S side of road
Smithsburg Quirauk Mountain	39°41'44" 77°30'49"	649	4	S02°E	3.6	On tread surface
Smithsburg Mt. access road	39°41'49" 77°30'52"	646	4-10	N60°W	3	W side of road
Smithsburg Mt. access road	39°41'48" 77°31'12"	543-555	13	N75°W	6	WNW side of road
Smithsburg Mt. access road	39°41'44" 77°31'17"	549-556	15	N60°W	11	Both sides of road
Catoctin Furnace Power line	39°37'02" 77°26'41"	408-421	11-15	N32°E	6.6	Between towers RCM 113 & 115
Catoctin Furnace Salamander Rock	39°34'38" 77°29'18"	555-561	1.5-6	S15-30°E	5.9	N, NE, & E of tower
Catoctin Furnace Hamburg Tower	39°30'36" 77°28'27"	469-488	4-6	N70°E	9	ENE of tower

Only several occurrences of well-developed and well-preserved sorted stone nets are known (Table 8), but more probably occur on isolated high flat areas that lack road or trail access. The nets occur either immediately downslope from talus breakdown only a few meters below summit levels or on high flats without surface evidence of former outcrop. Downslope the nets are transitional to sorted stripes on slightly steeper land surfaces. Many of the sorted nets may have once been well-formed sorted polygons, as suggested by slightly angular corners in the stone meshes.

Sorted stone stripes, by contrast, are much more common on South and Catoctin Mountains and tend to occur on land with slightly higher slope gradients (Table 9). Individual sorted stripe length may vary from < 10 m to about 100 m. Sorted stone stripes occur as solitary forms, and also in conjunction with sorted stone nets on gently-sloping upland surfaces (Figure 15), block slopes, and block streams. It is common to find sorted stripes merging either downslope into block streams or having an apparent surface source upslope from block slopes or block streams. Some to much of the block and boulder material, however, may have come from subsurface bedrock sources.

Interpretation. As noted above some sorted stripes are physically continuous with block streams; stripes enter some block streams from upslope directions, apparently as feeders, and emerge at the distal termini of some block streams, apparently as disseminators. Sorted circles, defined as nests of smaller and more rounded cobbles and boulders, are present in block streams. Thus there is a physical, if not genetic, linkage between the origins of at least some block streams and some sorted patterned ground.

Most but not all large-scale (> 2 m mesh or stone border diameter/width) sorted patterned ground is of cold-climate, or periglacial origin (Washburn, 1980; Williams and Smith, 1989). The requisite conditions for other types of origins for these large features are lacking on South and Catoctin Mountains. Neither are there soils with large percentages of expanding clay minerals (Vertisols), nor are there soils that contain large amounts of salts (certain Aridisols). Goldthwait (1976) interpreted sorted polygons and nets over 2 m in diameter as features requiring permafrost for development, but Washburn (1980) urges caution. Regardless of the ground thermal state that accompanied their development, the above authors agree that the development of large-scale patterns over broad areas occurs only above treeline. The only known exceptions are local azonal conditions, as for example small sites with high water tables and areas where forest cover has been removed by human activity.

If these sorted patterned ground features are old, how have they survived tree growth and tree throw? The answer may lie in the subsurface, as these features are of the deep "rooted" variety, with tabular blocks tending to be oriented with their ab planes vertical and anchored firmly in the subsoil. There are few places where the maximum depth of stone concentrations can be seen on South and Catoctin Mountains, but for the larger stripes observed the approximate depth to base of contiguous stones ranges from about 1 m to a maximum of 1.7 to 1.8 meters. If permafrost was present during sorted stripe development, this stone depth might represent the thickness of the active layer. On the other hand, depth to base of contiguous blocks may simply

Figure 15. Sorted stripe trending from lower right to photograph center, then splitting into sorted net area upslope, Snowy Mountain. Tor, composed of shattered rock derived from the Weverton Formation, is on left skyline. Length, with handle, of 50-m tape resting against survey monument is 33 cm.

represent the effective depth of highly-disruptive seasonal frost activity. As negative evidence, surface periglacial features that lack a blocky armor are not conspicuous in the South Mountain-Catoctin Mountain area. If well-developed surface features such as nonsorted patterned ground, solifluction lobes, and small nivation hollows once formed here during Pleistocene cold phases, they may have been obliterated by the subsequent 10 Ka of tree-root activity in the Holocene. A few vague forms that may represent old solifluction lobes and nivation hollows may be seen in some places. For example, northwest of the topographic profile across High Rock Road (called Three Springs Road on the Iron Springs 7.5-minute quadrangle, but High Rock Road on the Michaux State Forest road map and sign posted as such) (Figure 16C) are very subtle lobe- and hollow-like microtopographies that might easily be overlooked, but that are visible to the believer.

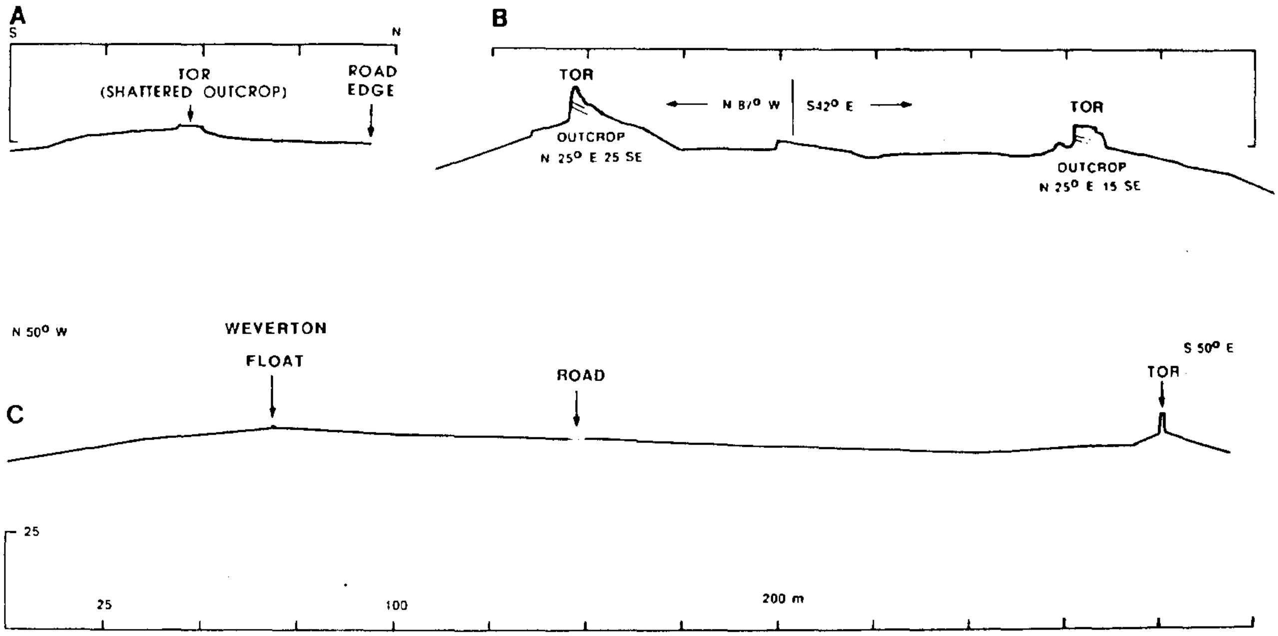


Figure 16. Topographic profiles across edge of upland surfaces, risers, and treads; no vertical exaggeration: A. Snowy Mountain site, profile line is due South (left)--North (road); B. Cat Rock, note bend in line of section. Attitude of bedding in the Middle, or Ledgemaker Member, of the Weverton Formation varies away from lines of profile (Fauth, 1977); C. High Rock Road site; profile line is N50°W-S50°E.

Tors, Summits, and Upland Terraces

Introduction. One of the most striking and most scenic attributes of South Mountain is the gentle upland subdistrict, as typified by the Ridge Road Upland Area. The Ridge Road has been described as the Skyline Drive of Pennsylvania, albeit at a degree and a half higher latitude and about 500 m lower elevation than otherwise comparable topography in Virginia. High flats along parts of the Skyline Drive, as in the vicinity of Stony Man, Hawksbill, and Big Meadows, Virginia are examples. As these features occur on the zonal, subdistrict, and even district topographic scale units, they form important aspects of South and Catoctin Mountain landscapes, and will be treated in some detail. Some of the flats in South Mountain have been named, such as Big Pine Flat Ridge, Big Flat Ridge, and East Big Flat Ridge on the Caledonia Park quadrangle and Big Flat on the Walnut Bottom quadrangle.

Despite the work of Monmonier (1967; 1968; 1971) and Hack (1975; 1989) on the major relationships between rock type and structure, which indicates that peneplanation is not a viable explanation of Appalachian uplands, the problem of the origin of flat upland surfaces still remains. Both South and Catoctin Mountains display outstanding upland "flats". In areas where bedrock is exposed, it can be shown that this gentle topography truncates lithologic differences and structural trends. Closer inspection of the "flats" in many areas reveals that this general appearance is a collage of smaller tread- and riser-like features that, in the aggregate, give the visual appearance of flatness. For example, Godfrey (1975), Middlekauff (1987), and Olson (1989) have all referred to the steplike nature of zone-, locale-, and compartment-sized landscape units on uplands on South and Catoctin Mountains.

Methodology. Areas on South and Catoctin Mountains in Pennsylvania and Maryland were chosen for investigation to see if field relationships at significant distances (about 200 km) from Late Cenozoic continental ice margins (Figure 11) have experienced major topographic modification. This is because general agreement exists today that periglacial effects near the glacial border were severe and that extremely harsh environmental conditions also extended to low elevations (Clark and Ciolkosz, 1988; Watts, 1979). For example, Watts (1979) reported clear evidence of tundra vegetation in association with colluvium in areas within 35 km of the Late Wisconsinan border at the Longswamp, Berks County, Pennsylvania site (40o 29' N; 75o 40' W) at an elevation of only 192 m. Marsh (1991, written commentary) remarked on the great number of flat uplands in the Ridge and Valley province in central Pennsylvania, and noted that contour maps from digital elevation tapes display flat, abrupt-edged uplands better than standard 7.5-minute quadrangle maps, even with the same contour interval. As one example noted by Marsh, part of a digital elevation map covering the Hartleton, Pennsylvania, 7.5' quadrangle shows excellent examples of these highest flats that Marsh reports break abruptly into blocky slopes.

Topographic maps covering South and Catoctin Mountains at 1:24:000 scale (Figure 11) were examined to identify local broad upland sites on the zonal scale (Table 4). Many of the identified uplands display prominent topographic benches and some show chimney-like summit protuberances. Localities for field study were chosen where comparable-scale geologic maps were available, or where it was known from other sources that bedrock exposures were present. Topographic profiles were constructed from traverses run at right angles to topographic slopes. Slope profiles and surficial materials were described utilizing standard geomorphological procedures (Gardiner and Dackombe, 1983). Although most locations have road access corridors, several sites do not, including locations along the Appalachian Trail or near power transmission lines. Locations in the following section on description of sites refer to Figure 12 and Tables 10 and 11.

Description of Sites. On and around Hammond's Rocks, Freedman (1967, Figure 21 and Plate 1) mapped structural elements that detail the trend of the Hammond's Rocks anticline. Dips of bedding, jointing, and cleavage are discordant to the topography, but are also nearly perpendicular to the overall topographic trend of tor-like features on the upland (See Stop 8, Day 2). Structural details mapped by Potter and others show a striking discordance between structure and topography on and around Hammond's Rocks (See Stop 8). Farther northeast, along the Hammond's Rocks Ridge Road, flat upland surfaces exist that are underlain by the Montalto Member of the Harpers Formation as mapped by

Table 10. Locations of selected South and Catoctin Mountain local broad upland sites.

Site No. (Table 11)	Quadrangle Locality name	Latitude Longitude	Elevation (ft) Elevation (m)	Feature name Aspect
1	Mount Holly Springs, PA	40°04'05-20"	1405-1520	Summit with tors
	Hammond's Rocks area	77°14'30-55"	428-463	Crestal area
2	Dickinson, PA	40°03'35"	1460-1470	Southerly sloping area
	High Mountain Road	77°22'15"	442-463	Southerly sloping area
3	Iron Springs, PA	39°50'03"	2020-2040	Summit with tor remnant
	Snowy Mountain	77°29'30"	616-622	Crestal area
4	Iron Springs, PA	39°48'02"	1760-1800	Summit with tor
	Three Springs Road	77°27'08"	536-549	Longitudinal divide
5	Blue Ridge Summit, MD-PA	39°43'44"	1680-1694	Summit area
	Mount Dunlop	77°29'23"	512-516	Crest
6	Smithsburg, MD-PA	39°41'44"	2120-2140	Riser, terrace
	Quirauk Mountain	77°30'48"	646-652	S10°W
7	Smithsburg, MD-PA	39°39'38"	1560-1580	Summit tor
	Buzzard Knob	77°32'20"	475-481	Crest
8	Catoctin Furnace, MD	39°36'58"	1520-1560	Tors, subsummit flat
	Cat Rock	77°26'54"	463-476	N86°W-S41°E
9	Catoctin Furnace, MD	39°30'36"	1600-1620	Tor, risers
	Hamburg Lookout Tower	77°28'31"	488-494	Crest
10	Keedsville, MD	39°25'00-52"	1320-1400	Risers, terraces
	South Mountain	77°38'19-27"	402-433	N62-82°W

Table 11. Form and material attributes of South and Catoctin Mountain upland sites. Numerical values given are for representative specific features measured.

Site No. (Table 10)	Width (down slope) Length (along slope) (m)	Tread gradient Riser gradient (degrees)	Tread material Riser material	Tread features Riser height (m)
1	30-40	4°	Soil with blocks, 1-8 m a-axis	Sparse stones
	65	20-38°	Outcrop, loose blocks	4-12
2	80	4°	Blocky soil	Sparse stones
	200	15-25°	Outcrop, block rubble	15
3	43-105	2-4°	15-25 % blocks	Sorted stripes
	170	8.5-15°	Outcrop, shattered outcrop	2-3
4	230	1-2.5°	Relatively stone-free soil	Smooth slopes
	300	20-90°	Tor, loose block rubble	1
5	40	0-11°	Outcrop, stony soil	Shallow
	150	summit	Summit	N/A
6	190	3.5-6°	Stony soil	Sorted stripes
	190	>10°	Blocks	5, disturbed
7	50	2-4°	5-35% blocks	Float
	62	27-30°	Outcrop above scree apron	15
8	120	1.5-16.5°	0-30% blocks	Shallow
	215	52.5-113°	Bedrock	4.5-5.8
9	330	1-5°	Outcrop, stony soil	Two treads
	275	30-85°	Outcrop, shattered outcrop	1-4
10	40-115	2-10°	< 5% stones	Few blocks
	235-445	44-53°	Blocks, outcrop	3-11

Freedman (1967). There are few outcrops on the upland surface, however, and structural, lithologic, and topographic relationships are obscure.

Snowy Mountain, also on South Mountain, provides an example of a summit tor landform rising above bordering treads that constitute the local broad upland surface at this site (Figures 15 and 16A). Bedrock exposures in the area comprise thick-bedded, coarse-grained sandstone of the Weverton Formation (Fauth, 1978). Dip of rock cleavage and bedding are strongly discordant to the local tread topography. Large-scale, sorted patterned ground on the tread can be traced upslope toward talus breakdown below the outcrop.

Across the High Rock Road, there is upland topography between a bedrock tor developed in quartz-vein rich Weverton Formation, a bordering area of loose blocks derived from this outcrop, and gently-sloping upland surface northwest of these features. A topographic profile (Figure 16C) shows these gentle slopes, risers on both sides of the tor, and the topographic expression of the tor.

On Mount Dunlop, the summit topography is characterized by gentle slopes over distances of hundreds of meters. The Weverton Formation bedrock (Fauth, 1978, Plate 1), dips from 10° northwest to 75° overturned to the southeast. Outcrop-level structures display even more spectacular discordances between topography and bedrock attitudes. Farlekas (1961) mapped folds in a thin but locally continuous "quartz" unit over a portion of the summit area, and showed the striking discordance between the bedrock and summit topography.

Along South Mountain in Washington County, Maryland, bedrock attitudes in the Weverton Formation (Edwards, 1978) are discordant to the trend and slope of local upland topographic surfaces. Summit topography on Quirauk Mountain, Maryland, follows in general the major structures and outcrop belt in the Weverton Formation as do most of the summits and knolls on South Mountain (Godfrey, 1975, p. 5). At the outcrop level, however, there are discordances between topography and the sense of bedding in shattered outcrops. An excellent example of cross-cutting relationships occurs south of the U. S. Army Information Systems Command Site C installation where most of the riser and tread topography is still preserved. Large-scale sorted stripes on the tread can be traced up slope gradient to contiguous blocks that comprise the riser. Also present are funnel-shaped depressions and microhollows visually similar to features that Demek (1969, Photo 15 and 16) termed respectively "solifluction forms" and "nivation funnels". Farther south on South Mountain at the Buzzard Knob site, tor microtopography and a bordering terrace, both discordant to structure, are well displayed. At the South Mountain site, Hedges (1975) noted excellent riser and terrace development that crosscut bedding in the Weverton Formation along the Appalachian Trail between the Townsend Memorial and Lambs Knoll.

Similar antipathy between bedrock attitudes and local summit topography also obtain at Cat Rock (Figure 16B) and east of Hamburg Lookout Tower on Catoctin Mountain. Whitaker (1955b) mapped and drew cross-sections of an area east of Hamburg Fire Tower in detail (Figure 5, p. 454) and showed the contrast between asymmetrically overturned shear folds in two members of the Weverton Formation and the plateau-like bench and scree summit topography. Mapping by Fauth (1977) also illustrates discordance between structure and topography at these two locations. Taken together, the Maryland Blue Ridge sites demonstrate that local summit topography is discordant to the dip of bedding structure and lithology on both limbs of the South Mountain Anticlinorium where a variety of structural attitudes exist. Nor are distinctive treads and risers confined to the clastic rocks. Excellent tread and scarp topography is developed in the Catoctin Formation along the Foxville Fire Tower road near the boundary between the Catoctin Furnace and Myersville Quadrangles, and also in the Cunningham Falls vicinity in the Blue Ridge Summit Quadrangle.

In conclusion, rock weathering and surface soil horizon characteristics at all of the sites provide qualitative evidence of relatively prolonged slope stability. When broken open, both bedrock ledges and float blocks show differential weathering effects on top *versus* bottom areas. Large blocks are weathered and broken up in place, with little separation of the constituted fragments. Gently-inclined surfaces of some large quartzite blocks show well-developed Opferkessel (weathering pits in quartzite) that show no morphological evidence of block disturbance during the time they have developed. Hedges (1969) assembled convincing evidence that Opferkessel are contemporary features that develop slowly under modern environmental conditions, although the rates of formation are unknown.

Visual evidence of active surficial processes on local broad uplands is largely confined to weathering, and the estimated rates of denudation are low. Ciolkosz and others (1990) reviewed estimates of modern rates of soil formation from sandstones on ridge crests that range between 0.026 and 1 cm/ka. Godfrey (1975) calculated chemical erosion rates for metaquartzites of 1.8 m/Ma for a drainage basin; ridge top rates might be lower or higher. Unless estimates of current rates of weathering and denudation on resistant rocks that underlie these uplands are radically in error, topographic development during Holocene time has been negligible, requiring that summit landscape development occurred in Pleistocene or earlier time.

Proposed Hypothesis: Palaeoperiglaciation. It is therefore hypothesized that the individually-small (zone or locale; Table 4) forms and materials developed on South and Catoctin Mountains are relict cryoplanation features that formed under rigorous periglacial environments (with or without permafrost) that are no longer operative. One suggestion that would explain the evenness of skyline impression is that the aggregate visual effect of these features, developed to common elevational ranges in local areas, is that of a much larger single summit "plain" when seen from a distance. This interpretation argues that large areas of Northern Blue Ridge local broad uplands can be viewed as relic incipient-to-essentially-complete assemblages of much smaller individual surfaces of cryoplanation (*cf.* Figure 17), as opposed to remnant fragments of peneplanation. A further conjecture is that in the aggregate the visual "evenness of skyline" effect

Figure 17. Clearcut area showing terrace wraparound of outcrop area, South Mountain, Pennsylvania. (Dickinson, PA 7.5-minute quadrangle; High Mountain Road; elevation 442-463 m.).

that these summit and bordering summit landforms have on our visual perception is that very impression which led early workers to adopt the peneplain remnant hypothesis. This hypothesis is presented as an alternative to the peneplain hypothesis which long fascinated geologists both here and abroad (*cf.* Thornbury, 1965; Adams, 1975; Sevon and others, 1983; Fulton, 1989) but does not disprove the peneplain hypothesis. Nor does it disprove the possible production of planation surfaces by the etchplain mechanism argues by Budel (1982), or through a process of topographic reversal that has been demonstrated to work on steeper lower slopes (Mills, 1981). These other hypotheses are not reviewed in detail here; see Clark and Hedges (in press).

The term "cryoplanation" (Bryan, 1946) is understood to mean cold climate land reduction with concomitant development of conspicuous upland flats, risers, terraces, and other features (e.g., Demek, 1969: p. 5-8; Washburn, 1980: p. 237). Although periglacial environments have also been considered as regions of extremely active valley incision and destruction of plains (e.g., Budel, 1982), modern workers increasingly recognize the presence of planar upland landforms that presumably have been produced under periglacial environmental conditions (e.g., Priesnitz, 1988). Extensive documentation exists on cryoplanation forms from a wide variety of localities (e.g., Demek, 1969). Given suitable bedrock structure, lithology and, climate, one overall impact of periglacial environment on summit- and near-summit level landforms is the production of upland flats, risers, and treads, although all of the responsible processes and required ground frost environments have yet to be elucidated (Washburn, 1985).

Mention of cryoplanation in the Central Appalachians is not new. Peltier (1949, p. 30, 67-69) invoked cryoplanation as a process to explain the form and relief of mountain tops in the Susquehanna River Valley. North of the glacial border Berg (1975, p. 32) indicated that surface morphology of Wisconsinan-age till has been modified by periglacial cryoplanation. Godfrey (1975, p. 7) noted that areas with "flat outcrops" along South Mountain, Maryland closely resemble the "cryoplanation terraces with frostriven scarps" of Demek (1972). Middlekauff (1987) and Olson (1989) also noted the presence of steplike landforms in the Blue Ridge Province in Maryland. Péwé (1983, Figure 9-11, p. 169 and Table 9-7, p. 177) reported an unpublished observation of cryoplanation for Mt. Davis, Pennsylvania.

That periglacial processes might be responsible for major modification of summit topographic forms over wide areas to the extent that they could be mistaken for surfaces of other origins is, however, a more sweeping proposal. Russell (1933) studied geomorphic form in western United States and stated this concept clearly:

> "Herein lies the explanation of numerous forms, ranging in size from minute, steplike benches to slopes covering whole mountain sides and broad surfaces across highlands which may readily be mistaken for parts of peneplains." (1933: 939).

There are few references to cryoplanation summits and terraces as major elements of Appalachian ridge tops over wide or disparate areas. Hedges (1975) proposed that cryoplanation had produced the truncation of bedrock structure and lithology on Sugarloaf Mountain on the Maryland Piedmont, and also noted well-developed terraces at the South Mountain, Maryland, site. Clark (1989a) offered the option that Central Appalachian ridge crests might be incipient cryoplanation features rather than remnants of peneplains, but did not publish site data in support of this speculation.

The morphological features reported in this study are similar or identical to forms described as cryoplanation features by previous workers. Tors on the local broad upland summits resemble tors that have been interpreted as periglacial in origin. For example, Ehlen (1990) studied topographically-similar summit landforms developed on granites in Dartmoor. She found that the summit tors are composed of rocks that are highly resistant to erosion, and are parted by highly-spaced vertical joints and medium-spaced horizontal secondary joints. Ehlen (1990, personal communication) indicated that there is a consensus among researchers that the Dartmoor tors are of periglacial origin. In the Appalachians, Braun and Inners (1990) and Braun (1990, personal communication) interpreted tors morphologically identical to those reported in this paper as periglacial in origin and not as relict Tertiary features.

Cryoplanation summit flats described by Demek (1969, p. 7) and those illustrated by Péwé (1970) (see Figure 18) resemble those of the uplands, and can be quite extensive, as in the Ridge Road Upland Area. On the outcrop level, trend and dip of bedding is discordant to local upland surfaces at the localities studied, although of course over broad areas (as major folds) there is a general positive relationship between major bedrock structures, lithologies, and mountain summit topography.

The form, relief, and materials of risers described (cf. Table 11) conform remarkedly well with descriptions of cryoplanation risers reported in the literature (cf. Table 12). Features composed of firm bedrock, shattered bedrock, and slumped to jumbled blocks singly and in various combinations are present in the study sample.

Table 12. Geomorphic form and material attributes of reported cryoplanation terraces compiled from Demek, 1969; Priesnitz, 1988; Péwé, 1975; Reger and Péwé, 1976).

Characteristic	Reported data		
Slope and slope position			
Summits	Summit cryoplains, with or without tors		
Slopes	Interrupt middle and upper slopes; generally several levels		
Width (along contour)	30 m (low)	400 m (common)	> 10 km (high)
Length (normal to contour)	5 m (low)	100 m (common)	> 1 km (high)
Riser or scarp height	1 m (low)	6 m (common)	> 50 km (high)
Gradient			
Tread or flat	1^{o} (low)	6^{o} (common)	14^{o} (high)
Riser or scarp	9^{o} (low)	30^{o} (common)	90^{o} (high)
Tread material	< 1-3 m debris depth, rarely more; sorted patterned ground		
Riser material	Bedrock, shattered bedrock, block rubble 1-3 m thick		
Rock type	Resistant lithologies that produce coarse debris favorable		
Lithologic control	Commonly initiate at or near breaks in lithologic reistance		

The sizes, shapes, surficial geomorphic materials, and site factors of treads are also similar to features of cryoplanation terraces described in the literature (cf. Tables 12 and 13). For example, Demek (1969, p. 55-56) noted that, although cryoplanation terraces are rare to absent on steep slopes, their development is not excessively dependent on leveled surfaces. The same topographic situations are present at the study sites. One shortcoming of this study is that detailed subsurface data that would permit mapping of the bedrock-soil interface are lacking. Many investigators consider cryoplanation terraces as transport slopes, with 3 m or less of debris mantling the bedrock. On the other hand, Demek (1969: p. 6-8) also recognized less common cryoplanation terraces with external parts composed of loose material. He termed these complex features compound cryoplanation terraces.

Important problems remain, moreover, and some new ones are raised. The volume of debris remaining on terraces cannot be computed until the soil-bedrock interface is mapped with tomographic geophysical techniques. Rates of removal cannot be computed and the geomorphic effectiveness cannot be assessed until

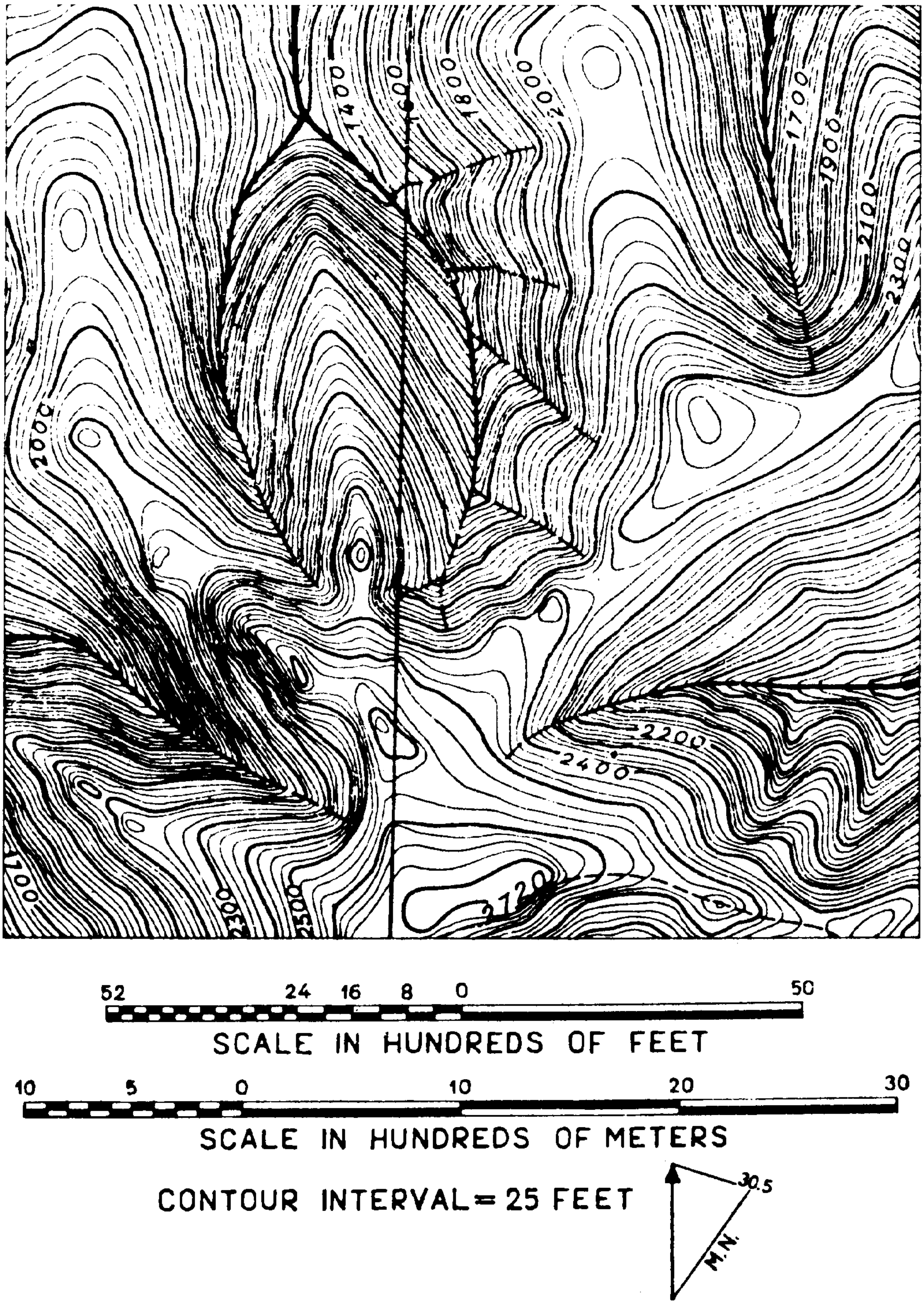


Figure 18. Topographic map of ridge at head of Kokomo Creek, 33 km NE of Fairbanks, Alaska showing cryoplanation terraces cut on Birch Creek Schist at elevations of 610-825 m (2,000-2,700 feet). Contour interval 25 feet. (From Péwé, 1970, Figure 2D. p. 361).

Table 13. Environmental (site) associations on cryoplanation terraces (compiled from Priesnitz, 1988; Péwé, 1975; Reger and Péwé, 1976).

Influencing factor	Favorable conditions	Unfavorable conditions
Wind effects on snow drifting	High	Low
Exposure to prevailing winds	Lee slopes	Windward slopes
Proportion of snow in total precipitation	High proportion	Low proportion
Sublimation rates	Low	High
Relation to local snowline	300-500 m below to very little above snowline	Rarely down to 1000 m below snowline
Summer temperatures	Cold	Warm
Freezing index	> 1000 degree days (^{o}C)	---
Permafrost required		
Continental areas	Yes	---
Maritime areas	No	---
Formation time (estimated)	ca. 10,000 years	> 10,000 years

age(s) of the terrace deposits and the downslope diamictons into which they grade are determined. Numerical ages will, in most cases, have to be measured with new methods of dating of geomorphic surfaces and materials. The time span required for cryoplanation terrace development has only been estimated (Table 13). Further complication is indicated by evidence that individual terraces may record more than one episode of development (e.g., Lauriol, 1990).

A hypothesis has been presented that relates the origin of local broad uplands on South and Catoctin Mountains to cryoplanation during cold phases of Late Cenozoic time, and, in speculation, that the accordance of their summit levels may be due to the effects of periglacial processes that worked to produce a common elevation range. Rather, it is hoped that this hypothesis will stimulate detailed geomorphic research on upland rocks, soils, and landforms which can be tested with newly-developing field and laboratory techniques. These techniques will also be useful to search for and evaluate evidence for or against deep weathering stripping (Büdel, 1982) as a process group that could also help to explain the formation of local broad uplands and perhaps summit accordance as well. The same methodologies should also help unravel geometries and depositional histories of the complex diamicton deposits that border these Blue Ridge upland surfaces.

It is of course premature to make palaeoclimatic interpretations until definitive studies of both the Appalachian features and comparable active analogs have been completed. Some notions, though, of the potential of cryoplanation terraces for providing palaeoenvironmental information are in order (cf. Table 13). Karte (1982) noted that, in addition to forms found in treeless, cold continental permafrost environments (e.g., Reger and Péwé, 1976), cryoplanation terraces are forming above the forest limit in maritime "Icelandic type" areas without permafrost where mean annual air temperatures are between -1^{o} to -2^{o} C. When definitive work on active forms becomes available, moreover, terraces may provide opportunity for other types of palaeoenvironmental inferences in a region where such data are difficult to obtain. In Alaska, for example, the altitudes of cryoplanation terraces plot about 100-300 m below the Wisconsinan snow line (Péwé, 1975, Figure 8, p. 22). The proportion of rain to snow in total precipitation is another possible palaeointerpretation. Péwé (1975, Figure 8, p. 22) showed that the elevation of cryoplanation terraces in central Alaska is lower in areas where the proportion of snow to rain is higher. Palaeowind interpretations and other aspect inferences may also be possible in the future (Table 10).

The genesis of accordance in elevation or "evenness of skyline" remains unaddressed. Tarr (1898) suggested that rock above tree line undergoes more rapid weathering and erosion than in areas below the forest limit. Daly (1905, p. 120-123) discussed the accordance of summit elevations above the forest limit in mountains and argued for the effectiveness of alpine processes in both rock disintegration and removal of rock waste. On South and Catoctin Mountains, much work must be done in order to define montane palaeoclimatic altitudinal zones, the landscape-forming processes that operated there, and their geomorphic effectiveness.

Diamicton Deposits

Introduction. Another problem at least spatially related to the ridge uplands is the presence of complex, thick, and extensive diamicton deposits along many mountain flanks that border the upland surfaces especially in the Ridge and Valley and Blue Ridge provinces (e.g., King, 1950; Pierce, 1966; Godfrey, 1975, Moss, 1976; Clark and Ciolkosz, 1988; Clark and others, 1989; Braun, 1989b; Ciolkosz and others, 1990). Previous workers have referred to these sediments as: "mountain wash", "alluvial mountain wash", and "alluvial cones of mountain wash". On South and Catoctin Mountains, what were the specific source areas for the debris, what were the mechanisms of entrainment, transport, and deposition of the sediments, and when and under what conditions did these events occur? Except in several areas in the Appalachians, (e.g., Braun, 1989b), little is known about the geometry and chronology of diamicton emplacement, so that questions of erosion magnitude and rates, geomorphic effectiveness of the responsible process groups, and Quaternary landscape history have been difficult to address in a quantitative manner.

Methodology. Potter (unpublished data) used the Cumberland, Franklin, and Washington County soil survey reports to map diamicton deposits on the north and west flanks of South Mountain. Data were compiled at a map scale of 1:24,000. These data have been generalized and are shown in Figure 12. The older county soil surveys are not as useful in identifying and mapping diamicton parent material in detail. Thickness to bedrock of these unconsolidated deposits is known only at a few localities. Becher and Root (1981) mapped diamicton thickness from well logs on the northwest flank of South Mountain in part of Cumberland Valley. This mapping indicates that thickness is highly irregular but in general decreases away from the mountain front. A maximum depth to bedrock is 137 m, although within 1-2 km of the mountain front well depths to bedrock of about 60 m are more common. Nutter (1973) studied the hydrogeology of these sediments in the Frederick and Hagerstown Valleys, and mapped their thickness using well log information along the foot of South Mountain on the east side of the Hagerstown Valley. Maximum reported depths to bedrock were 390 feet (119 m) and 400+ feet (> 122 m) 2.5 miles (4 km) and 7 miles (11.3 km) SSW, respectively, of Smithsburg, Maryland. Overall, Nutter found that the diamicton deposits decrease in thickness away from the contact between the Antietam and Tomstown Formations. A few seismic refraction lines have been run in several areas and generally corroborate the well log data. It would be helpful to know the form of the soil-bedrock interface as well, but sufficient detail to do this is lacking.

The Mainsville quarry (Stop 6) provides a temporary opportunity to study some of the diamicton material in plan and section, and the main aspects of that site are discussed later. Despite the information presently available from the Mainsville site, there are no absolute answers on when and under what palaeoclimatic, vegetational, and soil conditions the deposit formed, although Early through Middle Pleistocene time constitutes a reasonable temporal interval.

Interpretation. If development of the several types of block accumulations, sorted stone stripes, cryoplanation landforms, and other periglacial features involved deep erosion and transport of bedrock, periglacial processes may indeed be responsible for removal of much material from the mountain uplands (see Liestl, 1961 for a modern-day example), but how were the volumes of coarse to fine clastic sediment transported and deposited? Are the present-day channels strictly features formed in Holocene time, are they only the sites of former channels that are now being further deepened, or have they shifted laterally, perhaps by some mechanism of gully gravure? Until recently, few references were made to sources or transport mechanisms for these great volumes of sediment. Carter and Ciolkosz (1986) conducted a systematic study of bedrock depth on a ridge crest in the Ridge and Valley province, and found that the soil is 3 to 4 m deep. Based on seismic and rock fragment orientation data and the presence of sorted and stratified material in the upper 1 to 2 m of the profiles, they concluded that the upper 1 to 2 m of the soil is colluvium and the underlying material is residuum. Braun (1989b, p. 249) calculated that 8 to 10 periglacial episodes of the last 850,000 yr have been capable of tens of meters of ridgetop reduction. Ciolkosz and others (1990) showed that the upper parts of ridge-crest residual soils in the Ridge and Valley province of central Pennsylvania were truncated during Late Wisconsinan time and then either buried with local colluvium or cryoturbated. These nearly in situ parent materials have been stable since Late Wisconsinan time as evidenced by the nature and properties of soils developed in them. It is not known how applicable the studies in the nearby areas of the Ridge and Valley province are to South Mountain, but they can certainly serve as a guide for discussion and future research.

Could erosional magnitudes sufficient to form the features interpreted as periglacial landforms and deposits described on South and Catoctin Mountains have accounted for at least some of the thick diamict deposits that underlie simple side slope and complex lobe-shaped landforms in this part of the Blue Ridge province? The development of periglacial features described above would have produced large volumes of debris that would have been transported valleyward, although the rates of many periglacial processes are not well known (Washburn, 1980). In speculation, if widespread and deep periglacial weathering, erosion, transport, and deposition accompanied a number of cold-climate phases, the cumulative production of debris may have been sufficient to account for the materials both at the base of the Blue Ridge and those spread out into the bordering geomorphic provvinces (Figure 12). Suggested processes include, but are not limited to: snow and slush avalanches, alpine debris flows and mudflows, and catastrophic high-flow fluvial transport. Much better knowledge of the origin and age of these diamicton deposits will be required to evaluate these scenarios, however.

SYNTHESIS

Present climates in the Northern Blue Ridge are humid continental. Soil temperature regimes in the Central Appalachian mountains are mesic on the lower ridge crests and frigid at the higher elevations (Carter and Ciolkosz, 1980). Cryic soils are unknown and frost pockets are unlikely in topographically well-drained areas. Under natural forest conditions with snow cover, Central Appalachian mountain soils are frozen in winter to depths of less than 25 cm (Carter and Ciolkosz, 1980). Leffler (1981), moreover, demonstrated that Appalachian summit temperatures can be estimated reliably; his method has been refined further by Schmidlin (1982). These summit-level temperature estimates are characteristic of middle-latitude forested mountains. The predominant natural summit area vegetation on South and Catoctin Mountains is deciduous forest cover (Braun, 1950).

Set against the modern climatic environment are reports refining the Pleistocene glaciation sequence in Pennsylvania (e.g., Braun, 1989a), and on palaeoclimatic and palaeovegetation data beyond the ice sheets in eastern North America as a whole (Delcourt and Delcourt, 1981; 1984), including the last deglaciation (Jacobson and others, 1987). There is palaeobotanical evidence that the South Mountain-Catoctin Mountain area experienced severe cold-climates during Late Wisconsinan time as the work by Watts (1979) illustrated. Sediment influxes present in core samples suggest that the cold climates were accompanied by landscape disturbance on the district to lower landform scales (Table 4) on the lower slopes of the South Mountain area surrounding Crider's Pond (Watts, 1979). Physiographic location of sample sites is important for interpreting the palaeoenvironmental record. Mountainous areas such as the higher parts of South Mountain probably lacked arboreal vegetation and effective ground cover, had more severe palaeoclimates, and experienced greater landscape disturbance than lower-lying areas because of several reasons.

Important contributing topographic factors are: temperature decrease with increasing elevations (adiabatic lapse rates), greater and more varied exposure to atmospheric precipitation and wind, redistribution of snowfall, cloud cover, and steeper slopes with greater potential energy of gravitation. Thus, the envisioned reconstructed upper mountain climatic-vegetational environments would have been favorable for severe landscape disturbances that are interpreted to have had profound effects on Appalachian Highland landscapes (see Braun, 1989b).

It is not yet possible to construct an altitudinal zonation of periglacial features on South and Catoctin Mountains. In other parts of the world, when azonal features produced by unusual local conditions are excluded from the inventory, a common sequence of zones (1 to 7) from lower to higher altitude is as follows: (1) ploughing and braking blocks (Wanderbl_cke); (2) terrace-like garlands; (3) earthflows; (4) miniature patterned ground; (5) sorted or stony lobate forms; (6) large-scale sorted patterned ground; and, given resistant lithology, (7) a mantle of riven rock (Felsenmeer). These zones overlap in space in present-day periglacial environments, and it is important to realize that what is seen in palaeoperiglacial zonation is a collage of features resulting from waxing and waning environments as altitudinal climatic belts moved up and down environmental gradients. All of the above situations are further complicated by exposure and ground conditions. Thus, the interpretation of altitudinal zonation is extremely difficult, and leads to a subset of questions, not the least of which is whether palaeoperiglacial features can be used to answer them. Where were treelines? Did permafrost exist and above what elevations? Can mean annual soil or air temperatures and their probable ranges be deduced?

Although modern basic geomorphology as practiced in the United States is highly process driven, still there is interest in geomorphic form. On South and Catoctin Mountains two overall kinds of landscapes on the district to subdistrict scale levels (Tables 4 and 5) can be seen in the field and on large-scale, small-contour-interval topographic maps. These are: smooth convexo-concavo debris-mantled slopes, and gently-inclined summit and near summit planar features. The convexo-concavo landscapes are magnificently developed on the outer sides to lower slopes of South and Catoctin Mountains and can be seen from many vantage points in the Piedmont and Ridge and Valley provinces. These debris-mantled slopes merge down gradient into the complex lobe-like to sheetlike topographies that overlie the complicated and often thick diamicton deposits. On the crests and in the interior of the mountains and away from incised fluvial landscapes are the gently inclined summit surfaces with or without tors, that may be flanked by subtle risers and subjacent topographic terraces that have been interpreted by Clark and Hedges (in press) as cryoplanation features. Karrasch (1974) studied alpine landscapes in Europe and in the United States and described both landscapes with smooth rounded forms (Glatthänge) and landscapes with stepped flat-and-riser slopes (Treppenhänge). He attributed the development of Glatthänge landscapes to development conditioned by high-relief mountain environments and the formation of

Treppenhänge landscapes to environments with lower relief that are underlain by structural-lithologic irregularities (Karrasch, 1974). Albeit with less relief, the same general topographic and substrate conditions are met on South and Catoctin mountains with the same topographic results.

Assessment of the quantitative effects of long-term erosional activity on overall relief between mountain summits and bordering valleys is beyond the scope of this presentation. Godfrey (1975) measured and compared present chemical and physical rates of erosion in the South Mountain area, Maryland. He speculated that cold-climate processes would have reduced overall relief between ridge crests and the intervening Middletown Valley, whereas temperate climatic weathering and erosional processes would tend to increase relief. During times of temperate climate Godfrey (1975) estimated that the Catoctin Metabasalt beneath the valley would be eroded at a rate three to four times faster than the Loudon and Weverton Formations on the ridge crests. Under Pleistocene cold-phase conditions he concluded that the ridge crests would have been lowered faster than the valley, causing an overall reduction in landscape relief. Braun (1989b) hypothesized that Pleistocene periglacial erosion should be the dominant Pleistocene erosional process in the Appalachian Highlands, and that present processes are shaping the landscape to provide slope forms necessary to transport the periglacial debris. To address the hypotheses and speculations presented by Godfrey (1975) and Braun (1989b), the form, volumes, stratigraphy, and ages of summit erosional landforms and diamicton deposits will need to be determined. New methods for the numerical age dating of geomorphic surfaces hold great promise, as do evolving geophysical tomographic techniques that image subsurface soil and rock conditions. Only then can we begin to address concepts of landscape equilibrium and disequilibrium in a quantitative fashion in areas where surficial evidence indicates that effects of Late-Cenozoic environmental change have been severe.

Caution in making conclusions about specific palaeoenvironments must still be urged for several reasons. Washburn (1980) documented the dearth of information about formative environmental conditions of features active in present day periglacial environments, and he noted little improvement five years later (Washburn, 1985). To further complicate reconstruction of palaeoperiglacial environments, the degree of similarity between Central Appalachian conditions their modern day analogs, is not clear. These analogs occur either at much higher elevations in other mountain chains (for similar latitude and sun angle) or at much higher latitudes (for similar elevations). Also, the lack of subsurface information and numerical age dates restricts our ability to use analogs in present-day periglacial environments and to compute process rates and make correlations. For example, Büdel (1982, p. 78-80) interprets certain bedrock erosional surfaces as pre-Quaternary features that have only been overprinted by Pleistocene cold phase palaeoperiglacial processes.

CONCEPTUAL HYDROGEOLOGIC FRAMEWORK OF A REGOLITH-MANTLED CARBONATE SYSTEM, CUMBERLAND VALLEY, PENNSYLVANIA

by
Douglas C. Chichester
U. S. Geological Survey, Lemoyne, PA

INTRODUCTION

As part of the Appalachian Valleys and Piedmont Regional Aquifer-System Analysis (APRASA) project, the U.S. Geological Survey (USGS) is conducting a type-area study in northeastern Cumberland Valley, Pennsylvania (Figure 20). The study is designed to quantify the hydrogeologic characteristics and to improve understanding of the ground-water-flow system of the highly proddductive and complex regolith-mantled carbonate-rock aquifer on the northern flank of South Mountain. This study area is believed to be typical of regolith-mantled carbonate-rock aquifers that occur in other areas along the southeastern margin of the Valley and Ridge physiographic province from Pennsylvania to Alabama (Swain and others, 1991).

The scope of this study includes (1) estimation of hydraulic properties of the mantling regolith, carbonate aquifers, and streambeds; (2) assessment of the role of regolith in storage and flow of ground water to the underlying carbonate aquifer; (3) assessment of the role of springs and a diabase dike in movement and discharge of water from the ground-water system; (4) assessment of the depth of the regional flow system and affects of fracturing on flow in the carbonate aquifer; (5) development of a water budget for the study area; and (6) quantification of inter-basin transfer of water. Later in the project, these quantities, estimates, and assessments will be tested for reasonableness by means of a computer model of ground-water flow and mass (water) balance.

This paper describes (1) the hydrogeologic setting of the study area, (2) conceptualization of ground-water flow in the study area, and (3) water budgets for three stream basins having contrasting geology.

Description of Study Area

The study area lies within the Conodoguinet and Yellow Breeches Creek watersheds, in the northeastern part of the Cumberland Valley, in southcentral Pennsylvania (Figure 20). The area is bounded to the north by Blue Mountain, to the east by the Susquehanna River, to the east-southeast by the drainage-basin divide of the Yellow Breeches Creek, to the south by South Mountain, and to the west by the drainage-basin divide of Middle Spring Creek. Within the study area, ground water discharges locally to streams and springs, and surface water flows to the Conodoguinet and Yellow Breeches Creeks, both of which parallel the axis of the valley and drain eastward into the Susquehanna River. The study area has approximately 30 springs that discharge more than 1 ft^3/s (cubic foot per second) in the

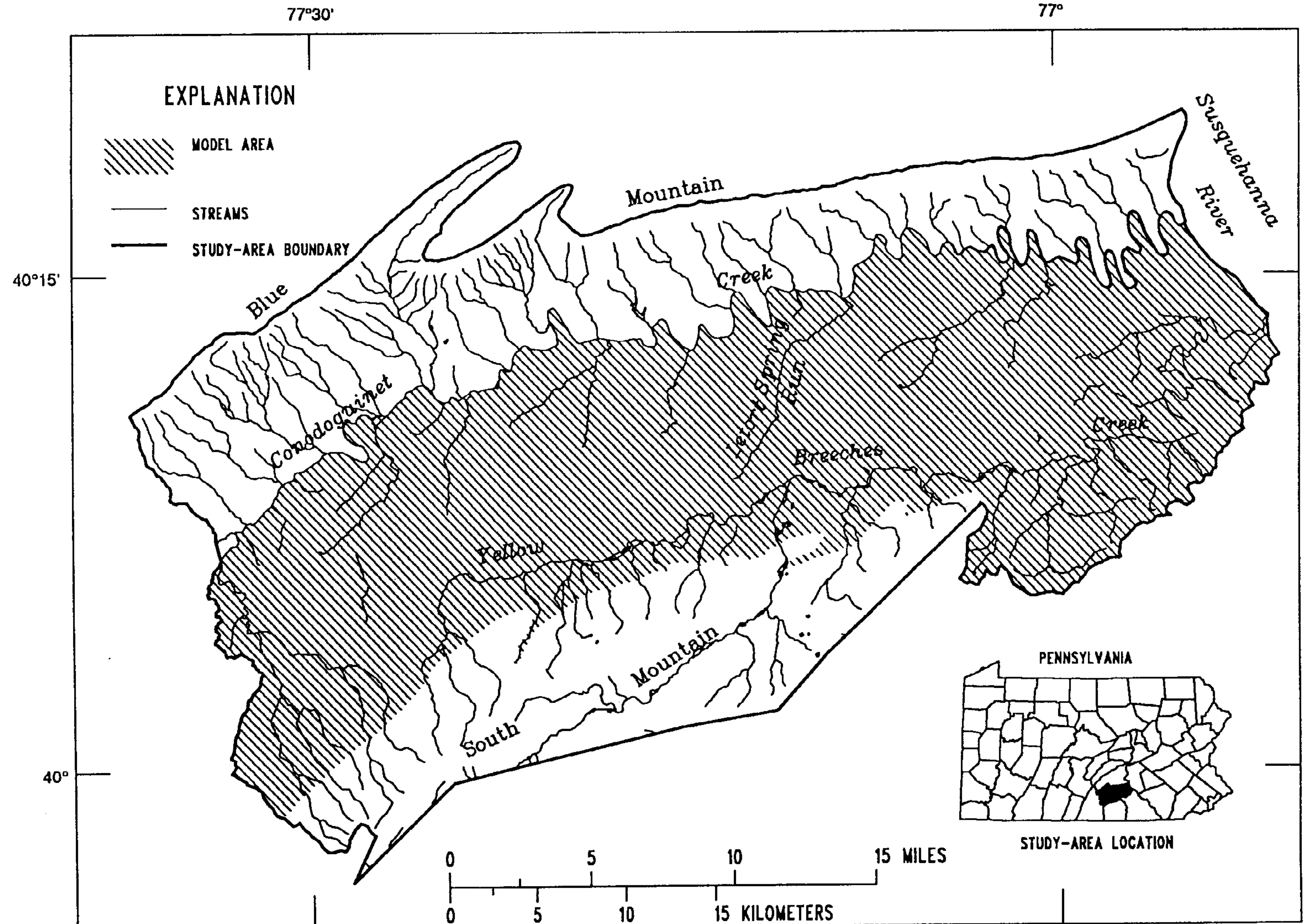


Figure 20. Location of study area.

valley.

Physiography

The study area is located in part of the Ridge and Valley, Piedmont, and Blue Ridge physiographic provinces. The northern and central part of the area is in the Great Valley section of the Ridge and Valley physiographic province. Cumberland Valley, which extends from the Pennsylvania-Maryland border to the Susquehanna River, is in the central part of the Great Valley section. South Mountain is in the Blue Ridge physiographic province. The eastern-southeastern part of the study area is in the Gettysburg-Newark Lowland section of the Piedmont physiographic province.

Previous Investigations

All or part of the study area has been the subject of several geologic and hydrologic investigations. The geology and hydrogeology of Cumberland County was described by Becher and Root (1981). Flippo (1974) and Saad and Hippe (1990) compiled and summarized discharge of measured springs within the study area. Gerhart and Lazorchick (1988) included the valley in the lower Susquehanna River basin ground-water-flow model. The geology of parts of Cumberland and York Counties was described by MacLachlan and Root (1966) and Root (1977, 1978). The geology of parts of Cumberland and Franklin Counties was described by Fauth (1968). Root (1968, 1971) described the geology of parts of Franklin County. The hydrogeology of Franklin County was described by Becher and Taylor (1982).

Data Availability, Collection, and Manipulation

The APRASA type-area studies rely primarily on existing data, with some supplemental data collection, to develop an understanding of the hydrogeologic characteristics of the local ground-water flow systems. In addition to published reports, a substantial data base exists for the Cumberland Valley study area. These data include USGS Ground-Water Site Inventory (GWSI) well and spring data, Pennsylvania Water-Well Inventory (PAWWI) data, continuous water-level records from wells, aquifer-test information, continuous streamflow records, precipitation records, and water-use records.

Field work for this study has focused on obtaining additional measurements of ground-water levels and ground-water discharge to streams and springs. Ground-water levels will be measured at selected wells during average ground-water conditions. These data will then be compared to water levels depicted in the map of the water table in November 1972 drawn by Becher and Root (1981, plate 1). Seepage-run data were collected to improve an understanding of the relation between surface water and ground water in the study area, to differentiate between ground-water recharge and discharge areas, and to calibrate the computer simulation with respect to the direction and magnitude of water flow through streambeds.

A Geographic Information System (GIS) is being used to compile, calculate, and store data; develop computer-simulation grids; input data to components in the model; and present simulation results. The following information is in the GIS data base: county, study-area, and model-area boundaries; hydrography; topography; bedrock geology; GWSI and PAWWI data; precipitation; thickness of unconsolidated and saturated regolith; hydraulic conductivity; seepage-run data; and model-grid and node data necessary for calibration and testing using the USGS ground-water-flow model MODFLOW (McDonald and Harbaugh, 1988).

HYDROGEOLOGIC SETTING

General Geology and Hydrology

The distribution and occurrence of general rock types for the area to be modeled are shown in Figure 21. The geology along the Conodoguinet Creek and to the north, is characterized by shales and graywacke, and resistant sandstone that forms Blue Mountain (not shown in Figure 21). Between the Conodoguinet Creek and South Mountain, carbonate rocks prevail, with local argillaceous carbonates, calcareous shales, and shales. In the eastern third of the area, a diabase dike is exposed that trends north-south. (Note that the width of the dike shown in Figure 21 is exaggerated for modeling purposes.) In the eastern-southeastern part of the area, the geology is characterized by red sedimentary rocks and diabase intrusives (dikes and sills). To the south, the geology is characterized by resistant quartzite and schist that form South Mountain (not shown in Figure 21).

Median sustained yields of water from rocks in the area differ greatly-- from less than 10 gal/min (gallon per minute) (Root, 1977; 1978) from the diabase intrusives in the east and east-southeast, to greater than 1,000 gal/min (Becher and Root, 1981) from cavernous dolomite underlying the regolith mantle along the flank of South Mountain. Reported median specific capacity is 0.15 to 1.4 (gal/min)/ft (gallon per minute per foot of drawdown) for shales, siltstones, and graywacke (Becher and Root, 1981). The low values are indicative of shale with few joints, fractures, and bedding-plane separations, whereas the higher values are indicative of calcareous shale or graywacke with extensive primary and secondary porosity and permeability. Median specific capacity of carbonate rocks range from 0.20 (gal/min)/ft for argillaceous limestone to 19 (gal/min)/ft for cavernous dolomite (Becher and Root, 1981).

Along the northern flank of South Mountain, an extensive deposit of regolith has formed on the carbonate rocks of the valley bottom. The regolith is a wedge-shaped deposit, as thick as 450 ft (feet)(Becher and Root, 1981), consisting of residual material, alluvium, and colluvium. The residual material, mainly insoluble clastic material from weathered carbonate rocks, ranges from 170 (Pierce, 1965) to 200 ft thick (R.S. Hughes, Gannett Fleming, written commun., April 1991). Becher and Root (1981) have indicated that the chemically aggressive water flowing off

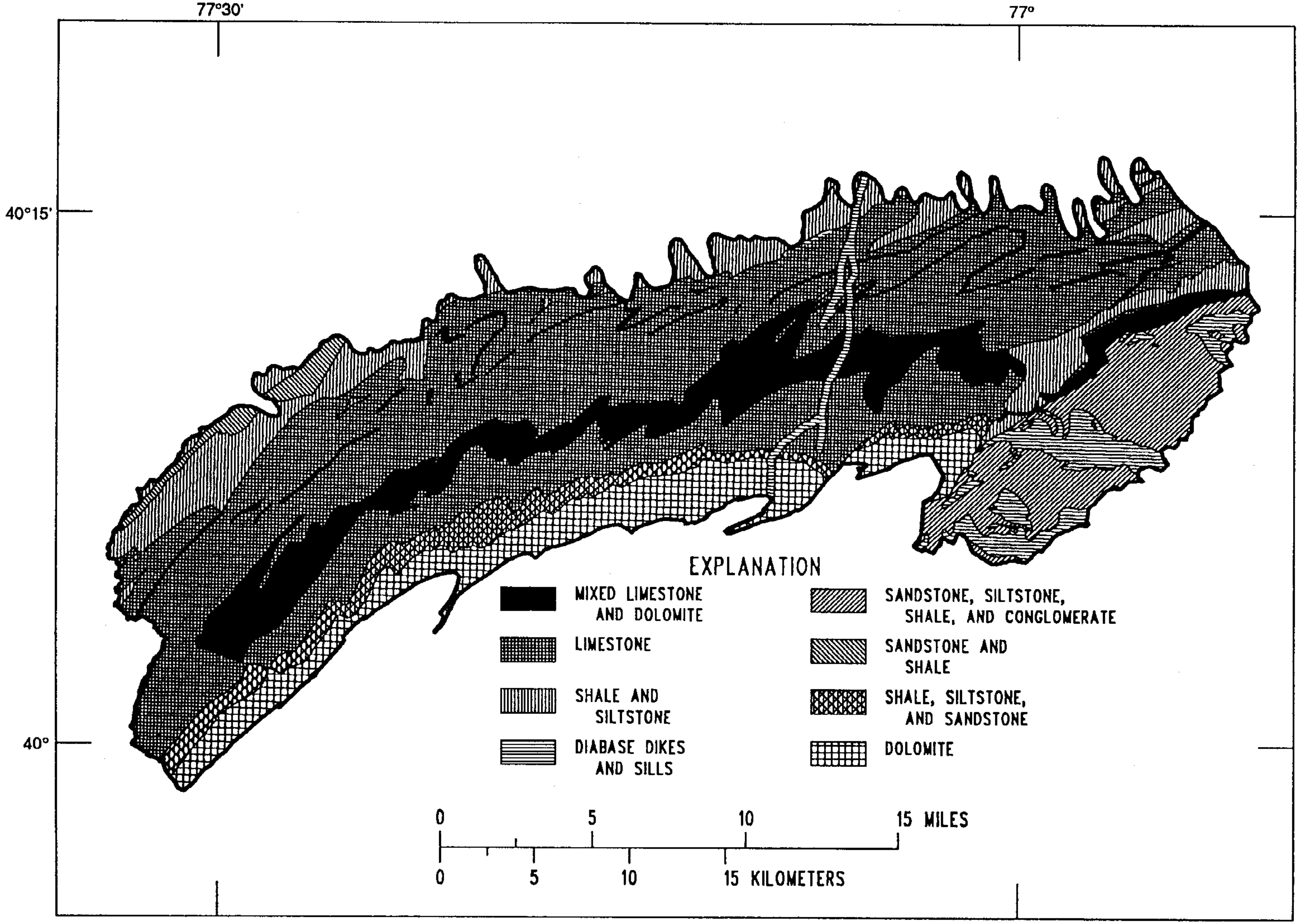


Figure 21. General rock types of the study area (modified from Berg, Sevon, and Abel, 1984).

South Mountain has dissolved the carbonate rocks adjacent to South Mountain and produced the topographically low area presently occupied by the Yellow Breeches Creek.

Thick deposits of alluvium and colluvium overlie the residual material. The alluvium consists of floodplain and alluvial-fan deposits that have coalesced to form thick alluvial slopes. Alluvial deposits are composed of detrital debris and siliciclastic material derived from South Mountain. The alluvial deposits can be as thick as 200 to 300 ft in buried river channels incised in the carbonate rocks (Root, 1978). Colluvial deposits are interspersed in the alluvium and are composed of similar, yet coarser, siliciclastic material. The colluvial deposits can be greater than 100 feet thick in areas near the source material along South Mountain (Root, 1978).

The water-yielding capacity of the regolith varies according to its composition. Becher and Root (1981) reported a median specific capacity of 1.4 (gal/min)/ft for colluvium, and Hollyday and others (U.S. Geological Survey, written commun., Feb. 1991) reported a value of approximately 10 (gal/min)/ft for alluvium throughout the Ridge and Valley physiographic province.

A generalized north-south hydrogeologic section of the study area is shown in Figure 22. This figure shows the general distribution of bedrock and regolith for an area typical of the central part of the study area.

Gains and Losses in Streamflow

Seepage-run data were collected during a period of "average" streamflow at selected reaches of the Yellow Breeches and Conodoguinet Creeks, their major tributaries, and springs. These data are used to quantify the groundwater discharge from the aquifer as well as to determine areas of losing and gaining reaches along the streams. In order to schedule the seepage investigation, median streamflow conditions were determined statistically by hydrograph-separation techniques (Pettyjohn and Henning, 1979) at three USGS continuous-record gaging stations (Figure 23) (USGS Station IDs: 01571500 - Yellow Breeches Creek near Camp Hill, Pa.; 01570000 - Conodoguinet Creek near Hogestown, PA; and 01569800 - Letort Spring Run near Carlisle, PA). These median base flows were compared to existing stream stages at the gaging stations to select the days when median ground-water discharges could be measured. On June 13 and 14, 1990, seepage-run measurements were made at 81 sites in the study area (Figure 23), and were published by the USGS (U.S. Geological Survey, 1991, p. 194-199).

Results of seepage runs indicate that most stream reaches in the lower and middle part of the Yellow Breeches and Conodoguinet Creek basins are gaining water from the ground-water reservoir, which maintains streamflow during average ground-water-discharge conditions. Data collected for tributaries along the northern flank of South Mountain indicate that many reaches are losing water to the ground-water reservoir. Indeed, some of these reaches go dry and only regain water in the lower reaches of the tributary. These reaches lose water because of several factors:

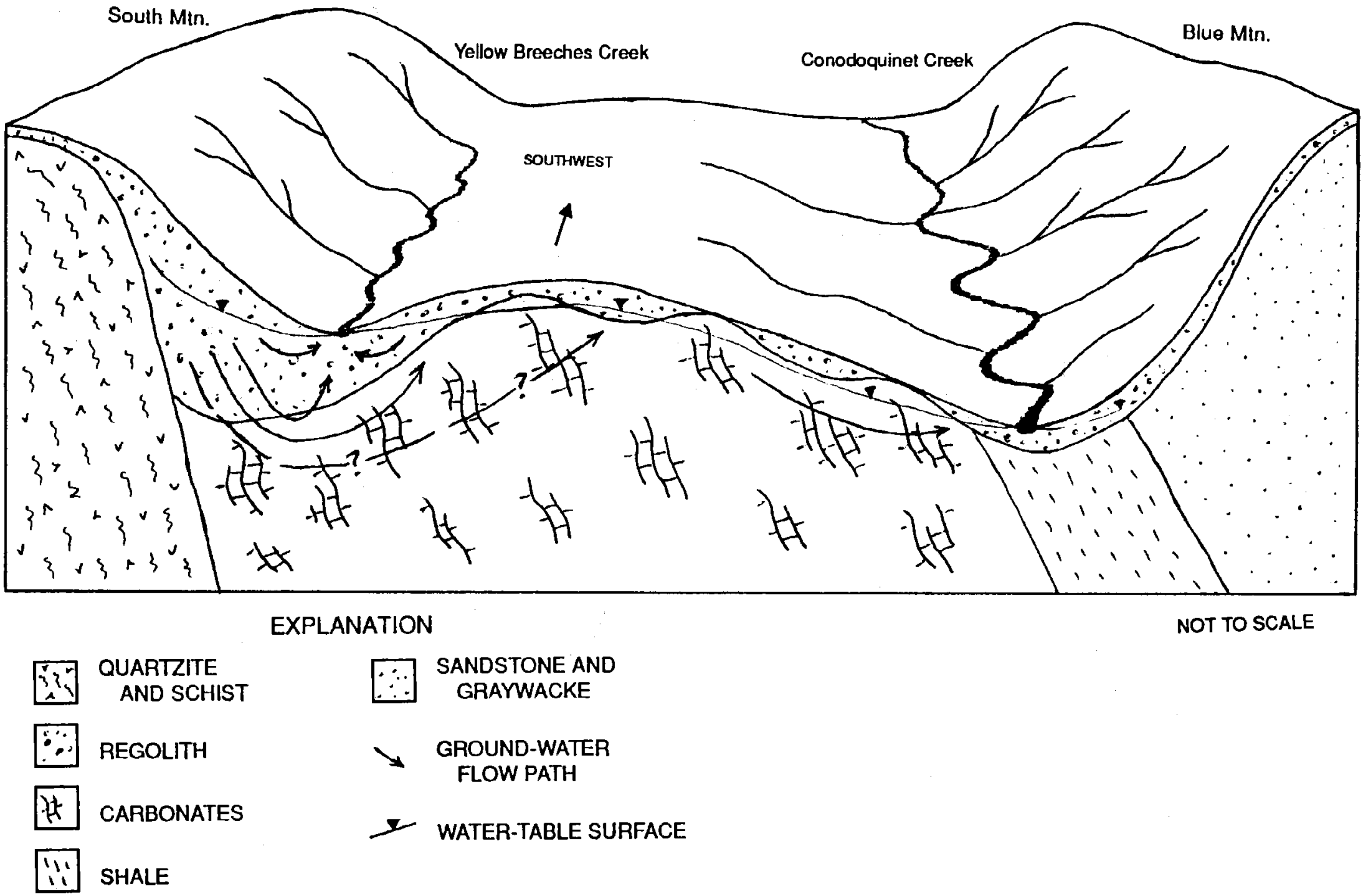


Figure 22. Generalized block diagram and hydrologic cross section of the study area.

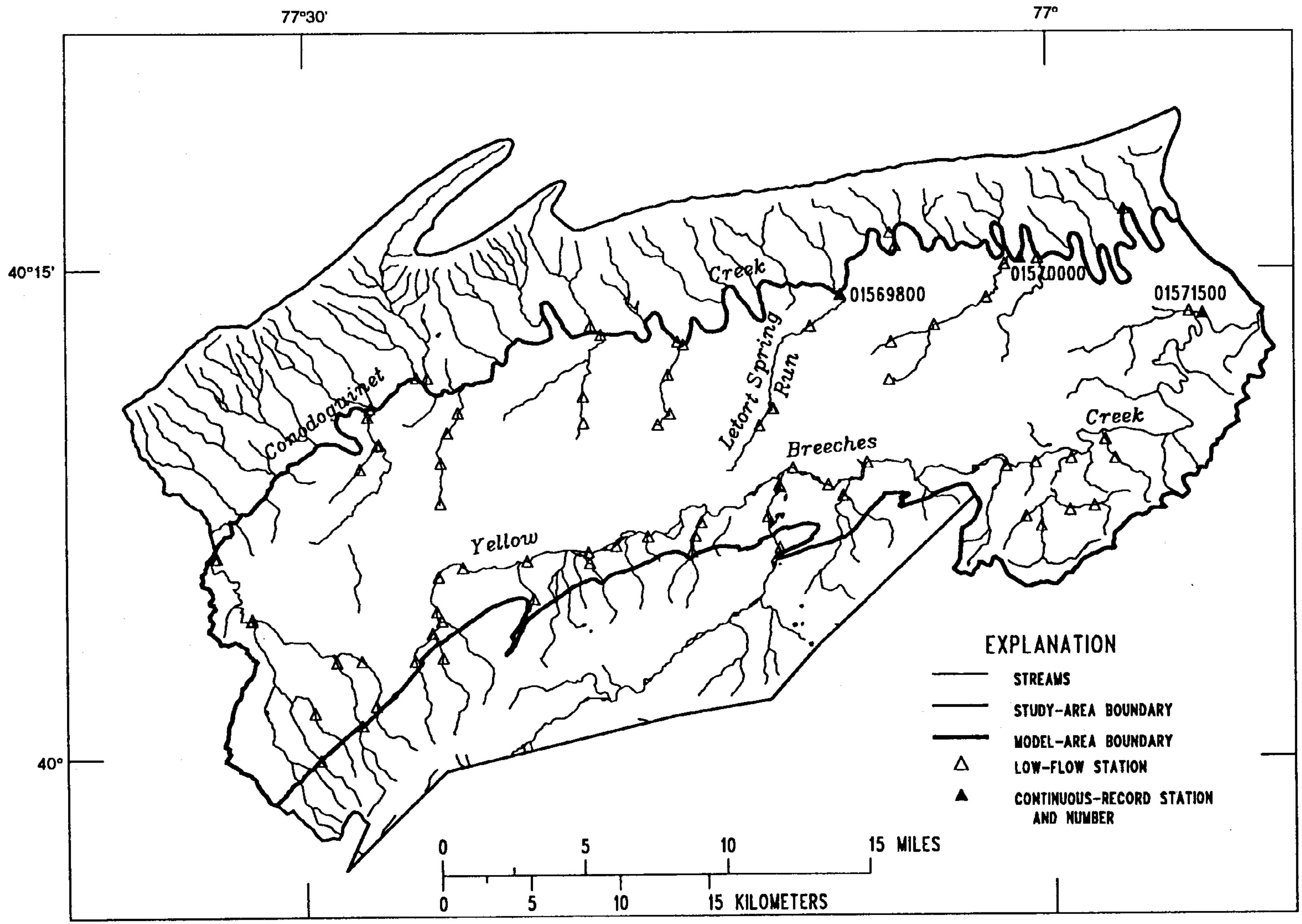


Figure 23. Location of streamflow-gaging stations.

Tributary streams flowing near hilltops of South Mountain have steeper gradients and flow over material of lower permeability (quartzite and schist) than do the streams near the valley walls and adjacent to the valley bottom; as a result, hilltop streams lose little or no water to the ground-water reservoir. When the tributary streams flow over the valley wall areas, the low gradients and high streambed permeability (regolith), enable infiltration and percolation of surface water to the ground-water reservoir (Figure 22).

CONCEPTS OF GROUND-WATER FLOW

A water-table map of the area was constructed to help determine recharge and discharge areas, conceptualize ground-water flow, determine the effects of geology on the water table, and improve the accuracy of computer simulations (Figure 24). The map was constructed, in part, from the map of the water table in November 1972 by Becher and Root (1981, plate 1). In areas outside that mapped by Becher and Root, the water-table map was completed with GWSI and PAWWI data from a period of average ground-water conditions, and from streamsurface elevations at gaining-stream reaches. During average ground-water conditions, ground-water levels will be measured in select wells to determine how much, if any, the present-day water levels have changed from the surface mapped by Becher and Root (1981).

Horizontal Component of Flow

The water-table configuration and gradient are strongly related to the underlying geology and reflect general topography of the land surface. In areas where the bedrock has low porosity and permeability (e.g., shale, diabase, or argillaceous limestone), the gradients increase and closely follow the areal distribution of the particular rock type. Conversely, in areas where bedrock has high porosity and permeability (e.g., limestone or dolomite), the gradients decrease.

The eastern-southeastern part of the area is underlain by diabase sills and red sedimentary rocks, including sandstone, siltstone, shale, and conglomerate (Figure 21). These rocks are characterized by low porosity and permeability and, in this area, have little secondary porosity. The water table in areas underlain by these rocks (Figure 21) has gradients greater than 25 ft/1,000 ft (Figure 24).

In the east-central part of the area, a diabase dike trends north-south through the valley. The water-table configuration shows a displacement and damming effect as the ground water tries to move around, over, and(or) through the dike (Figures 21 and 24).

In the northwestern part of the area, the water table contours reflect mounds of ground water in two places (Figure 24). These mounds overlie areas of shale bedrock, whose hydrologic characteristics contrast significantly with those of the surrounding carbonate rocks. Also, northeast of the diabase dike and in the east-central part of the study area, two other areas

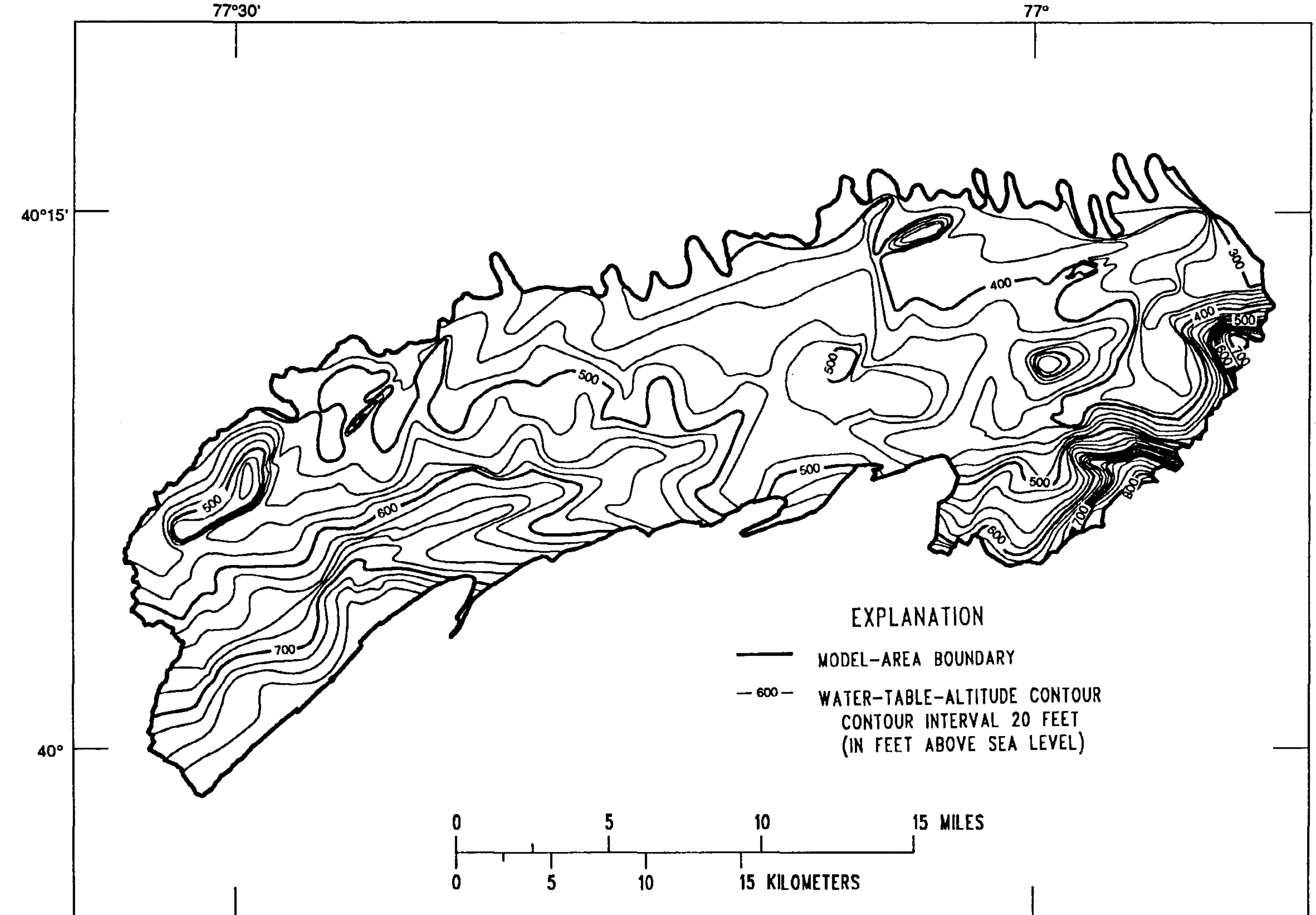


Figure 24. Altitude of water table within the model area (modified from Becher and Root, 1981, Plate 1).

of shale have resulted in mounding of the ground water.

In the area of the Letort Spring Run, the underlying geology is characterized by limestone and dolomite that has been developed by solution. The water-table configuration in this area has low relief with gradients of approximately 4 ft/1,000 ft.

In the southern part of the study area, along the flanks of South Mountain, the water table has gradients of approximately 10 ft/1,000 ft. This gradient reflects not only the topography of the valley walls, but also the large amount of recharge resulting from infiltration of precipitation and water from losing streams.

Vertical Component of Flow

Along the flank of South Mountain, the ground-water system is recharged predominantly from losing streams and precipitation. In areas southeast of Shippensburg, the water-table aquifer consists of saturated regolith as thick as 240 ft that overlies cavernous dolomite. From the saturated regolith, flow is downward to the underlying carbonate-rock aquifer, then laterally and upward to springs and streams (Figure 22).

In the center of Cumberland Valley, between the Yellow Breeches and Conodoguinet Creeks, the ground-water system is recharged largely from infiltration of precipitation. Here, the aquifer is predominantly in carbonate rock, because the regolith thins northward toward Conodoguinet Creek, and locally is either unsaturated or discontinuous. Within the carbonate rock, ground-water flow is through joints, fractures, bedding-plane separations, and cleavage openings that have been enlarged by dissolution of the folded and faulted carbonate rocks.

Becher and Root (1981) have indicated that there is intrabasin and interbasin flows of ground water in the valley. Within the Yellow Breeches Creek basin, ground water infiltrates into the reservoir south of the creek, flows under the creek, and discharges to springs north of the creek (Figure 22). In addition, Becher and Root indicated that interbasin flow of ground water occurs from Yellow Breeches to the Conodoguinet Creek.

WATER BUDGETS

Mean annual precipitation at five National Oceanic and Atmospheric Administration stations in or adjacent to the study area ranges from 38.8 to 46.4 in. (inches) and averages approximately 40 in. (see also Table 6). Typically, because of orographic effects, precipitation amounts are greater on ridges and hilltops than in the valley bottoms. Precipitation is uniformly distributed throughout the year except during summer months when precipitation amounts increase slightly because of local storms.

The water budgets for the study area (Table 14) were determined from precipitation data and use of stream-hydrograph-separation (Pettyjohn and Henning, 1979) and recession-curve displacement techniques (A.T. Rutledge, U.S. Geological Survey, written commun., Feb. 1991) for the period of record for each of three station. The data shown in Table 14 result from hydrograph separation. These data are very similar to those calculated by

Table 14. Major components of water budgets for Yellow Breeches Creek, Conodoguinet Creek, and Letort Spring Run. Percent = percent of total precipitation; mi^2 = square miles.

	Yellow Breeches Creek		Conodoguinet Creek		Letort Skpring Run	
	inches/year	percent	inches/year	percent	inches/year	percent
Surface runoff[1]	3.5	9	5.7	14	1.6	4
Ground-water Discharge[1]	15	38 (81)*	12	30 (68)*	23	57 (93)*
Evapotranspiration	21	53	22	56	16	39
Precipitation	40	100	40	100	40	100
Basin area (mi^2)	216		470		21.6	

*Base-flow index, or ground-water discharge as percent of total streamflow.
[1]A. T. Rutledge, written communication, February, 1991.

Becher and Root (1981) using methods of Rorabaugh (1964).

The data in table 1, in particular the base-flow index, reflect the different lithologic and topographic characteristics of each surface-water basin. The Letort Spring Run base-flow index of 93 percent reflects a valley basin in carbonate terrane of low relief (approximately 200 ft). In this basin, only 7 percent of the flow is surface runoff; the remainder is ground-water discharge. The ground-water system is drained predominantly by solution-enlarge conduits in the carbonate rock.

The Conodoguinet Creek base-flow index of 68 percent reflects a basin in carbonate and shale terrane with high relief (approximately 1,900 ft). In this basin, nearly one third of the total flow is surface runoff; the remainder is ground-water discharge. Drainage of ground-water is more diffuse in this basin than in the Letort Spring Run basin.

The Yellow Breeches Creek base-flow index of 81 percent reflects a basin in quartzite, schist, and mantled-carbonate terrane with a basin relief slightly less than that of the Conodoguinet Creek basin (approximately 1,700 ft). In the Yellow Breeches Creek basin, the saturated regolith provides a slow, steady release of water to the stream as base flow. Surface runoff is only about 19 percent of the total flow; ground-water discharge comprises about 81 percent of the streamflow. The ground-water system is predominantly drained by diffuse flow through a porous media and solution-enlarged openings.

SUMMARY AND CONCLUSIONS

The regolith-mantled carbonate-rock aquifer of Cumberland Valley is a highly productive and complex aquifer. The aquifer is characterized by complexly folded and faulted carbonate bedrock in the valley bottom and, locally in the north and east, by shale, graywacke, and red-sedimentary and diabase rocks. Near the southern valley hillslope, the carbonate rock is overlain by wedge-shaped regolith (as thick as 450 ft) consisting of residual material, alluvium, and colluvium. Residual material, comprised mostly of weathered carbonate rock, is up to 200 ft thick. Alluvium and colluvium consist of siliciclastic materials derived from a resistant upland source of quartzite and schist to the south. Locally, the thickness of regolith that is saturated exceeds 240 ft. The topographic relief of the carbonate, carbonate and shale, and regolith-mantled carbonate basins are approximately 200, 1,900, and 1,700 ft, respectively.

Anomalies in water-table gradient and configuration are a result of topography and differences in the character and distribution of overburden material and bedrock. In general, the water-table surface is a subdued representation of the land surface. Ground-water flow is generally to the north and east to the Susquehanna River; locally, however, ground water is mounded, has steep gradients, and its flow is diverted by rocks of low permeability. In areas of solution-affected carbonates, the water-table gradient is low.

Seepage-run data indicate that stream reaches near valley walls are losing water from the stream, through the regolith, to the ground-water system. Most stream reaches in the lower and middle part of the Yellow Breeches and Conodoguinet Creek basins are gaining water from the ground-water system. Results of hydrograph separations indicate that base flow in stream basins dominated by carbonate, carbonate and shale, and regolith-mantled carbonate bedrock is 93, 68, and 81 percent of total streamflow, respectively.

Analysis of the McNeal ore.

Metallic iron,	46.250
Metallic manganese,	.432
Sulphur,	.033
Silicious matter,	15.940
Phosphorus,	.437
Phosphorus in 100 parts iron,	.944

Pond bank No. 1.

This is the first opening north from the railroad, formerly worked as an open cut, 80 feet deep, but now idle and filled with water to within 30 feet of the surface. There is absolutely nothing to be seen at this opening, although the ore taken from it in past times must have been large and is said to have been of a superior quality. The stripping varies from 10 to 30 feet.

Between this opening and the Little Pond bank to the north a shaft was put down by the Mont Alto Iron Company, with the following results:

Earth and white clay,	10 feet.
Sharp light-colored sand,	5 "
Clay, sand and pigment,	25 "
Black close-grained clay,	1 "
Lignite,	4 "
Gray sandy clay,	1 "
Lignite,	18 "
Sand,	1 "
Variegated clay,	6 "

The record of this shaft 71 feet deep was kindly furnished by Col. George Wiestling, and is certainly very interesting on account of the occurrence of the two beds of lignite. From the bottom of the shaft drifts were driven towards the two banks for the purpose of draining them.

Little Pond bank No. 2.

This lies about 200 yards north of the railroad, close to the base of Little mountain. It was formerly worked as a large open cut, although that method has now been abandoned for underground mining. Gangways have been driven southwards from the bottom of the shaft almost to the railroad, which is evidence of the flat structure of the ore-mass,

From d'Invilliers, 1887.

THE ACIDIFICATION OF STREAMS AND LAKES IN THE SOUTH MOUNTAIN REGION: TWO CASE STUDIES

by
Candie C. Wilderman
Department of Geology, Dickinson College, Carlisle, PA
Robert J. Schott
PA Department of Environmental Resources, Harrisburg, PA

INTRODUCTION

Streams and reservoirs throughout the Blue Ridge geological province have been shown to be particularly vulnerable to acid deposition, due to the extremely low buffering capacity of the bedrock and soils in the region. This report summarizes the results of studies in the South Mountain regions of Cumberland and Franklin County, documenting the acidification of two streams and one drinking supply reservoir (Figure 25).

ACID DEPOSITION: THE PROBLEM

In recent years, thousands of scientific studies have documented the serious impacts of acid deposition in North America and in Europe. The mounting evidence of the harm being done by acid deposition was summarized recently in a report on the status of the scientific knowledge about acid deposition (Camuto, 1991). The report, published jointly by seven prominent scientists, noted that the scientific community now understands the sequence of physical and chemical steps that lead from the emission of oxides of sulfur and nitrogen in North America to eventual changes in widespread ecosystems. The authors also noted that the biological consequences of high rates of sulfur deposition on aquatic ecosystems are causing widespread damage, including loss of bicarbonate, increased acidity, and higher concentrations of toxic metals, all of which have resulted in several important species of fish and invertebrates having been eliminated over substantial parts of their natural ranges. Other less well documented effects include high forest mortality, accelerated corrosion of buildings and monuments, alteration of agricultural yields, and health impacts through degradation of drinking water quality.

Results from a six-year monitoring study by the National Atmospheric Deposition Program (National Atmospheric Deposition Program, 1989), based on hundreds of sites across the nation, indicate that the Northeast receives the highest sulfate and nitrate deposition, as well as acidic rain of any region in the nation. In fact, studies by the Pennsylvania State University Monitoring Network (Lynch and others, 1986, 1987) have demonstrated that Pennsylvania receives the most acidic deposition of any state in the nation. Monitoring stations throughout the Commonwealth reveal that the pH of our rainfall averages approximately 4.0, having declined from an estimated pre-industrial pH of approximately 5.0. Other studies (Webb and others, 1989a; 1989b) have noted that ion deposition in precipitation has increased as much as ten-fold from pre-industrial times, primarily in sulfate,

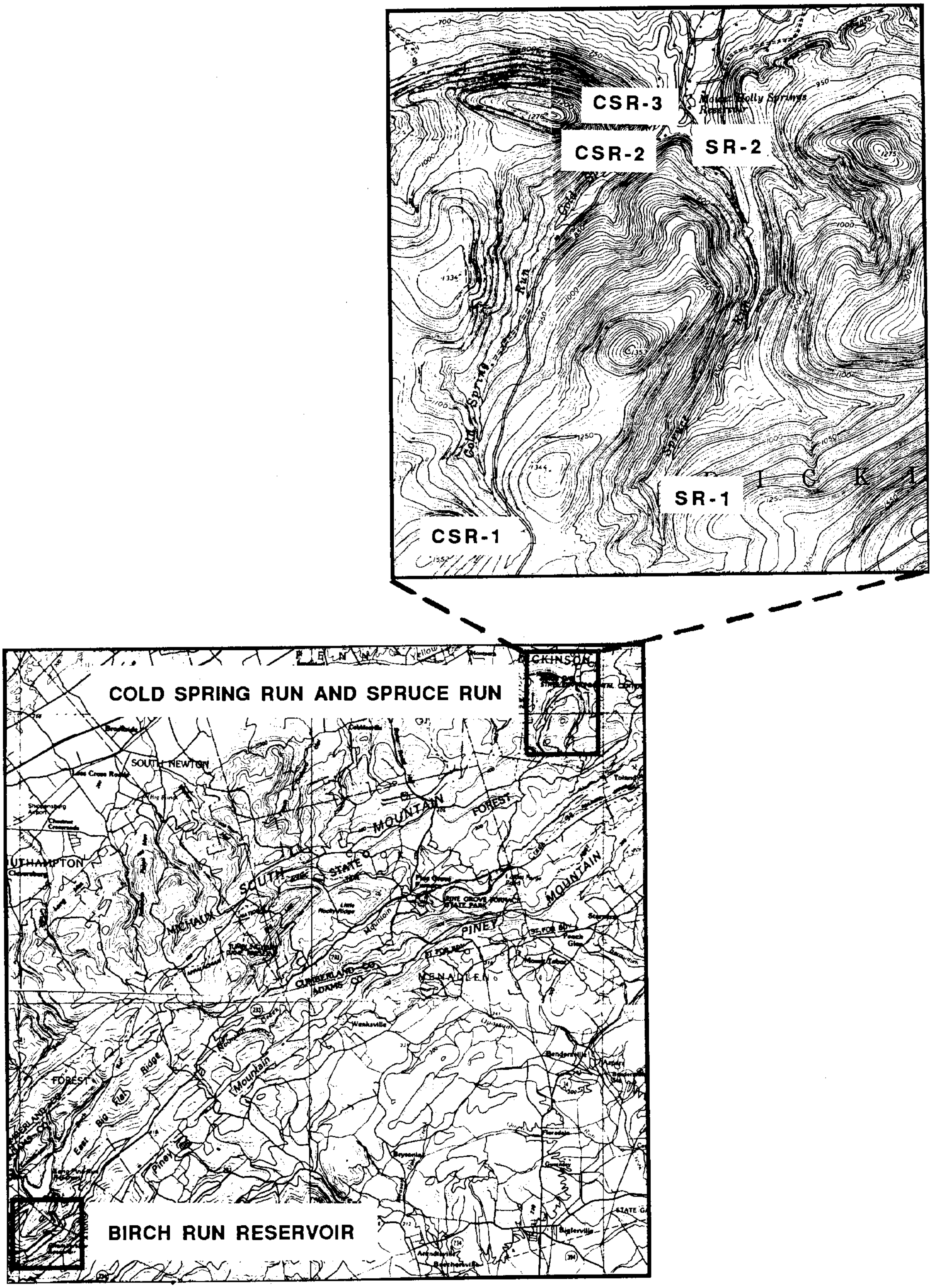


Figure 25. Location map, showing the Birch Run Reservoir and the Cold Spring Run and Spruce Run watersheds. Inset shows the sampling site locations on Cold Spring Run and Spruce Run. hydrogen, and nitrate ions.

Additional studies are documenting the threat to Pennsylvania's streams and lakes. The Pennsylvania Fish Commission reported in 1986 (Johnson, 1986) that 30 percent of the state's 5000 miles of stocked trout streams and 45 percent of its 9000 miles of unstocked trout water were vulnerable to acidification (alkalinity <10 mg/l). They projected the loss of 3000 miles of trout water by the end of the century. Since the late 1950's more than 80 streams have been subject to trout stocking management changes as a result of increasing acidity. The United States Environmental Protection Agency's (EPA) National Stream Survey (NSS-I), where 500 stream reaches in 17 states were samled, identified more acidic streams in Pennsylvania than any other state in the study (Kaufmann and others, 1988).

Studies on data collected from streams across the Commonwealth by trained citizen volunteers, as part of the Alliance for Acid Rain Monitoring (ALLARM) project at Dickinson College (Wilderman, 1990; Wilderman and Reuss, 1991), reveal that episodic pH and alkalinity depressions due to rainfall events are common in Pennsylvania, even in streams that are categorized as "resistant", based on their average annual alkalinity. (Figures 26 and 27). A Lehigh University study (Bradt and Wimpfheimer, 1987) shows that out of 160 lakes in the Pocono region for which there are data, 70 percent were sensitive to acid deposition and 8 percent were already acidified. A subset of small, high altitude lakes (greater than 500 m in elevation and less than 20 ha in surface area) showed that 96 percent were sensitive and 30 percent were already acidified. Finally, published findings on streams in the Laurel Hill area of southwestern Pennsylvania (Sharpe and others, 1984) blame acidification for the absence of trout in 20 percent of 61 headwater streams and changes in benthic macroinvertebrate communities.

IMPORTANCE OF GEOLOGY IN DETERMINING STREAM RESPONSE TO ACID DEPOSITION

Different regions may respond differently to acid deposition, depending on the region's natural ability to "buffer" or neutralize the incoming acidity. The single most important factor in determining stream response to acidic input is watershed bedrock type.

A stream's sensitivity to acid deposition is generally measured by its alkalinity concentration, that is, how much acid neutralizing capacity (ANC) it has. Although different authors vary somewhat in their category definitions (Figure 28), a stream is generally considered sensitive if its average alkalinity is less than 10 mg/l (200 ueg/L). It is extremely sensitive and vulnerable to acidic episodes associated with heavy rainfall and snowmelt if its alkalinity is less than 5 mg/l (100 ueg/l) and it is acidic if its alkalinity is less than 0 mg/l (0 ueg/l). It is important to note that the ANC of any stream is finite and irreplaceable, because it is derived from soils. Generally streams in carbonate regions have high buffering capacities, whereas streams associated with basaltic bedrock are marginally sensitive; those which flow over granitic bedrock are sensitive and

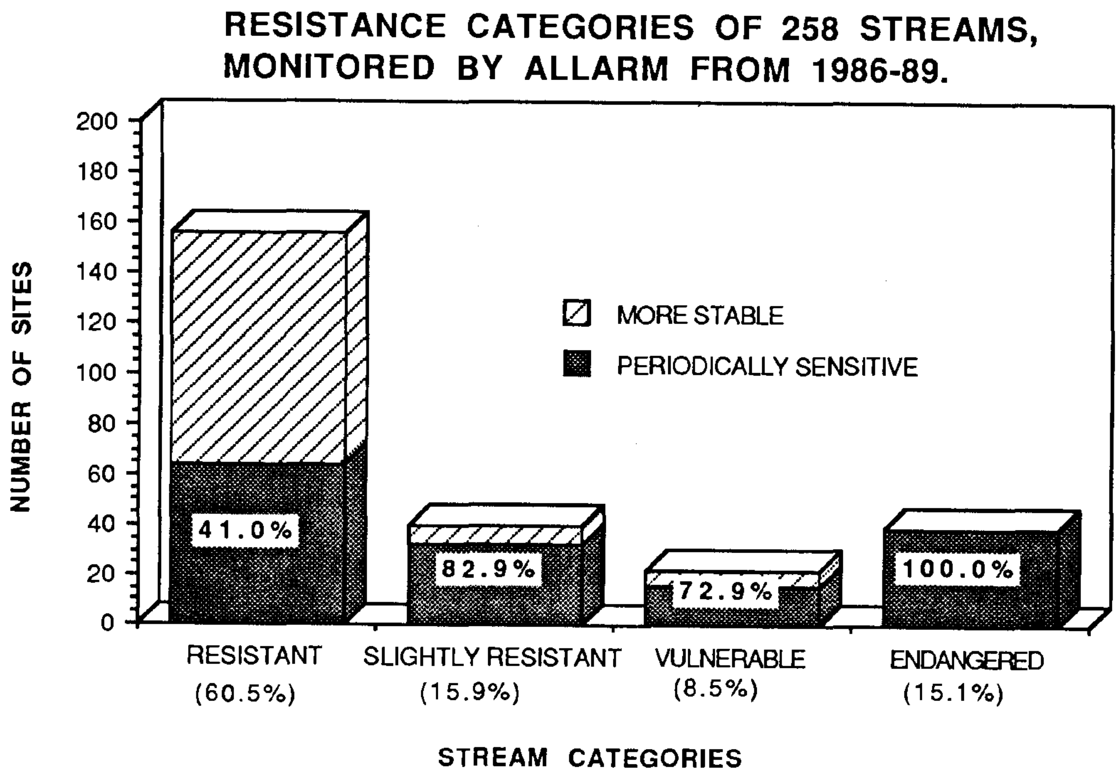
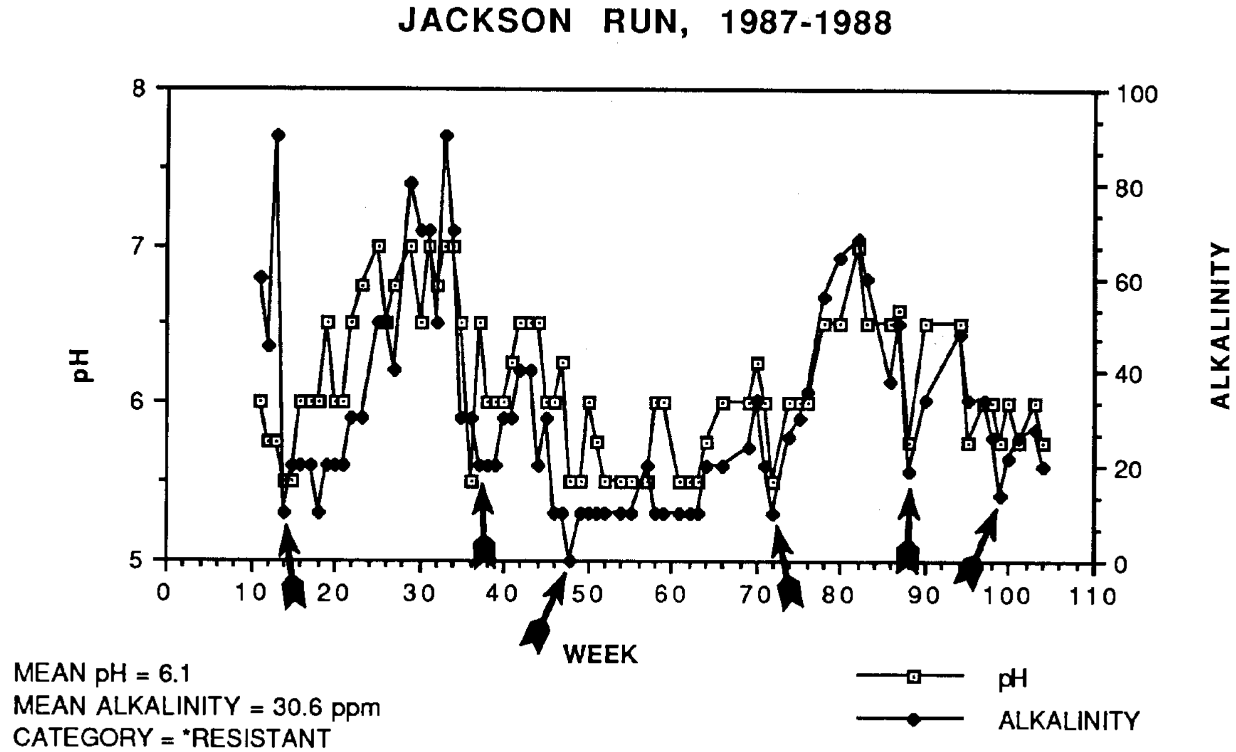


Figure 26. Data from 258 streams in Pennsylvania, indicating that a large percentage of streams in all resistence categories undergo periods of sensitivity in response to rainfall events.

those in watersheds dominated by siliciclastic bedrock (quartzite, sandstone and phyllitic shale) are extremely sensitive (Webb and others, 1989a).

Studies on 91 streams monitored biweekly or weekly in 1989 by the ALLARM project, show that there were significant differences in the number of endangered and vulnerable streams in the four geologic provinces, with Blue Ridge having the highest percentage, followed by Allegheny Plateau, Valley and Ridge, and Piedmont respectively (Figure 29). EPA also recognized in its NSS-I study (Kaufmann and others, 1988) that low pH and low ANC stream reaches were most common in upland, forested sites of the Allegheny Plateau, Valley and Ridge Province, Blue Ridge Mountains, and Cumberland Plateau. They singled out the Southern Blue Ridge as having the highest proportion of streams sensitive to acidification and they noted the high vulnerability of streams in watersheds underlain by metamorphic rock. Studies on streams in the Shenandoah National Park found that watersheds dominated by siliciclastic bedrock such as quartzite, sandstone, and phyllitic shale were most sensitive to acid input. A more extensive survey of steam water chemistry (Webb and others, 1989b) corroborated these regional differences.

In addition, whether or not the region has been glaciated plays a critical role in the rate of release of sulfate into the



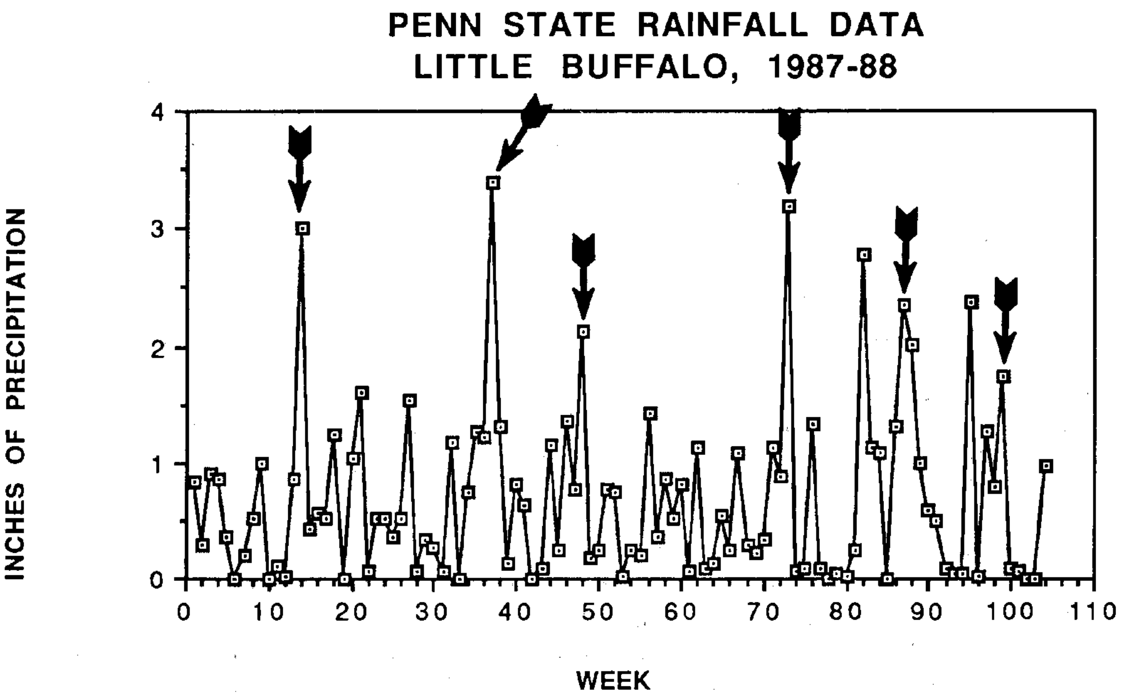


Figure 27. Plots of weekly alkalinity and pH determinations for a monitoring site on Jackson Run in Perry County, sampled by ALLARM volunteer Edith Brown, compared to weekly rainfall amount (inches) at the closest Pennsylvania State University Deposition Monitoring Station. Arrows indicate weeks of heavy rainfall and depressed alkalinity.

ALKALINITY (mg/l)	CANADA/US 1983 MEMO	ALLARM 1988
		RESISTANT
20		------
19		
18	RESISTANT	
17		
16		
15		SLIGHTLY
14		RESISTANT
13		
12		
11		
10	------	------
9		
8		
7		VULNERABLE
6	MODERATELY	
5	SENSITIVE	------
4		
3		
2	------ EXTREMELY	ENDANGERED
1	SENSITIVE	
0	------	------
	ACIDIFIED	ACIDIFIED

Figure 28. Two closely related sensitivity category schemes. The first is based on the Canadian/US 1983 agreement and the second is a scheme commonly used by other groups such as ALLARM.

aquatic system. Unglaciated regions generally have a deeper, older soil cover which adsorbs sulfate and provides a temporary reprieve for the streams. However, this provides only a delay in the acidification process; as the adsorption capacities of watershed soils are exhausted, the sulfate concentrations and acidity levels in the streams will rise (Cosby and others, 1986).

CASE STUDIES IN THE SOUTH MOUNTAIN REGION

(1) Cold Spring Run and Spruce Run

Cold Spring Run and Spruce Run are first order streams

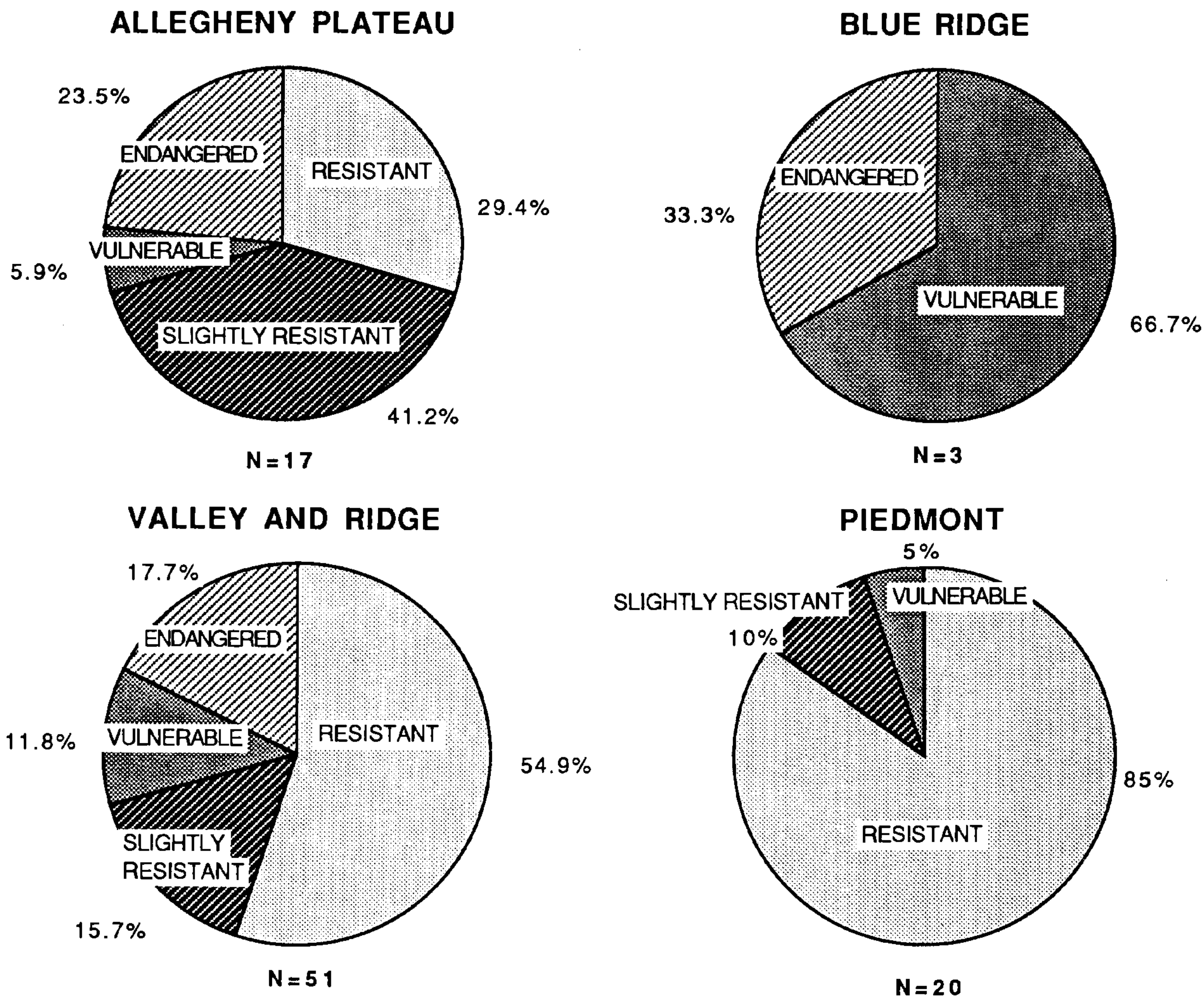


Figure 29. Percentage of streams in each resistance category in each of four geologic provinces in Pennsylvania. Data were collected by ALLARM volunteers and is based on weekly monitoring of 91 streams.

originating on South Mountain in Dickinson Township, Cumberland County (Figure 25). The two streams converge and flow into a small reservoir used by the Borough of Mount Holly Springs as a public water supply. Both streams are similar in size at their confluence, having widths of approximately 1 m and depths of approximately 10 cm. The majority of flow in Cold Spring Run originates in one large spring, whereas Spruce Run receives its flow from numerous small springs and seeps along the valley through which it flows.

Both drainage areas are underlain predominantly by the Mont Alto Member of the Harpers Formation, which is a medium-grained quartzite with interbeds of dark greenish-gray phyllite and schist. The upper portion of Cold Spring Run, above its confluence with Spruce Run, travels directly along the Cold Springs Fault, an east-dipping steep thrust (Root, 1970). Below the confluence and the Mount Holly Springs reservoir, Cold Spring Run travels across the colluvium on the flank of South Mountain, to its confluence with the Yellow Breeches (Becher and Root, 1981).

The upper reaches of both streams lie within the Michaux State Forest. The forest is a mixed oak-hard pine with chestnut oak (*Quercus promus*), scarlet oak (*Q. coccinea*), red oak (*Q. rubra*), black oak (*Q. velutina*), and pitch pine (*Pinus rigida*) dominating the community. In the lower reaches above the reservoir, white pine (*Pinus strobus*) and eastern hemlock (*Tsuga canadensis*) become more dominant.

Residents in the area report catching brook trout in both streams until the early 1970's when the fishery simply disappeared (Walter Mohn, personal communication). Unfortunately, no historical data exist for the two streams.

Biweekly samples were collected from three sites on Cold Spring Run and two sites on Spruce Run from January, 1985 to December, 1986 (Figure 25). Analyses for total aluminum, dissolved aluminum, and alkalinity were performed at the Department of Environmental Resources laboratories; pH and temperature were measured in the field using an Orion Model 399A pH meter. Qualitative sampling of the benthic community was conducted and fish communities were surveyed using electrofishing in April, 1985.

Figure 30 shows the alkalinity, pH and aluminum concentrations at Spruce Run (site #2) during the two years of the study period, as well as the amount of precipitation in the closest Penn State University Atmospheric Deposition Monitoring station. Alkalinity concentrations never exceeded 2 ppm, indicating a low buffering capacity. pH and aluminum concentrations are generally inversely related and all measurements are closely correlated to regional rainfall events. Figure 31 shows the alkalinity, pH and aluminum concentrations at Cold Spring Run (site #2) during the two years of the study period, as well as the amount of precipitation in the closest Penn State University Atmospheric Deposition Monitoring station. Although pH values were generally higher than in Spruce Run, relationships between variables and responses to rainfall events were similar.

There is a strong and rather consistent seasonal pattern in both streams, involving high pH's in January followed by a steady decline through February and March, a gradual increase from April to late September, with a dramatic decrease from October through December. Similar seasonal patterns were noted in about 2/3 of streams monitored on a weekly basis throughout the state of Pennsylvania by the ALLARM project (Wilderman, 1990; Wilderman and Reuss, 1991) in 1988 and 1989. Only those streams which were extremely resistant showed little seasonality.

Superimposed on patterns of seasonality are responses to extreme rainfall events. During such events pH and alkalinity reached temporarily low concentrations and aluminum levels increased. Examples for 1985 are marked with arrows in Figures 30 and 31 and include heavy rainfall events in early February, early April, early May, late September and early November, 1985. Responses to extreme rainfall events were also noted in over 80% of

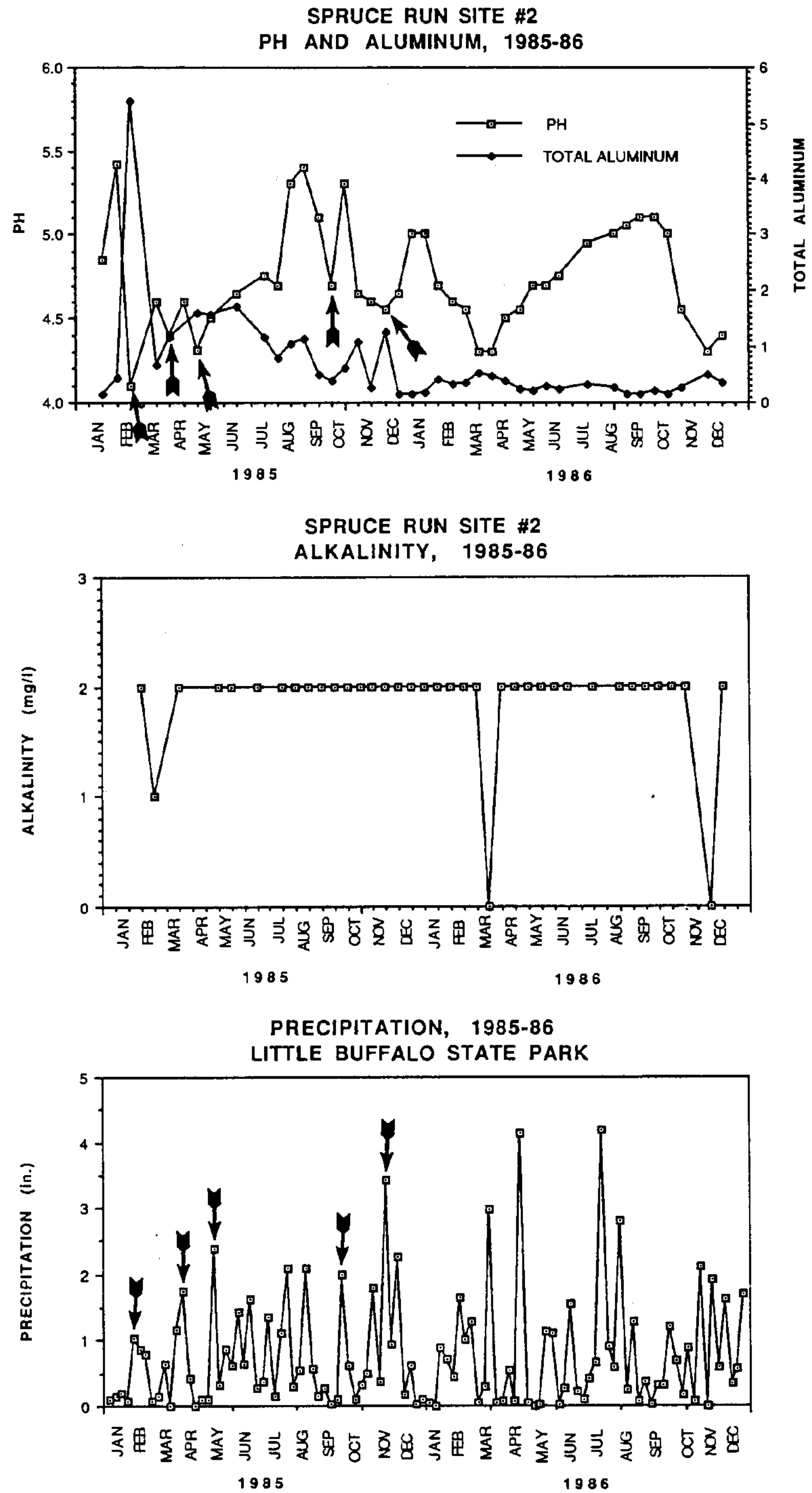


Figure 30. Data on alkalinity, pH, and aluminum concentrations collected biweekly from Spruce Run site #2 in 1985 and 1986. Rainfall data from the closest Pennsylvania State University Atmospheric Deposition site is included for comparison. Ar-Arrows indicate periods of heavy rainfall and depressed pH.

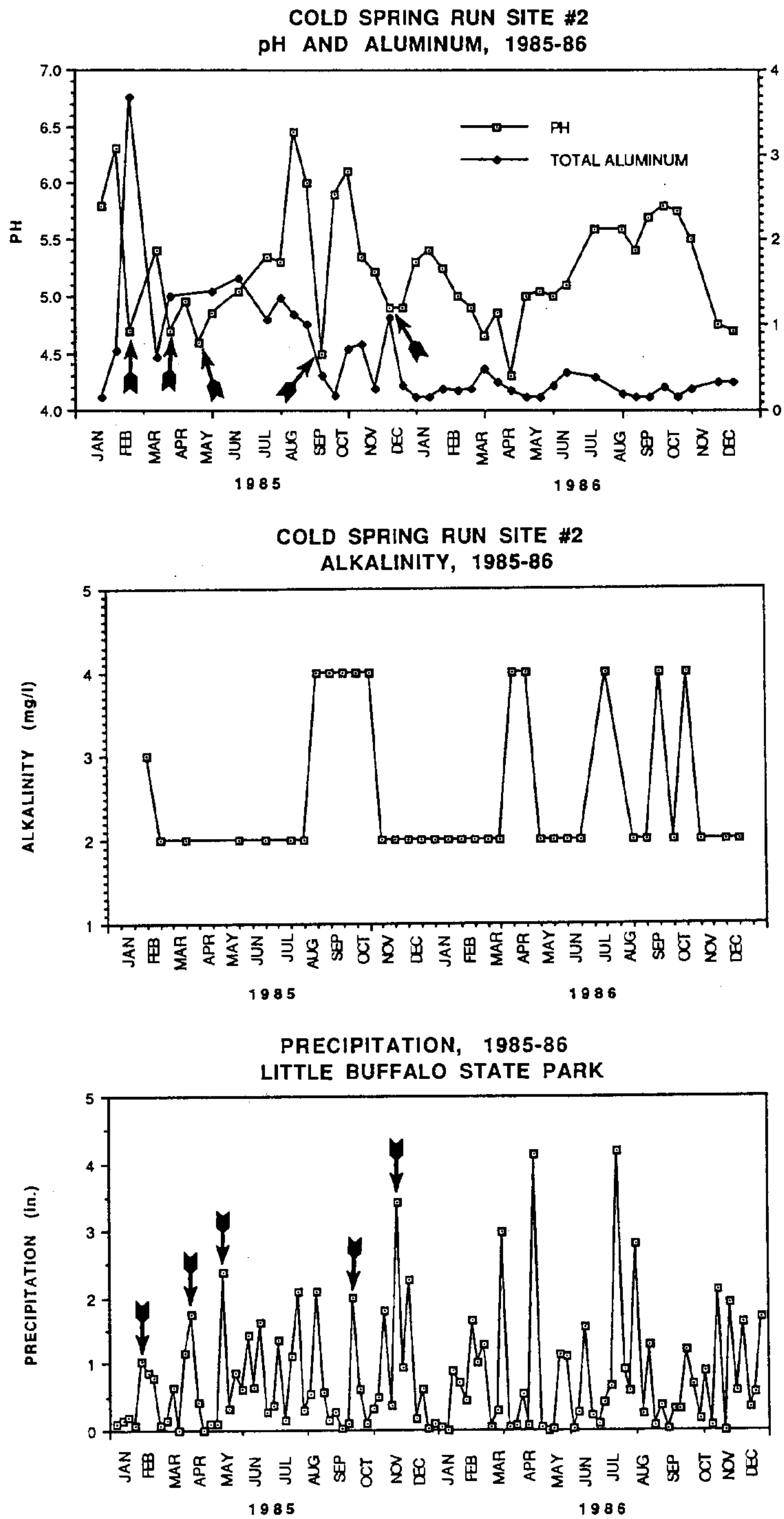


Figure 31. Data on alkalinity, pH, and aluminum concentrations collected biweekly from Cold Spring Run site #2 in 1985 and 1986. Rainfall data from the closest Pennsylvania State University Atmospheric Deposition site is included for comparison. Arrows indicate periods of heavy rainfall and depressed pH.

all streams sampled by ALLARM in 1988 and 1989, and consisted of depressions in both pH and alkalinity concentrations. In some cases, rainfall events decreased alkalinity concentrations to as low as 5% of the mean annual alkalinity concentration (Figure 27).

Due to the poor buffering capacity of the two watersheds, both Spruce Run and Cold Spring Run are extremely susceptible to the effects of acid deposition. The ranges of pH in Spruce Run and Cold Spring Run during the two year period were 4.10-5.42 and 5.24-6.45 respectively, with average pH's of 4.74 and 5.24, respectively. Although thresholds are subject to many variables, biologically critical pH values for coldwater streams are generally thought to range from 4.7 to 6.0 (Camuto, 1991). In fact, EPA's most recent assessment is that changes in acid-base chemistry, at pH levels below 6.0-6.5 result in changes in biological communities, involving both shifts in species composition and reductions in species richness. However, tolerances among species vary, and, within species vary according to other factors such as temperature, dissolved oxygen and habitat characteristics (Camuto, 1991).

The benthic communities in both streams reflect the acidified environs. A total of 16 taxa with a low total biomass was collected in Spruce Run in April, 1985. Spruce Run was virtually devoid of mayflies and had a community dominated by acid tolerant taxa, including the stonefly, *Leuctra* and the caddisflies, *Rhyacophila* and *Diplectrona*. These results coincide with findings on other acidified streams (Simpson and others, 1985; Kimmel and others, 1985; Haines, 1981). In Cold Spring Run, the presence of mayflies (*Stenonema*) and a seemingly greater biomass reflected the less harsh conditions compared to Spruce Run, although the overall community response was indicative of an acidic environment.

No fish were observed in 100 m of electrofishing in Spruce Run. In 100 m of Cold Spring Run, 5 brook trout were observed. Two were approximately 12 cm in length, while the other three were young-of-the-year, approximately 2.5 cm in length. Studies with brook trout have shown that dissolved aluminum concentrations as low as 0.2 mg/l are either toxic or result in sub-lethal effects such as poor growth or lower recruitment, depending on pH and duration of exposure (Siddens and others, 1986; Schofield and others, 1980; Gagen and others, 1987). During 1985, dissolved aluminum in Spruce Run exceeded 0.2 mg/l most of the year and was even greater than 1.0 mg/l after or during storm events. During 1986, the average dissolved aluminum concentration was 0.27 mg/l. Overall, the concentration of dissolved aluminum in Cold Spring Run was lower than in Spruce Run, but it averaged 0.74 mg/l during 1985 and 0.19 mg/l in 1986.

According to residents living near Mount Holly Springs Reservoir, brook trout could be seen in Cold Spring Run above and below the reservoir in the early 1970's. Other observers indicated that brook trout were commonly caught in Spruce Run many years ago, but that the fishery gradually declined and then disappeared altogether in the early 1970's.

In March of 1985, a cement bridge with limestone rip-rap was constructed upstream from site #3 in Cold Spring Run. The pH at that site was elevated almost immediately and continued to be significantly higher than at site #2, upstream from the new bridge (Figure 32). Especially evident were increases in alkalinity at the downstream site when pH at the upstream site was depressed, thus allowing for the dissolution of more calcium carbonate from the bridge. The bridge also slightly reduced total and dissolved aluminum.

Inspired by this unintentional mitigation project, the Cumberland Valley Trout Unlimited group sponsored a liming project on Spruce Run in October, 1990. In an attempt to provide buffering capacity for the stream, 11 tons of limestone rip-rap were placed in the stream at three locations. To date, the project has had no measurable impact on pH or alkalinity.

Questions requiring further research involve the apparent difference between the state of acidification of the two streams, which lie in very similar watersheds. Hypotheses regarding the differences include: recent forestry practices in the Spruce Run watershed, including clearcutting of large areas; the presence of a major fault through Cold Spring Run; and possible differences in the groundwater to surface water ratio in the two creeks.

Since both of these streams feed a public water supply, there is the additional concern that those residents receiving this water are being exposed to excessive levels of lead and copper, leached from household plumbing fixtures due to the corrosivity of the water. A 1984 DER survey of several residences and

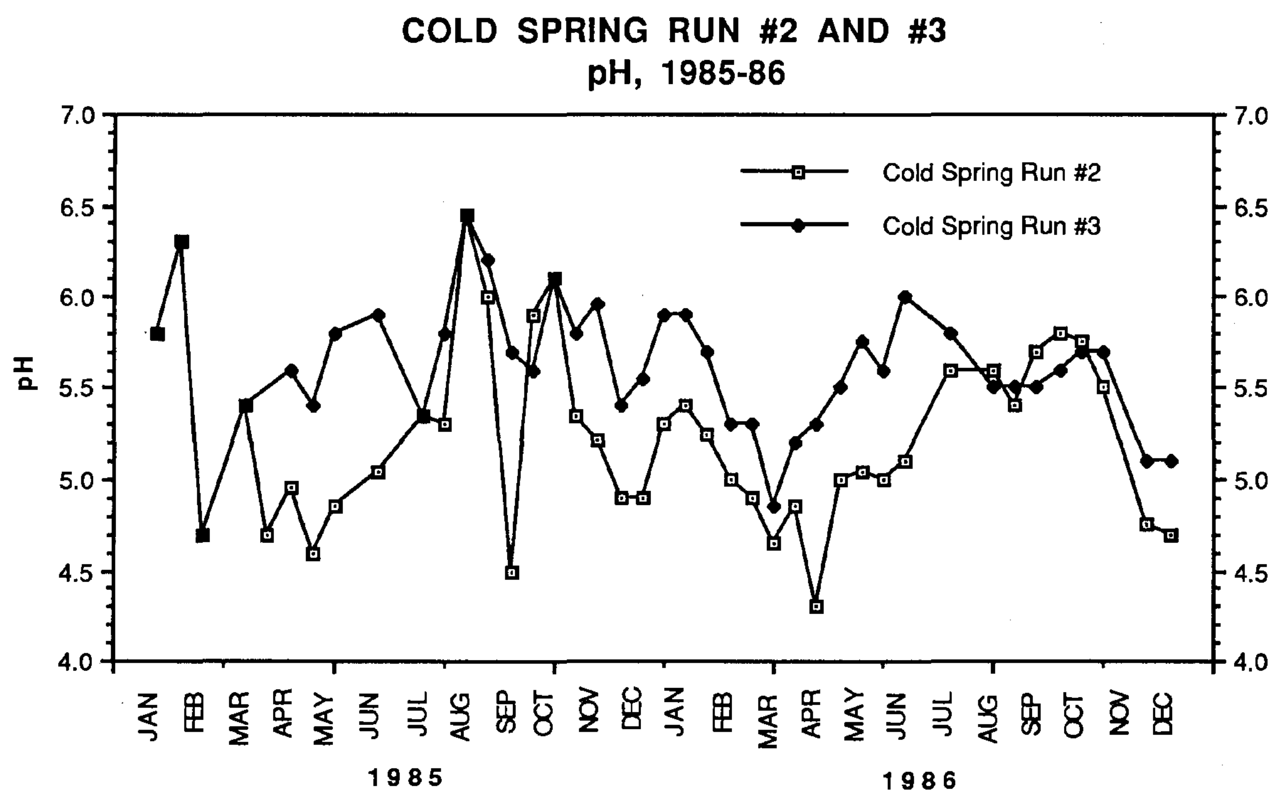


Figure 32. Measurements of pH at Cold Spring Run sites #2 (upstream) and #3 (downstream) demonstrate the buffering effect of the bridge, built between the sites in March, 1985.

businesses served by this reservoir revealed extremely high copper and lead levels, especially after the water had remained in the lines for several hours. However, concerns are not limited to Mount Holly Springs; the towns of Chambersburg and Waynesboro also receive their water from South Mountain reservoirs.

(2) Birch Run Reservoir and DeHart Dam Reservoir

Few investigators have focused efforts on studying the effects of atmospheric deposition on the quality of drinking water supplies over time and the potential implications for human health. Acid deposition can affect drinking water by enhancing the scavenging of atmospheric contaminants, by mobilizing existing contaminants in catchment areas, and by promoting corrosion of drinking water distribution systems (McDonald, 1985).

In Sweden, researchers have documented a national trend of acidification of water supply sources over the past 10 to 40 years, even among the generally less sensitive lakes and streams used as drinking water supplies. They conclude that acid deposition in Sweden is causing this acidification of drinking water supplies (Grimvall and others, 1986). In the United States, Taylor and others (1986) collected data from over 270 water supplies in 12 eastern states. They found that historical declines of pH and alkalinity were consistent with an acid deposition cause and effect hypothesis, and that such acidification had caused violations of EPA's standards for copper, pH, aluminum and lead in many samples.

Recently Wilderman and others (in press) examined patterns of water quality trends in two surface water reservoirs, the DeHart Dam Reservoir, servicing Harrisburg, PA and the Birch Run Reservoir, servicing Chambersburg, PA, and two ground water supplies (Philadelphia Suburban Water Company), all of which have unusually good historical records.

The Birch Run Reservoir and the DeHart Dam Reservoir were chosen to study for the following reasons: 1) data on pH, alkalinity and hardness of the untreated water are available from the early 1940's to the present, both watersheds are located in forested regions, where local land use has remained stable throughout the past 50 years; if water quality has changed in the reservoirs, it cannot be attributed to local land use changes; and 3) both reservoirs are located in watersheds with little buffering capacity, and therefore, should be responsive to a continued influx of acid deposition.

Considering the location and bedrock geology of the Birch Run Reservoir watershed, it is not surprising that historical records document an ongoing process of acidification over the last 50 years. The Reservoir is located east of Chambersburg in the Conococheague drainage basin (Figure 25). The dam was built as a Civilian Conservation Corps project in the 1930's and came into service as a water supply in the late 1930's. The water intake is located about a mile below the dam.

The catchment area above the reservoir is approximately 46 km^2 and is underlain by the Mont Alto Member of the Harpers Formation, the same bedrock that predominates in the Spruce Run and Cold Spring Run watersheds.

Historical records on pH and alkalinity of water coming from the reservoir vary in their frequency. The earliest reported data are from an engineering study (Gannet Flemming Corddry and Carpenter, 1945), and includes biweekly records of alkalinity, pH, and hardness from April 1941-March 1942. From 1958-1966, pH was recorded with an average of 63 readings per year. From 1971-85, pH and alkalinity were measured and recorded daily.

Regression analysis on the historical data shows that there is a statistically significant decrease in mean monthly pH ($p<0.001$), mean monthly alkalinity ($p<0.001$), minimum monthly pH ($p<0.001$), and the alkalinity to hardness ratio ($p<0.05$) from 1941-1985 (Figure 33). The most significant change in pH is the decline from 1941-1979; after 1979 there is no significant decrease in pH. The mean pH drops from 6.35 during the first decade of records (1941-51) to 5.34 during the last decade (1972-85). To better illustrate the change that occurred, Figure 34 shows a comparison of biweekly pH and alkalinity during the oldest year on record (1941-42) and a year chosen randomly from the last decade of record (1975-76). Although pH and alkalinity fluctuate throughout the year, the 1941-42 data are always greater than the 1975-76 data.

The mean alkalinity drops from a value of 8.02 ppm in 1941-42 to 1.16 ppm in 1972-85. However, because there is no alkalinity data from 1942 until 1971, the actual pattern of the decrease is not observable. The alkalinity to hardness ratio, (Wilderman and others, in press) decreases from a mean of 1.13 in 1941-42 to a value of 0.07 in 1985 (Figure 33). However, the lack of extensive data on hardness requires caution in the interpretation of this parameter.

Data from the DeHart Dam Reservoir (also located in an area with poor buffering capacity) show a similar pattern, and are displayed in Figure 35 for comparison. The DeHart Dam catchment area is in the Clarks Creek drainage basin, north of Harrisburg, and is underlain predominantly by the Mississippian Mauch Chunk Formation, consisting of grayish-red shales, siltstones, and sandstones.

Although Birch Run Reservoir has consistently lower pH and alkalinity measurements than DeHart Dam Reservoir, several similarities emerge. Both reservoirs experienced their maximum mean annual pH and alkalinity measurements early in the records (1940's and 1950's); both experiences their minimum mean annual pH and alkalinity in the 1970's. In addition, both show progressively decreasing minimum pH's over the time period studied.

If acidification proceeds in poorly buffered systems by acid input first exhausting the buffering resources of the system, and then impacting the pH, the data suggest that Birch Run Reservoir shows a typical pattern of acidification. The alkalinity was impacted sometime in the 1950's or early 1960's, followed in the 1960's by a drop in pH. The alkalinity at DeHart Dan Reservoir dropped at least 10 years later, and is still not as low as at Birch Run Reservoir; therefore, the mean pH at DeHart Dam Reservoir has not yet dropped significantly. Nonetheless, the minimum pH has dropped, indicating that the DeHart Dam Reservoir system

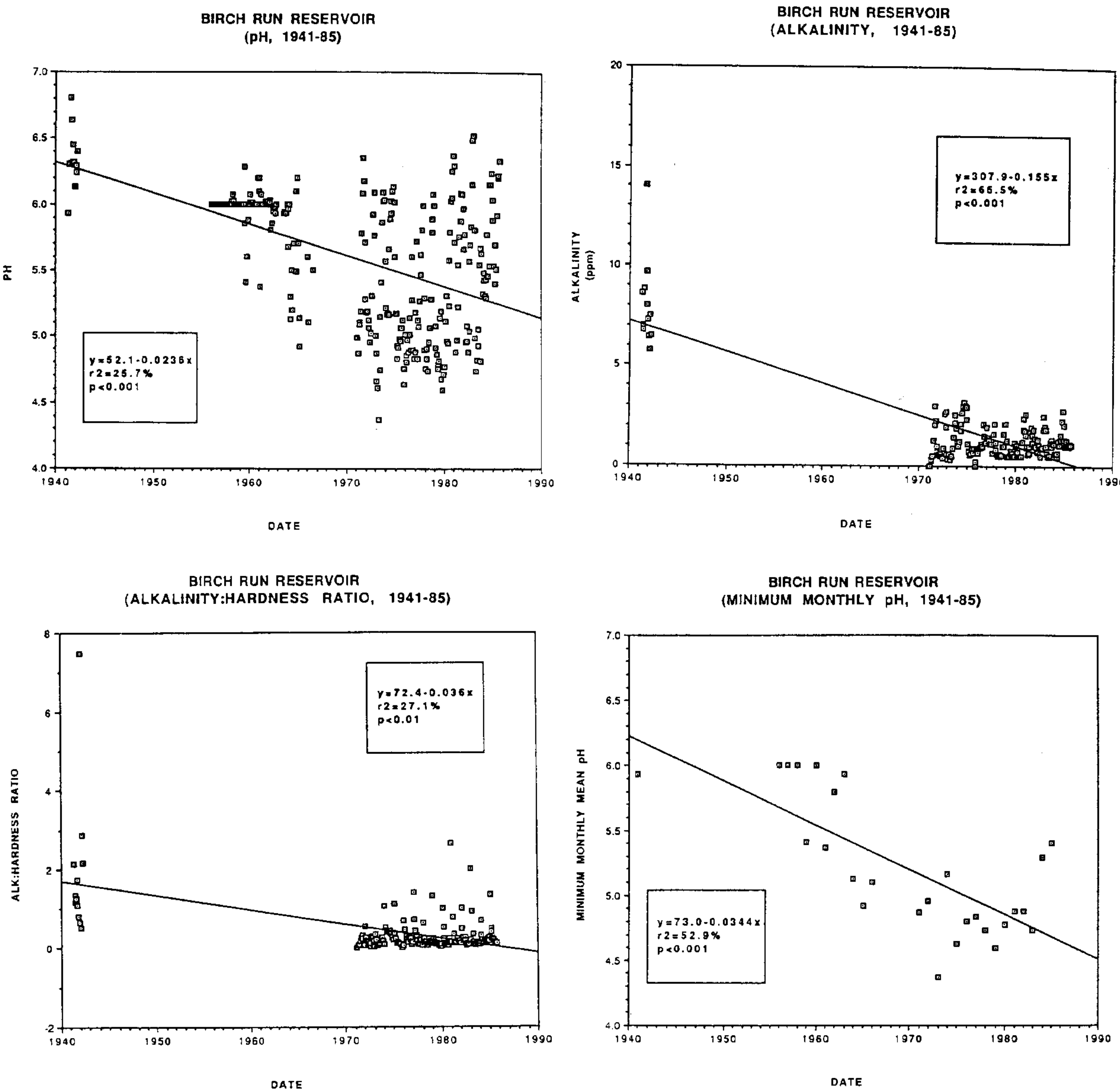
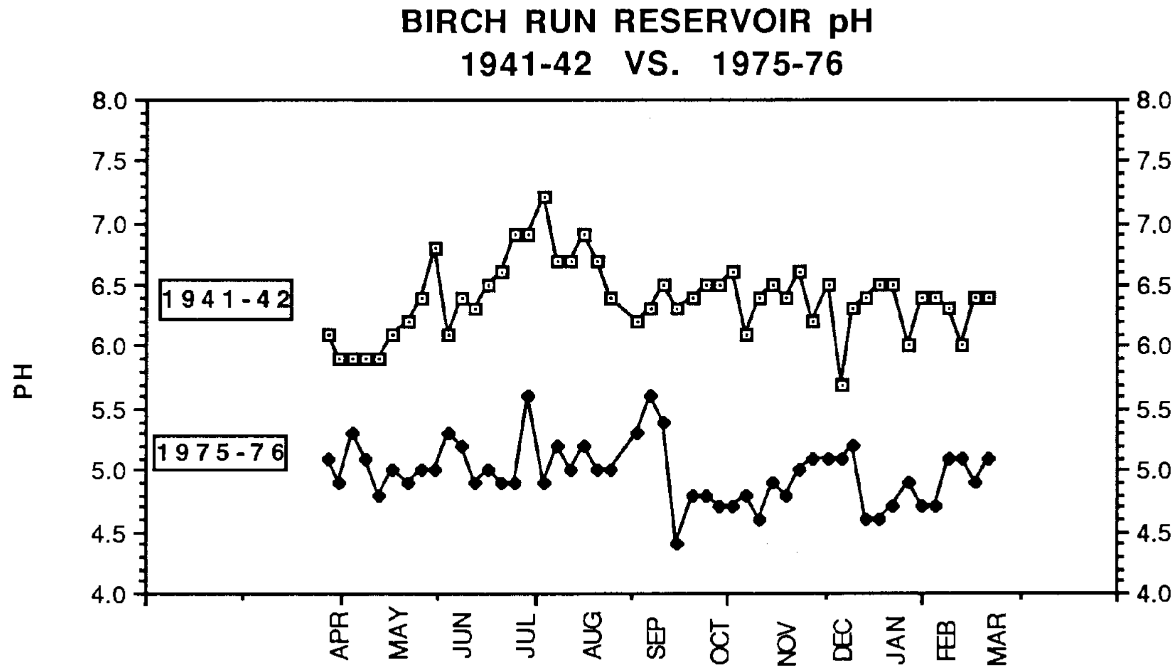


Figure 33. Plots of mean monthly pH, alkalinity, alkalinity to hardness ratio, and annual minimum monthly pH of the outflow from the Birch Run Reservoir, Chambersburg Water Authority from 1941- 1985.

is showing progressively more serious impact during certain periods of the year due to its decreased buffering capacity. We can conclude therefore that Birch Run Reservoir was acidified earlier and is therefore in a more advanced stage of acidification than the DeHart Dam Reservoir.

The differences in the rate of acidification of the watersheds is probably related primarily to the natural buffering capacities of the soils derived from the predominant bedrock in the



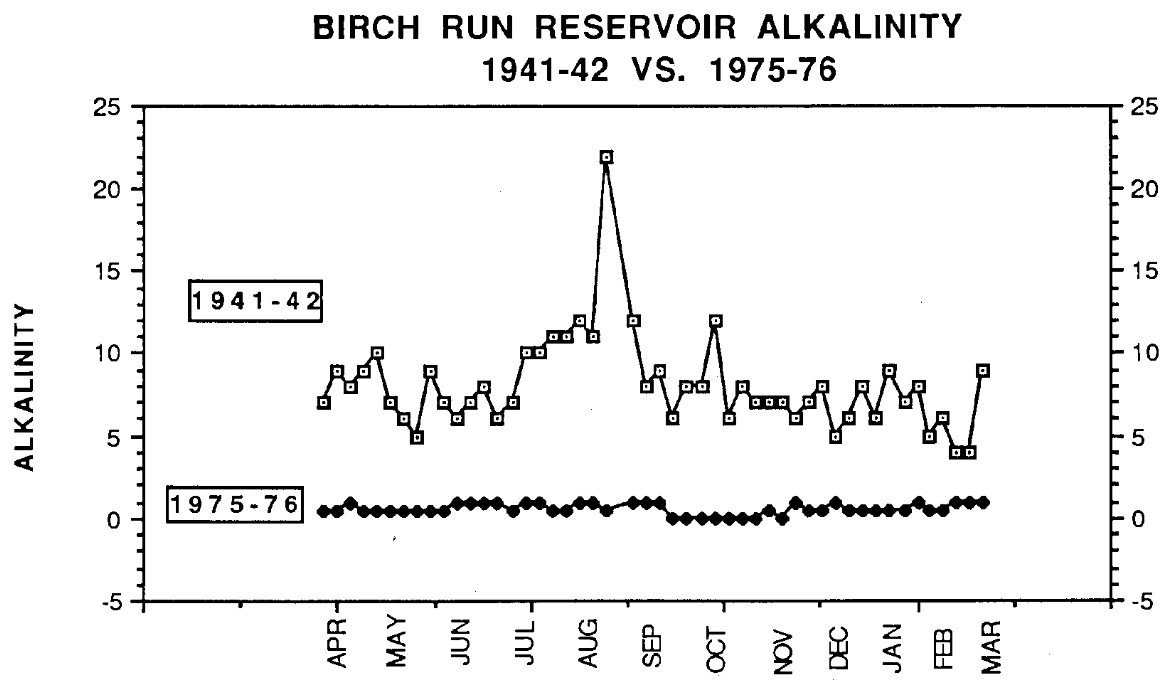


Figure 34. Biweekly pH and alkalinity determinations in 1941-42 and 1975-76 on the outflow from the Birch Run Reservoir.

watersheds. Because the metamorphic rocks of the South Mountain region are known to provide the least acid neutralizing capability, it makes sense that the Birch Run Reservoir would show more acidification than the DeHart Dam Reservoir. Other factors such as size, elevation, and differences in atmospheric deposition may play a role. In both cases, land use practices have not changed significantly in the catchment areas and therefore, atmospheric

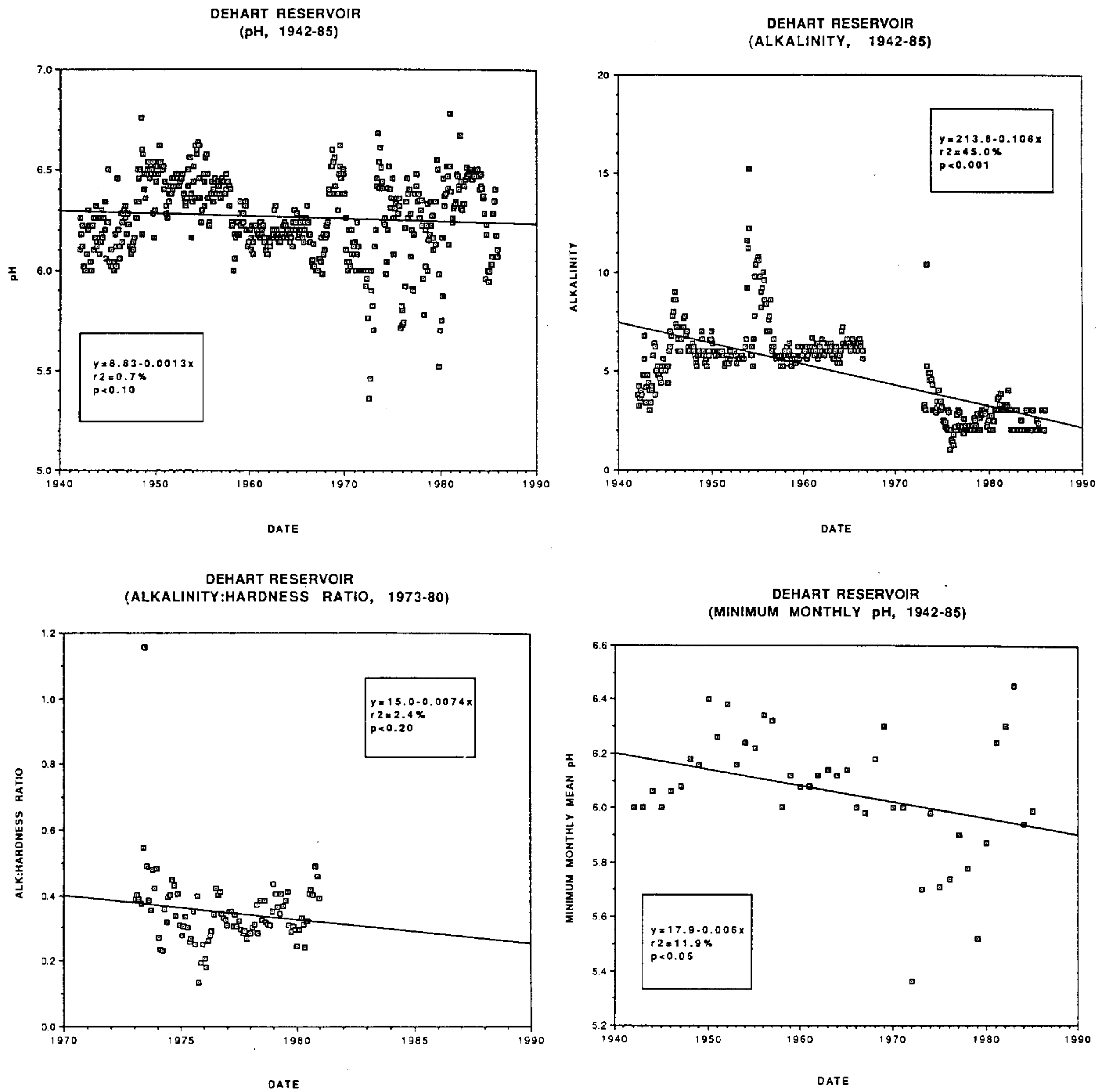


Figure 35. Plots of mean monthly pH, alkalinity, alkalinity to hardness ratio, and annual minimum monthly pH of water from the DeHart Dam Reservoir, Harrisburg Water Authority from 1942-1985.

deposition is the likely source of the acid input.

Interpretation of trend analysis over long periods of time is limited by biases that might have been introduced by changing techniques for pH and alkalinity determination. For this reason, studies documenting changes in diatom communities from bottom cores in both reservoirs are currently underway in an effort to corroborate the historical data.

As noted above, changes documented in these surface water supplies could increase the corrosivity of the water (McDonald, 1985). Although there is insufficient data to establish a cause and effect relationship between acidification of the DeHart Dam Reservoir and the recent elevated lead levels in Harrisburg's drinking water, the possibility of a causal relationship is suggested, and can not be dismissed without further study. The importance of geology in determining stream and reservoir response to acid deposition suggests that all South Mountain water supply systems need to be carefully monitored.

THE MT. CYDONIA SAND OPERATIONS OF VALLEY QUARRIES, INC.

by
Randal L. Van Scyoc
Valley Quarries, Inc., Fayetteville, PA

Valley Quarries, Inc., a subsidiary of the New Enterprise Stone and Lime Co. is located in the south central Pennsylvania counties of Franklin, Adams, and Cumberland and is a full range producer of construction materials. In addition to sand and gravel, Valley Quarries is involved in the extraction and processing of limestones, limestone conglomerate, and an argillaceous hornfels plus the production of ready mix and bituminous concrete, ag-lime, and dimensional stone. Incorporated in 1952, Valley Quarries today employs approximately 180 individuals at a dozen locations.

The sand division operates two processing plants near Fayetteville in Franklin County. Plant #1 is a dry process operation in the Cambrian Antietam Formation. Plant #2 is a wash plant, processing a mixture of the Antietam quartzites and quartzose sandstones and the colluvial sands and gravels from the base of South Mountain near Mainsville, Franklin County. Combined annual production for both plants is approximately 1/2 million tons/year.

Plant #1 is the oldest of all Valley Quarry operations, in operation since about 1915. Two pits are worked north and south along the strike of the ridge-forming Antietam Formation. The remnants of the original workings are still visible at the upper level of the current south pit. Multiple benches with face heights of 35-40 feet are advanced by drilling and shooting conventional vertical blast holes. Six-inch diameter holes are drilled on typically 16-foot centers and loaded with anfo and aluminized anfo blasting agents. Loading and hauling is accomplished with CAT 988 and 980 front end loaders and 30 ton Euclid rear dumps. Sand color segregation and blending is performed to the greatest extent possible at the face. Processing is straightforward with shot sand and stone first passing through a double rotor impactor primary crusher before being sent to the screen house and bins. From here, products are stockpiled via conveyor belt or stock truck.

Products are predominantly 1/4 inch-minus masonry and specialty sands such as asphalt, filtration bed, landfills, and golf course sand. Colors range from brown to orange to pink and buff to pure white. Prices for masonary sands are in the $5 to $7 per ton range. The Antietam is a relatively high purity silica deposit with the white and buff colored sands achieving an apparent S10 value of 99.5 percent or more. Although the possibility has been explored, there are no plans presently to enter the silica market at this location.

Plant #2, in operation since about 1960, is a wet process plant manufacturing a Pennsylvania Department of Transportation (PA D.O.T) type A washed concrete sand plus various other grades of aggregate. The first 20 years of production utilized on site alluvial reserves now near depletion. Substantial additional reserves of this alluvium within the Conococheague Creek Basin exist; however, local opposition has prevented further exploitation of these sites to date. A new permit for a 50 acre tract downstream has been held up in litigation for 5 years. In part, due to this issue, the company in 1981 opened a pit 12 miles away in the South Mountain colluvium. This tract was depleted and reclaimed in 1988. In 1989 a second, larger tract was permitted and opened (Mainsville Site #2). This pit is located about 1 mile away from the original site and lays against the western slope of South Mountain. After clearing, grubbing, and overburden removal, the alluvial/colluvial sand and gravel here is loaded onto triaxle trucks via hydraulic excavator for transport to the processing plant in Fayetteville. Five to six trucks make 10-11 round trips each per day for a total daily demand of 1,000-1,500 tons.

Today, a blend of the Mainsville alluvium/colluvium and shot material from Plant #1 are processed at Plant #2 to achieve the desired feed characteristics and end gradations. Both materials are surged adjacent to the processing plant where they are fed alternately to the 30 x 42 inch-jaw primary crusher via front end loader. Crushed stone and sand next goes to the primary surge where it can be drawn out at a uniform rate for sizing of the 8 x 24 feet triple deck screen. Oversize material goes for further crushing or stockpiling as coarse aggregate. Sand (3/8 inch minus) is washed, classified, and dewatered prior to stockpiling for sale. The plant operates 8 to 12 hours per day at a capacity of about 150 tons/hour. A closed-loop pond network is used to settle out silt and recycle process water for washing operations.

As with most sand plants, the biggest aggravation tends to be excess fines. The plant #1 raw material runs 5-10 percent minus 200 mesh (silt) whereas the Mainsville material is on the order of 10 to 20 percent. The product specifications require less than 3 percent, therefore the excess must be washed out. This creates a "loss" factor of approximately 25 percent. These fines are costly to transport as raw materials, to wash out of the products, and finally to retrieve and dispose of in settling ponds. A comparison of the relative characteristics of the two raw material sources now being blended would seem to indicate the Antietam from Plant #1 to be the obvious best choice, however, from a cost standpoint, the Mainsville alluvium/colluvium (1) does not require drilling and blasting and (2)is less wearing on the processing equipment.

Washed sands are used primarily in the production of ready-mix concrete and precast and prestressed concrete products. Sandstone aggregates are used in architectural concrete products, paving, driveways, landscaping, and for anti-skid materials. Prices F.O.B. are $8/ton for washed concrete sand and $7-9/ton for coarse aggregates. Sands typically have a wider market area than crushed stone. Some sand products are shipped into the Washington, Baltimore, Lancaster, and Harrisburg markets. Still, probably 95 percent is sold within a 40-mile radius of the source.

Valley Quarries is fortunate to have permitted sand and stone reserves sufficient to last well into the next century at a time when, country wide, many deposits are being depleted and operations shut down. The company is confident that through a continued effort toward good public relations and education and a dedication to quality work, Valley Quarries will continue to be a vital and growing entity in the years ahead.

ECONOMIC GEOLOGY OF THE SOUTH MOUNTAIN ANTICLINORIUM

by
Berkheiser, S. W., Jr. and Smith, R. C., II
Pennsylvania Geological Survey, Harrisburg, PA

As can be gleaned from Table 15, history records that the South Mountain region has been blessed with a rich and varied abundance of mineral resources. Somewhat ironically, evidence suggests that stone age aborigines mined silica-rich materials to make tools perhaps as much as 6,000 years ago, while today, nearby, a high-tech quartz crystal growing industry has developed in Carlisle, Pa.

Iron mining prior to the 1900's has probably left the largest trail of activity both in terms of geographic names of State Parks (Caledonia, Mont Alto, and Pine Grove Furnace State Parks, as well as Old Forge State Forest Picnic Area are the sites of former mines, mills, or furnaces), and land forms (ponds and pits) (see Way, this guidebook). As population increases, aggregate mining will play an ever increasing role in the local communities by providing basic construction materials. Fortunately, there is a potentially vast resource of quality construction materials within this region that can be mined in environmentally acceptable fashion (see Van Scyoc, this guidebook).

Potential exists for the Antietam Formation and some quartz veins to fill specialty high-silica niches in the future. The final chapter on South Mountain's mineral resources and man's use of these resources is far from over. It is limited only by our imagination, innovation, and needs.

Perhaps ultimately, the most precious of the "resources" that exist on South Mountain are timber and water. Much of this upland is State Forest and with proper soil retention management can provide future generations with both mineral resources as well as a source of clean water (see Chichester, and Wilderman and Schott, this guidebook) and renewable wood products.

TABLE 15. Summary of some past, present, and future mineral resources occurring in the South Mountain area.

Commodity	Producing Formation(s)	Uses or Grade	Potential Resource Available	Comments	References
ABORIGINAL ARTIFACTS					
Spear points, knives, scrapers, etc.	Catoctin metarhyolite	Dark gray to medium gray aphantic metarhyolite containing feldspar phenocrysts up to 3 mm. Certain clumsy authors can attest to the superior attributes of metarhyolite as a flesh ripping experience.	Locally common pits and "quarries" In recent centuries, demand has waned.	Abundant culls and blanks found at abandoned sites. Dr. Normand D. Keefer, P.C. to B.C. Smith, II, 1973, attested to the manufacture of Catoctin metarhyolite spear points, by Native Americans in Pa.	According to Wallace (1964), jasper [and possibly rhyolite] lost its importance in the arms race when the bow and arrow replaced the spear (i.e. ~3000 years ago).
AGGREGATES					
Pennsy Supply, Mt. Holly Springs (Bender Quarry)	Antietam	PennDOT approved, coarse and fine, with beneficiction specialty sand potential.	Flanks of South Mountain.	Cemetation and vitrification of sand governs abrasion and soundness losses.	Berkheiser (1985ab)
Hempt Bros. Inc. Toland Operation	Antietam Montalto Mbr Harpers	PennDOT approved, coarse and fine.	Major portion of the western 1/2 of South Mountain.	Cementation and vitrification of sand governs abrasion and soundness losses.	Berkheiser (1985ab), Berkheiser and others (1982), Freedman (1967)
Mount Cydonia Sand and Gravel Co., Plant No. 2	Chilhowee Group alluvium and colluvium	PennDOT approved, coarse and fine.	Limited to major drainages proximal to South Mountain.	Resource probably developed during glacial or periglacial activity.	Van Scyoc (this guidebook)
Mount Cydonia Sand and Gravel Co., Plant No. 1	Antietam	Commercial specialty sands and colored sands.	Flanks of South Mountain.	Mining is generally limited to an approximate 100-ft thick friable sand zone. Potential for benficiated specialty products.	Berkheiser (1985ab), Wilshusen and Sevon (1981, 1982), Van Scyoc (this guidebook)
Numerous abandoned and small semi-active sand pits	Antietam	Typically local needs such as borrow and fill and secondary road construction.	Flanks of South Mountain.	Typically developed in middle friable unit which has potential for high-silica specialty uses.	Stose (1932), Stose and Bascom (1929)
Aggregate potential	Catoctin volcanics	Potential for PennDOT approved aggregate. Used in the past for road metal and roofing granules.	Large	Metabasalts could have problems with "flats and longs," and A-T-A. Rhyolites historically have had binding problems. Might have excessive abrasion losses.	
MINERAL FILLERS					
Clay, typically white, but can be multicolored	Carbonate-clastic contact (Tomstown-Antietam)	Residual siliceous clay. Historical uses have included paper, vitreous tile, brick, paint, chinaware, horse liniment, production of white cement, impervious core, fill for dam construction.	Locally >1 million tons (1600 ft x 200 ft x >100 ft) in the region.	"Toland Clay Mines" are the most extensive pits in the region. Hosterman reports alunite in clay. No active clay operations. Potential impermeable barrier source.	Stose (1907, 1932), Hosterman (1968, 1969), Stose and Bascom (1929)

Table 15. continued.

Granulated Greenstone	Catoctin metabasalt	Natural and ceramic-coated coated roofing granules; mineral filler in composition stone and flooring, concrete blocks, portland cement, bricks; dressing for clay-based tennis courts; terrazzo.	Substantial	GAF is an active producer of roofing granules and has a robust fines pile waiting to be utilized.	Stone (1923), Nelson (1968), Berkheiser and Smith (1990)
Mica	Catoctin phyllite	Mineral fillers for plastics, paints, rubber, joint cements, pipe enameling, automotive, undercoating cement, porcelain, fire extinguisher powder, and protective coatings on wires; non-sticking agent, abrasive, insecticide carrier.	Unknown, but appears substantial	Historically other mica-rich rocks have been mined as platey mineral filler in the region. Gross Minerals in Aspers, Pa. mines a mica-rich tuff near Mt. Hope, Pa. Phyllites appear to have future potential in these fields. Could we recognize "silica sinter" in this terrane?	Stose (1932), Anonymous (1985)
Pyrophyllite	Catoctin "phyllite"	Same as above plus refractory and ceramic uses	Probably very limited, but unknown.	Several localities mined for insecticide carriers in the past. Wish we had some Au assays.	Freedman (1967)
Silica	Quartz veins and Chilhowee Group	Mineral filler in plastics, paints, rubber, ceramics, and pottery; non-sticking agent; terrazzo; refractories; metallurgical.	Quartz veins limited (100,000 tons reported), Sandstones large. Friable middle unit of Antietam (~100 ft thick) has best high-silica potential.	Veins up to 65 feet wide exist. Other than aggregates, no known current specialty uses or mines.	Smith and Berkheiser (1985), Berkheiser (1985), Stose (1932)
Copper, native	Typically Catoctin metabasalts, but showings in rhyolite, phyllite, and superadjacent clastics.	Typically grades of 2% Cu across ~8 ft. Deposits occur in narrow discontinuous belt trending N20°E for about 9 miles.	Limited, Rose (1970) estimates value of past production at $10,000.	Rose estimates 15 localities or prospects on record. Several generations of prospecting apparent, from mid 1800's until 1950's(?).	Rose (1970), Stose (1910)
Iron, typically limonite (bombshell ore)	"Valley ores" - "brown ore" found at carbonate-clastic contact (Tomstown-Antietam). Disseminated octahedral magnetite in Chilhowee Gp.	Usually made into pigs. "Valley ores" typically ~40% FeO. One thousand ton ore pods not unusual.	Limited and mined prior to 1900's to support furnaces at Montalto, Caledonia, and Maria (2 mi W Fairfield).	White and varicolored clay mining superseded iron mining in many of the same localities--Old Forge Park was the site in 1832 of a rolling mill and in 1835 a nail works (M'Cauley, 1878). This local industry went the way of buggy whips at the turn of the century.	Stose and Boscom (1929)
Manganese	Carbonate-clastic contact (Tomstown-Antietam), occurs as nodules in clay	Alloy; coloring agent for bricks, ceramics, and pottery.	700,000 t of ore in place SE of Boiling Springs (Foose, 1945). Mostly high in phosphorus.	Smith (1978) reports iron ore deposits become richer in Mn toward SW-Penn Products Corp. recently mined manganiferous clays used in stoneware from the old "Reading Banks" (Berkheiser and others, 1985).	Foose (1945ab)

Table 15. continued.

ORNAMENTAL STONES - GEMSTONES					
Metabasalt and metarhyolite	Catoctin	Polished arts and crafts, jewelry, etc.	Locally limited	Greenstones locally contain amygdules, fractures, and clasts, containing epidote, albite, quartz, red jasper, cuprite, and native copper. Various combinations of mineralogy and textures can be quite pleasing. Aporhyolite (altered and devitrified rhyolite) especially from the Bigham mine is a mottled light gray and pale pink variety that that is splotched with bright green colors in various tints. Banding, amygdules, mottling, breccia, and phenocrysts add variety in both texture and color.	Stose (1932)
Piedmontite	Catoctin metarhyolite	Polished arts and crafts, jewelry, etc.	Limited to rare.	A Mn-rich epidote having a deep red color and a fine radiating fibrous texture.	Stose (1932), Gordon (1922)
Turquoise	Carbonate-clastic contact (Tomstown-Antietam)	Small, blue masses in clay; unsuited for lapidary.	Very rare	Is this another clue from a polymetallic ore-body?	Geyer and others (1976)
PHOSPHATE					
Wavellite	ditto	Massive; botryoidal; white radial clusters. Chiefly used for making matches.	American Phosphorus Co. mined 400 tons in 1905.	Moors Mill--about 3.5 mi W of Mt. Holly Springs is the most famous locality although others exist.	Geyer and others (1976)

PINE GROVE IRON FURNACE AND EARLY AMERICAN IRON MAKING

by
John H. Way, Lock Haven University, Lock Haven, PA

PREFACE

Nestled, almost hidden, in this verdant Mountain Creek valley of South Mountain in southern Cumberland County, are the remains of Pine Grove's iron-making industry. This partially restored, mountain-stone furnace stands as a monument to those nameless individuals, whose efforts, along with countless others in similar settings throughout the colonies, created an industry that was ultimately destined to move a struggling colonial America through the eighteenth and nineteenth centuries and catapult it into the twentieth century as a world power.

The serendipitous association of iron ore and limestone combined with an abundant supply of timber and water here in Pine Grove and its subsequent exploitation is only part of this story. The history that is represented here integrates such diverse human endeavors as politics, economics, science, and technology with the natural environment in a fascinating story worthy of many more pages than can be devoted to the subject at this time. For more information the reader is referred to Bining (1979), Clarke (1968), d'Invilliers (1887), Flower (1975), Kemper (1968?), Kurjack (1958), Pearse (1970), Rogers (1858), Walker (1966), and Way (1986).

SETTING THE STAGE

Looking back through the pages of history, we can see that following on the heels of the stifling and oppressive Middle Ages, the Renaissance rekindled a refreshing atmosphere, one conducive to a spirit of inquiry and experimentation and marked by geographical exploration and discovery. In Europe, it marked a time of transition from manual labor to machine power, when science and technology combined to furnish an increasing number of life's conveniences as well as to create solutions to agricultural and industrial problems.

New scientific principles led to the invention of many profitable devices that facilitated increased production of manufactured goods and resulted in significant financial gain for successful enterprises. Expansion of commerce created new markets, and greater demand for manufactured items further added to capital return. Thus, within the favorable economic climate of the seventeenth and eighteenth centuries, Europe was propelled into an era of rapid industrial expansion.

These countries underwent a shift from the traditional agricultural-based economy to one that was substantially industrial based. As a result, local small-scale problems grew to become serious regional problems. Severe constraints to land use and the decreasing availability of mineral-resources created national concerns. In addition, human resources were negatively impacted by stresses brought about by rapid population expansion and declining environmental conditions, especially air and water quality.

Reduction of surplus population, development of new and expanded economic market places, and need of new sources of raw materials for industries at home certainly were among the pressures influencing governmental decisions encouraging merchant-capitalists to develop the colonies. Recognizing that burgeoning populations burdened governments with increased relief expenses for the poor and contributed to a general reduction in the standard of living for the nation as a whole, the colonies were seen as a place where large numbers of people could migrate in order to relieve this problem. In addition, impending shortages of raw materials for manufactured items became a growing concern. These supplies could be most economically derived from the colonies rather than from other nations where prices and politics were uncertain and the balance of trade was always in jeopardy. It was for these reasons that several European governments, including Great Britain, actively vied for footholds in the New World.

Colonial America became a home for countless thousands of individuals for a variety of reasons. Some sought religious or political freedom, others hoped for personal financial gain, and still others searched for adventure and the challenges a new, virtually uninhabited land would offer. In addition, many came against their will, conscripted and forced to work for others. Overshadowing the myriad of individual reasons were the advantages these lands offered to those European governments that were responsible for founding the colonies. The new world served as an escape valve, relieving some of the pressures created by rapidly increasing industrialization.

THE PLAYERS

Many individuals were either directly or peripherally involved with the iron-making industry in America during its evolution from embryonic beginnings in the early 1700's to the great industrial complex that was in place by the early 1900's. Great Britain and her iron and steel industry played a primary role in the growth and development of America's iron-manufacturing industry.

The British entrepreneurs, ironmasters, manufacturers, and merchants all sought to manipulate colonial iron production through the British government. Each special interest group fought to preserve their economic interests and to survive in an increasingly more competitive situation.

On the other side of the ocean, it was the American iron plantation owners, many of whom were not under direct British control, the local ironmasters, master colliers, teamsters, woodchoppers, blacksmiths, and multitude of common workers who also struggled to survive. By performing their various duties, these pioneers were ultimately responsible for carving out of the American wilderness a immensely successful iron and steel industry that was to have far reaching and lasting impact upon this developing nation.

Great Britain

"The forests were consumed in melting the ore by a kind of foot-blast, before the introduction of better engines, and the workmen frequently shifted their ground as the wood or ore decreased, till they had overrun the whole country." (Pearce, 1970, p. 7).

Throughout the sixteenth and seventeenth centuries, the manufacture of iron in England moved from small bloomeries and simple forges to cold-blast furnaces, more elaborate forges, and slitting mills. Iron production and trade had flourished, but at a significant price to the countryside. Of greatest concern was the growing scarcity of wood for use as a fuel in the furnaces and forges of the ironmaking districts. Timbering necessary to satisfy the voracious iron works was so extensive that cries arose from those in nearby communities that sufficient fuel for domestic purposes would be impossible to obtain if this denudation continued (Swank, 1892, p. 50). As early as 1558, it was forbidden to use timber trees of oak, beech, or ash for charcoal and one of the English Lords of Trade, Evelyn, cajoled "certainly, the goodly rivers and forests of the other world (America) would much better become our iron and saw mills than these exhausted countries" (Pearce, 1970, p. 7).

Great Britain sought to loosen the constraints that threatened to snuff out the flame fueling her iron manufacturing industry. The plan seemed simple enough. America would serve the motherland as the source for pig and bar iron and English soil would be saved from further decimation. The colonies would be using their vast resources of timber, iron cre, and water to produce this raw material. It would be shipped across the Atlantic where it would be manufactured into products for England, but also for sale to the colonies. English businessmen and merchants began the iron-making industry in America with an eye to exploitation and profit. To the colonists, it began as a means of existence. Later, however, it would be cast into a truncheon of strength for the American colonies, produce controversy and division within Great Britain, and ultimately contribute to the separation of the two nations.

Soon after iron manufacturing began in America, conflicts arose among various special-interest groups in England, each seeking to manipulate this colonial industry for their particular benefit. British ironmasters saw the colonies as direct competitors and sought high import tariffs on pig and bar iron. Iron manufacturers looked to America for cheaper raw materials and therefore promoted lower duties on imported supplies of iron. At the same time, however, fearing competition from the colonies, this group sought parliamentary prohibitions to manufacturing of ironware. Additionally, merchants and traders encouraged any legislation that would promote the exchange of colonial iron or ironware for all types of British merchandise.

This controversy continued throughout the period prior to the Revolution and Parliament was forced to consider several bills in an attempt to satisfy petitions by these groups. Most of this legislation failed to gain sufficient support. However, the Iron Act of 1750, did pass both houses and received royal assent (Bining, 1979, p. 139). This law eliminated all duties on colonial pig iron, limited importation of bar iron, prohibited the erection of any new mills and forges in the colonies and restricted production of manufactured items, and required reports from each colonial governor on the activity of ironworks within their jurisdiction (Bining, 1979, p. 139, 140).

In spite of some benefit to the iron-making industry in the colonies, most of the restrictive clauses of this act were not taken seriously and there were many infractions. Violation of this act "brought with it an attitude of defiance to the mother country, and the prohibition itself was a grievance that irritated the colonists" (Bining, 1979, p. 143-144). Clearly the Iron Act of 1750 can be seen as one more factor contributing to the call for revolution.

The American Colonies

"...in two places of the countrey specially, one about foure score and the other sixe score miles from the fort or place where wee dwelt, wee founde neere the water side of the ground to be rockie, which, by the traill of a minerall man, was founde to hold iron richly. It is founde in manie places of the countrey else. I know nothing to the contrarie but that it maie bee allowed for a good marchantable commoditie, considering there the small charge for the labour and feeding of men; the infinite store of wood; the want of wood and deerenesse thereof in England; and the necessity of balasting of shippes." (Thomas Hariot quoted by Bishop and cited in Swank, 1892, p. 102-103).

It was during the first attempts of the English to establish colonies in the New World that iron was discovered. Ore sent from Jamestown to England was smelted and declared to be "as good iron as any in Europe" (Pearce, 1970, p. 8). Despite Indian attacks, locally poor quality ores, the lack of financial backing, business failures and lawsuits, various iron-making efforts persisted and became successful enterprises. From these early manufacturing centers at first located in coastal regions or near navigable rivers, the iron industry gradually moved inland, spreading westward from one colony to the next.

Colonial Iron Plantations

"...a tie of common interest, stronger than exists to-day under similar relations, bound master and workman together. Whether the workman were their own masters or not they were virtually fixtures of the furnace or the forge. The ladies of the 'big house' disdained not their poorer sisters, but were often their teachers, often their nurses and physicians, and always knew them by name and would recognize and greet them with politeness. If daily toil was the common heritage of the workmen and their families it may be said that their wants were few and their aspirations were humble. If there were bare floors in the little log houses there was food and there was warmth within their

walls." (Swank, 1892, p. 189).

In general, iron-making throughout colonial America proceeded along similar lines with respect to the overall organization and set-up of the ironworks, methods of production, types of transportation, and distribution of the finished products. The industry at this time also had much in common with the early stages of iron manufacturing in Great Britain.

Communities grew up around the furnaces where the ironmaster, his workers, and all their dependents lived and labored. Flower (1975, p. 19) described the iron works community as "almost feudal in its dependence upon the Iron Master for food, shelter, and the work which made both possible. But it was, we are told, in the best sense of the word patriarchal. The woes of the men were not unheeded by the Iron Master, for so closely were capital and labor inter-dependent that the good of one was the good of the other."

Compact and largely self-sustaining, iron plantations comprised several thousands of acres of land. Most of this land was forested because of the enormous amount of wood needed to make charcoal, the fuel of the furnaces. Wood was also used for the cooking stoves and to heat the homes and shops within the community. A number of acres of cleared land was used to grow grain and raise food to support the community. The iron mines and limestone quarries were also part of these extensive land holdings.

Scattered across the countryside today are remnants of old furnaces, some still associated with a large stately stone mansion. With the exception of a few carefully preserved and restored iron-making centers such as Hopewell Village, a National Historical Site here in Pennsylvania, these few structures are usually all that remain of the once flourishing iron plantations. Long gone are other structures directly involved with iron making such as the cast house, a bridge house, the "coal" house, and perhaps a forge. Other buildings comprising a typical community included workers homes, the company store, possibly a schoolhouse, and subsidiary buildings including the blacksmith and wheelwright shops, barns and sheds. Gristmills and sawmills, built on the races and run by waterpower, contributed to the self-sufficiency of many communities.

The Iron-making Furnaces

"When the blast furnace was in operation, sparks and smoke billowed out from its stack. At times flames spurted forth....the red glow of the furnace was like a colossal beacon, visible for miles around....in the darkness it was awesome and terrifying." (Clarke, 1968, p 22-23).

BLOOMERIES. In the early stages of colonial iron manufacturing, both bloomeries and cold-blast furnaces were operating; bloomeries flourished in New England, whereas, blast furnaces came into use almost from the beginning in colonies to the south. The iron-making process in bloomeries was a relatively simple one.

Ore was broken into small pieces and heated. As the iron became semi-molten, it was stirred with a long bar until it gathered into a lump, called a bloom. This pasty lump of iron, which had never completely melted, was removed from the fire, hammered into a bar to be used by blacksmiths and artisans. This process was wasteful and inefficient; large amounts of charcoal were required and the iron was full of slag.

COLD-BLAST CHARCOAL FURNACES. Blast furnaces were the result of a technology that had slowly evolved from German and English modifications to the bloomery and simple forge. Usually built into hillsides, furnaces "in blast" were continuously filled from the top with charges, alternate layers of iron ore, charcoal, and limestone (Figure 36). Air forced into the furnace from below heated the mixture to a temperature high enough to cause the ore to melt and the iron to separate and accumulate at the bottom. The volatiles escaped through the stack, and the physical impurities from the ore combined with the molten limestone to produce slag, a waste product.

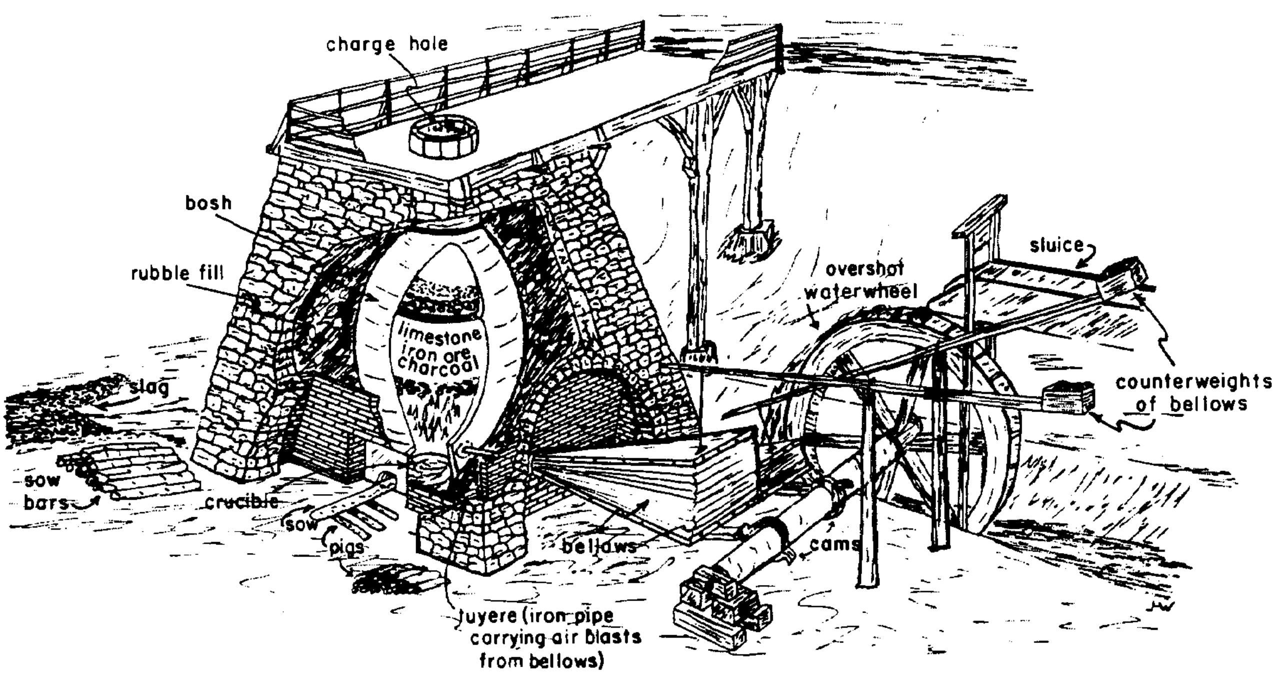


Figure 36. Sketch of a typical cold-blast furnace used to manufacture pig iron in colonial times (from Way, 1986, Figure 6-1, p. 12).

At least twice a day, the molten iron was tapped and allowed to run into molds made in moist sand on the floor of the casting shed. Typically, these molds comprised a long, narrow feeder trough with several smaller troughs or side gutters coming off at right angles, a shape resembling a pig and her little piglets; the terms sow and pig are still in use today. Molten iron was poured into different shaped sand molds to produce hollow ware such as pots, pans, or skillets, or delicately carved wooden molds for specialized castings such as stove plates.

Pig iron was further refined in forges, smaller operations that cast a variety of iron goods for household, agricultural, and industrial uses. It served as the raw material that was hammered into sheet iron in plating mills, or pressed into slit iron for nails in slitting mills, or later used in steel furnaces to produce blister steel for tools. Both pig and bar iron was exported to England, but never in the volumes predicted by the entrepreneurs or desired by the British manufacturers.

HOT-BLAST CHARCOAL FURNACES. The iron-making industry continued to evolve on both sides of the Atlantic. Out of necessity, technology developed in England such that coke (coal from which most of the volatile materials has been removed) and raw-coal, hot-blast furnaces came to be used prior to the Revolution, and by 1796, charcoal furnaces had all but been abandoned (Pearse, 1970, p. 129). Thus, in England the industry progressed due to abundant capital and advanced engineering skills combined with the availability of superior fuels. During this same period in America, bituminous coal was far to the west in the wilderness where transportation was poor to non-existent and anthracite was unknown as a fuel. Both the technology and capital were lacking, and as a result, iron-making lagged throughout the eighteenth and nineteenth centuries.

Pennsylvania

"...Joseph Farmer, a steelmaker of Birmingham [England] who testified in 1736 before a Parliamentary committee [that] he had tried the Potomac iron and found it too tender, though fit for ordinary uses. Pennsylvania iron he had tried in every form but steel, and found it entirely fit for all purposes" (Pearce, 1970, p. 10).

Bloomeries, furnaces, and forges began springing up in Berks, Chester, and Montgomery counties in southeastern Pennsylvania, in the early part of the eighteenth century, and by 1759, most "iron-works lay within a radius of forty miles from Philadelphia" (Pearce, 1970, p. 81) in the Delaware and Schuylkill valleys.

In the latter half of the eighteenth century, ironworks had sprung up along many of the tributaries of the Susquehanna River and within the Juniata Valley. Prior to the Revolution, few settlers ventured west of the Alleghenies, because of Great Britain's policy of restricting settlements to the coastal fringe territory and concerns over Indians and the large tracts of uncleared and unprotected lands (Bining, 1959, p. 52-54). After independence had been secured, increasing numbers of immigrants moved into and through the Monongahela region on their trek westward. Many stayed, finding rich ores and abundant timber, water, and limestone in sufficient quantities and began to produce iron. Before the beginning of the nineteenth century, the long list of iron works built in this section of the frontier illustrates the rapidity with which iron manufacturing took hold (Bining, 1959, Appendix A).

The Recipe for Iron Making

"...great ore beds, the thick woodlands assuring tremendous reserves of charcoal, and the bold streams promising water power..." (Kemper, 1968?, p. 2).

The description quoted above refers to Pottstown in southeastern Pennsylvania in the early 1700's, but today seems more appropriate to South Mountain and the area around Pine Grove. Nonetheless it points to those natural resources necessary in the making of iron: iron ore near the surface, abundant woodland to supply the fuel to smelt it, and fast-flowing water to provide power to the blast bellows at the furnace. In addition, a quantity of limestone was required as flux in the smelting process and it was desirable to have a hillside against which to anchor the charging bridge.

IRON ORE. Other than the occasional meteorite that is discovered on the surface of the earth, iron does not occur in its native state, but is found chemically combined with other elements and compounds. The percentage of iron in earth materials varies considerably, but when iron is concentrated in high enough amounts by any one of several natural processes, it becomes of interest as a raw material for the recovery of iron. Several kinds of iron-rich earth materials were mined as iron ore, including magnetite, hematite, limonite, and iron-carbonate.

In South Mountain, the ore was limonite, often referred to as brown hematite or mountain ore. Concentrated by processes operating as local limestone units slowly weathered, "ore banks" tended to accumulate along the edges of the valley at the foot of South and Piney mountains.

Many mining operations were small and shallow, and one or two operators could supply more than one furnace with ore. Frequently farmers supplemented their incomes by mining iron ore on their property. Under some circumstances, larger trenches or pits were developed, but these rarely exceeded 40 to 50 feet deep (Figure 37).

CHARCOAL. Throughout the eighteenth and into the nineteenth centuries, charcoal was the fuel used in smelting iron ore. "It made an ideal furnace fuel, being almost free of sulfur, and its ash, consisting largely of lime and alkalis, supplied part of the necessary flux" (Kurjack, 1958, p. 13-14).

"Coaling" a 10- to 14-foot-high stack of 30 to 50 cords of wood (Figure 38) was a job for specialists. A master collier and one or more helpers built the charcoal pit, lighted it, and then monitored it continuously, night and day, for several weeks until it was finished charring. As many as 10 colliers were needed to keep one furnace going. An average of 800 bushels of charcoal was consumed in a 24-hour period, and more than 240 acres of woodland might be required over a year.

LIMESTONE. A fluxing agent was necessary in the iron-making process to help separate the iron from the contaminants in the ore. In many regions, limestone adequately served this purpose. In coastal regions where limestone units were unavailable, oyster shells served as a substitute.

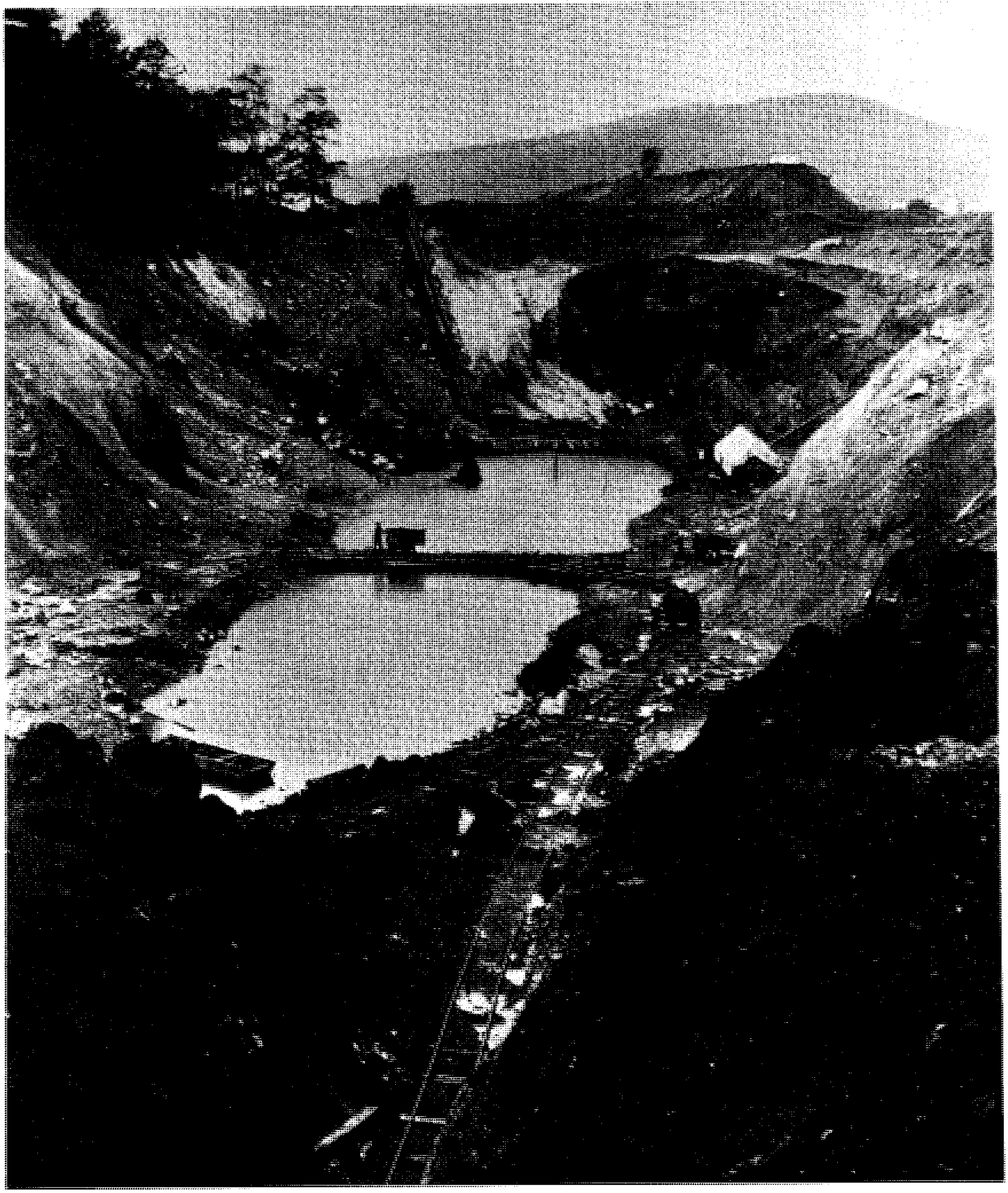


Figure 37. Ore pit at Pine Grove circa 1875. The view is to the northwest. Today this pit is filled with water and is called Fuller Lake. Photograph courtesy of Cumberland County Historical Society (Potter, 1981, Figure 10, p. 22; Way, 1986, Figure 7-5, p. 14).

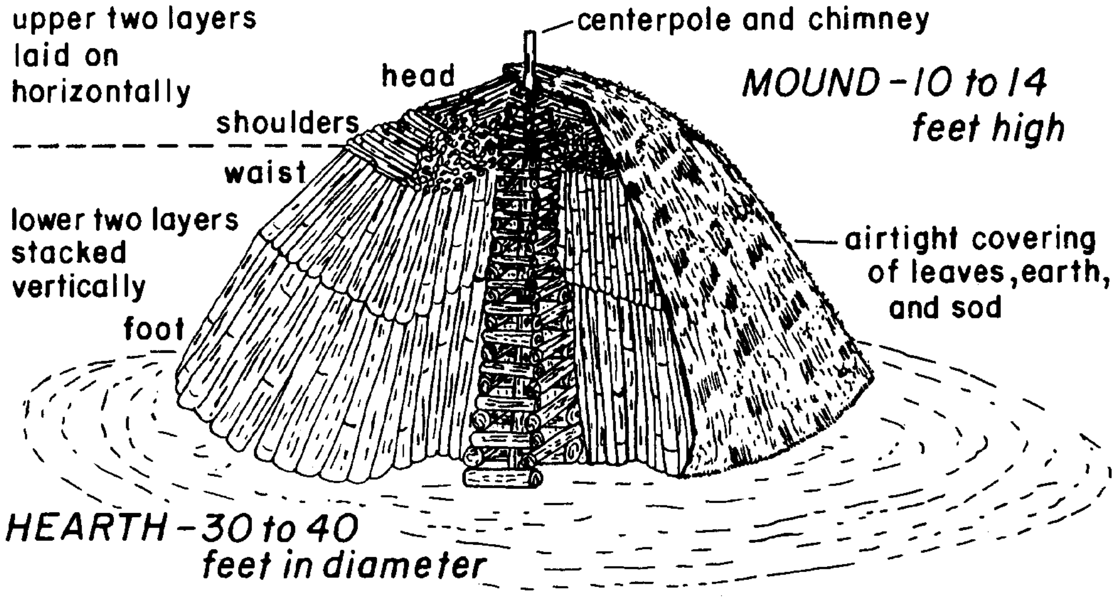


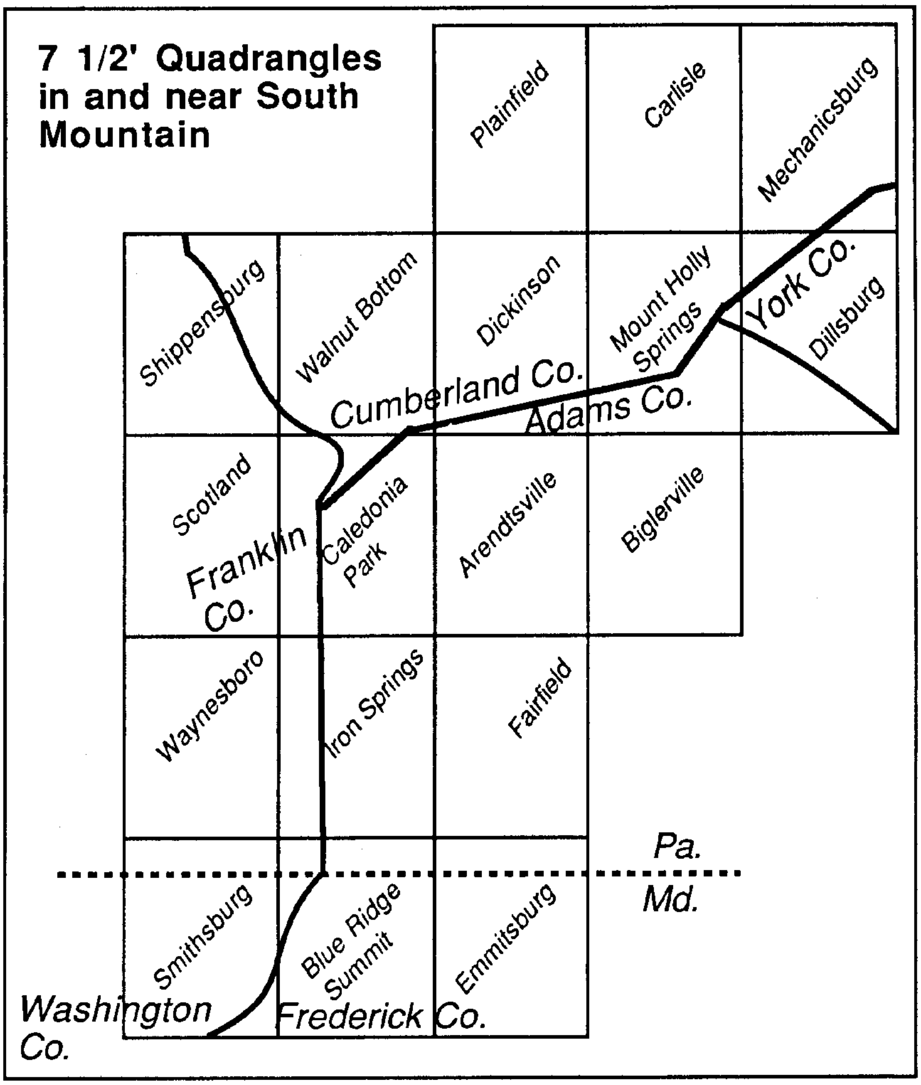
Figure 38. Sketch of a charcoal pit where timbers cut into 4-foot lengths are piled and "coaled." Charcoal produced in this manner was used as fuel for furnaces and forges of the area (Way, 1986, Figure 6-3, p. 12).

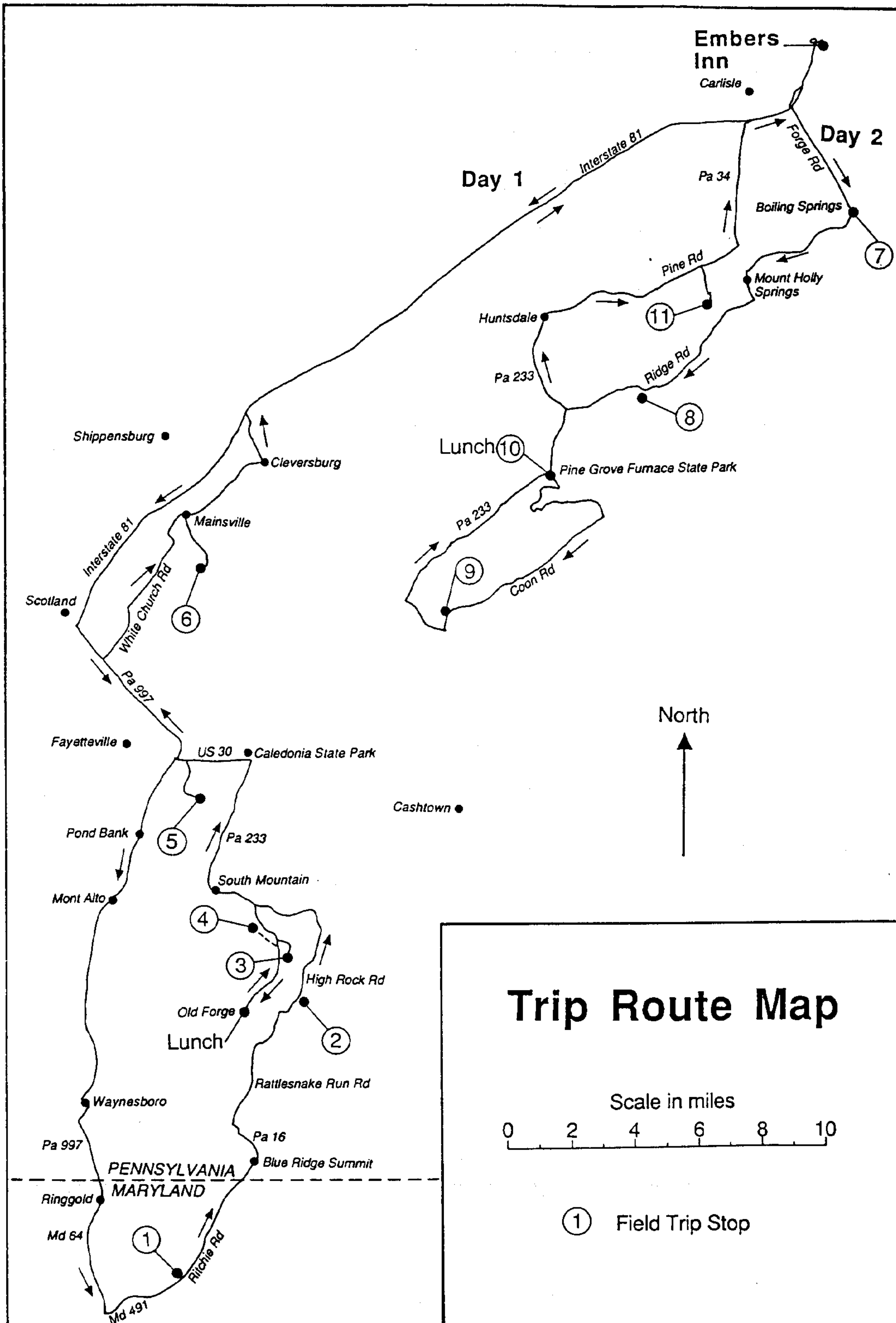
The melted limestone combined with non-metallic impurities in a liquid mixture that floated on top of the molten iron. Periodically this liquid was drawn off and allowed to cool forming a rock-like material called slag.

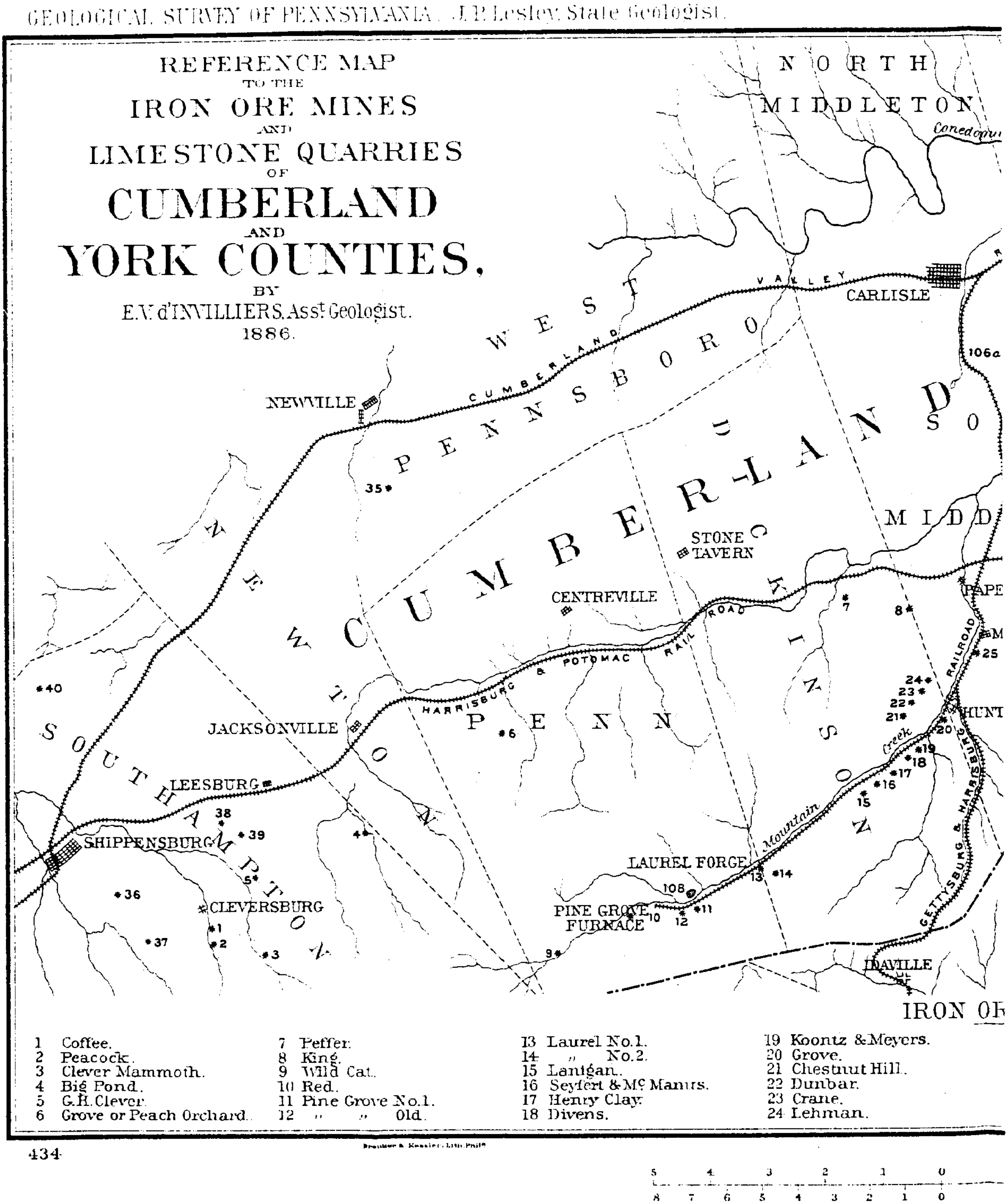
Slag, often found in the vicinity of old iron furnaces, varies in appearance and color. When slag contains numerous gas-generated holes, it resembles the vesicular igneous rock scoria, or when it has a glassy texture, it looks like colorful obsidian. A sky-blue color indicates the presence of manganese; high-grade iron rich in graphite carbon produces a gray color; and dark slag showed the iron was low in graphitic carbon (Bining, 1979, p. 69).

PUBLISHED GEOLOGIC MAPS FOR SOUTH MOUNTAIN AND ADJACENT AREAS

Scale	Area Covered	Author	Notes
1:250,000	Pennsylvania	Berg and others, 1980	State map
1:62,500	Various quadrangles	Berg and Dodge, 1981	Quadrangles not published at 1:24,000
	Fairfield quad.	Stose and Bascom, 1929	Caledonia Park, Arendtsville, Iron Springs, Fairfield 7.5-minute quads.
	Washington County, MD	Edwards, 1978	Entire county
1:50,000	Cumberland Co.	Becher and Root, 1981	Excludes South Mountain rocks
	Adams Co.	Taylor and Royer, 1981	Entire county
	Franklin Co.	Becher and Taylor, 1982	Excludes South Mountain rocks
1:24,000	Carlisle and Mechanicsburg	Root, 1978	
	Mount Holly Springs	Freedman, 1967	
	Scotland	Root, 1971	Western 1/3 of quad.
	Scotland	Fauth, 1968	Eastern 2/3 of quad.
	Caledonia Park	Fauth, 1968	Except SE corner
	Arendtsville	Fauth, 1968	NW corner only
	Waynesboro	Root, 1968	
	Iron Springs	Fauth, 1978	
	Blue Ridge Summit	Fauth, 1978	To MD/PA state line
	Blue Ridge Summit	Fauth, 1977	South of FMD line


7 1/2' Quadrangles in and near South Mountain
Plainfield
Carlisle
Mechanicsburg
Shippensburg
Walnut Bottom
Dickinson
Mount Holly Springs
York Co.
Dillsburg
Cumberland Co.
Adams Co.
Scotland
Franklin Co.
Caledonia Park
Arendtsville
Biglerville
Waynesboro
Iron Springs
Fairfield
Pa.
Md.
Smithsburg
Blue Ridge Summit
Emmitsburg
Washington Co.
Frederick Co.


Embers Inn
Carlisle
Day 2
Forge Rd
Day 1
Interstate 81
Pa 34
Boiling Springs
7
Pine Rd
Mount Holly Springs
Huntsdale
11
Pa 233
Ridge Rd
8
Lunch 10
Pine Grove Furnace State Park
Pa 233
Coon Rd
9
Shippensburg
Cleversburg
Mainsville
Interstate 81
White Church Rd
6
Scotland
Pa 997
North
Fayetteville
US 30
Caledonia State Park
Cashtown
5
Pond Bank
Pa 233
Mont Alto
South Mountain
4
3
High Rock Rd
Old Forge
Lunch
2
Rattlesnake Run Rd
Waynesboro
Pa 16
Pa 997
Blue Ridge Summit
PENNSYLVANIA
MARYLAND
Ringgold
Md 64
1
Ritchie Rd
Md 491
Trip Route Map
Scale in miles
0 2 4 6 8 10
1 Field Trip Stop



From d'Invilliers, 1887.

ROAD LOG AND STOP DESCRIPTIONS - DAY 1

Mileage Inc	Mileage Cum	
0.0	0.0	Leave parking lot of The Embers Inn. **TURN LEFT** onto US Route 11.
0.4	0.4	**TURN RIGHT** onto Interstate 81 South. Route ahead travels across carbonate terrain typical of the Great Valley Section of the Ridge and Valley physiographic province. This area is very suitable for further subdivision following the approach of Godfrey and Cleaves (1991). Such subdivision would separate the Cumberland Valley district in this area into two geomorphic subdistricts: Shale Uplands and Carbonate Lowlands. Here in the Carbonate Lowlands note such features as: no integrated surface drainage, common undrained depressions, relatively low relief, and occasional outcrops on sides and tops of low hills.
5.3	5.7	Pass Exit 14 which leads to PA Route 34 and Carlisle (N) or Mt. Holly Springs (S). Along the route ahead will be several views where South Mountain is seen on the left and Blue Mountain on the right. The expanse of the Great Valley Section in this area is about 12 miles.
2.9	8.6	Good outcrops of Ordovician Rockdale Run limestone occur in the golf course on the right.
9.9	18.5	Topography in this area has more relief than carbonate areas farther north. Presumably this is because the area is underlain by different rock units, some of which are sandier and they and the soil parent material generated by weathering are more resistant to erosion. The area is underlain by the Cambrian Elbrook Formation.
8.1	26.6	There is cave under the hillside on the right in the Cambrian Elbrook Formation. The cave passageways follow joint and bedding orientations very well.
0.3	26.9	Franklin County line.
1.4	28.3	Pass Exit 9 which can be used to go to Mainsville by exiting here. Turn left at ramp end onto PA Route Route 696 S, go 0.7 mile, turn left onto Woods Road, go 0.9 mile, turn left onto White Church Road which puts you at mileage 108.0 in this road log.
0.7	29.0	This low topographic crest is the surface divide between Conodoguinet Creek drainage to the north and and Conococheague Creek drainage ahead to the south. What the subsurface interflow directions and ground water flow lines are like in this carbonate terrane would be interesting to know, and of value in land planning.
4.4	32.7	**EXIT RIGHT** off Interstate 81 at Exit 8, Scotland, PA Route 997.
0.2	32.9	**STOP LIGHT. TURN LEFT** onto PA Route 997 S.
0.1	33.0	**STOP LIGHT. CONTINUE STRAIGHT AHEAD. STAY IN LEFT**

		LANE.
0.1	33.1	**STOP LIGHT. CONTINUE STRAIGHT AHEAD.** Note along the route ahead that many outcrops and field surfaces contain gravels derived from South Mountain.
1.6	34.7	Fields along the road are composed of gravel which overlies carbonate rock which has karst topography.
2.3	37.0	Caledonia Gap occurs in South Mountain ahead. Note the topographic offset in South Mountain. This offset is a reflection of the Carbaugh-Marsh Creek fault, described by Fauth (1968) and Root (1970; this guidebook). Conococheague Creek has excavated the valley in the fault zone. Here in the gap, the the fault is a subvertical tear fault with a right-slip displacement of about 2.5 miles inferred from displacement of Catoctin metavolcanics. According to Root (1970), to the west "near Chambersburg, it swings to a southerly course and becomes a steep thrust. Fauth and Root note that a marked change in trend of South Mountain and of of folds, faults, and cleavage occurs across the Carbaugh-Marsh Creek fault. Structures south of the fault are oriented about N20^{o}E; north of the fault, N40-50^{o}E.
0.7	37.7	On the right is the processing plant of Mt. Cydonia Sand, Division of Valley Quarries and our host at Stops 5 and 6.
0.5	38.2	**STOP SIGN. TURN RIGHT** following PA Route 997 and US Route 30 W.
0.1	38.3	**TURN LEFT** following PA Route 997 S.
2.2	40.5	Road to English Valley which enters from the left is adjacent to the Pond Bank lignite locality. This lignite, discovered and described in the 1800's, has been identified as Mid-Late Cretaceous in age by Tschudy (1965). The stratigraphic section illustra-by Pierce (1965) begs for coring and modern palynological and geomorphological analysis of this site.
0.5	41.0	Duffield Road, the center of Pond Bank.
1.0	42.o	The slight topographic rise is the surface divide between Conococheague Creek to the north and west and Antietam Creek ahead to the south. From here to the Potomac River, Antietam Creek lies between the base of South Mountain and the main stem of Conococheague Creek which lies in the western part of Cumberland Valley in the Shale Uplands subdistrict.
2.2	43.2	PA Route 233 goes to the left. **CONTINUE STRAIGHT AHEAD.**
4.6	47.8	Views ahead on the left of South Mountain. Note the considerable relief on the carbonate terrain in this area coupled with an almost complete lack of mapped sinkholes at contour interval of 20 feet. This area is on the Cambrian Waynesboro Formation. Tomstown is 0.75 miles NNE of here.
1.7	49.5	Waynesboro borough limit.
0.5	50.0	**STOP LIGHT. TURN LEFT** following PA Route 997 S at center of Waynesboro.

0.4	50.4	**STOP LIGHT. TURN RIGHT** following PA Route 997 S.
0.2	50.6	**STOP SIGN. CONTINUE STRAIGHT AHEAD.**
0.3	50.9	**TURN LEFT** following PA Route 997 S.
0.6	51.5	Cross East Branch Antietam Creek that drains Antietam Cove and the Old Forge and Waynesboro Reservoir areas.
1.9	52.8	Maryland state line. Now on MD Route 64 S.
1.0	53.8	**STOP SIGN AND BLINKER** where MD Route 418 crosses. **CONTINUE STRAIGHT AHEAD.**
3.6	57.4	Raven Rock Road on left.
0.3	57.7	**TURN LEFT** onto MD Route 491 N. Outcrops of gravel ahead on right.
1.6	59.3	Outcrops of phyllite assigned to the Harpers Formation on the left.
0.5	59.8	Outcrops of sandstone assigned to the Weverton Formation on the left.
0.2	60.0	**TURN LEFT** onto Fort Ritchie Road.
0.5	60.5	Cross small creek. South end of Raven Rock Hollow block field extends to here.
0.2	60.7	**STOP 1.** Block field is very visible from road. There is a gravel pull off on left, a primitive road to the block field just ahead of the pull off. Road curves right ahead.

STOP 1. RAVEN ROCK HOLLOW BLOCK STREAM

Discussant: G. Michael Clark

CAUTION: The blocks are very slippery if wet, especially if lichen-covered. Climb over the rocks at your own risk.

This block stream is one of a number of both forested and forest-free features inferred to be of periglacial origin in this part of the Blue Ridge, but is one of the few known block streams that lacks a vegetation cover mat over the central part of its axial length of about 1 kilometer. A description of other periglacial features on South Mountain is given in Clark (this guidebook). Because few of these block streams are free of living vegetation and accumulations of organic and inorganic matter, this stop permits us to examine block stream microtopography and the composition, size, shape, orientation, and packing of surface blocks and boulders.

Description of the Block Stream

Slope of the stream immediately upslope from the old logging road across it is 3.5°; farther upvalley the slope is 4°. Width of the treeless area is variable, but 30-35 m in the lower part is representative (Figures 39 and 40A).

The bedrock here is mapped as Catoctin Formation beneath the road and to the southeast, and Weverton Formation on the northwest side of the block stream. The contact runs nearly parallel to and along the axis of the hollow (Edwards, 1978). The surface material in this part of the block stream is derived from the

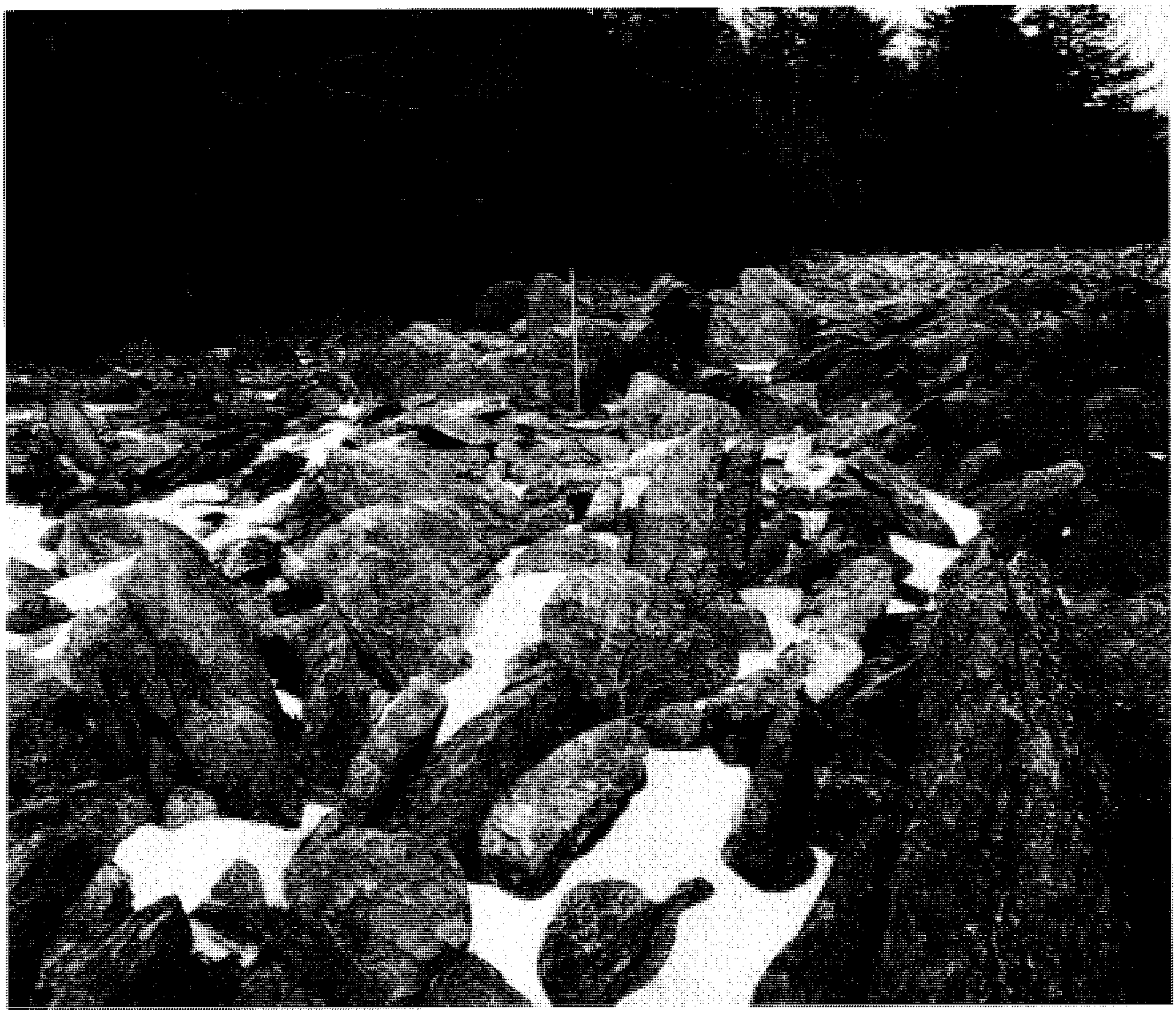


Figure 39. View upslope from lower part of Raven Rock Hollow block stream. Note tendency for larger tabular blocks to be oriented near vertical. Meter stick for scale.

Weverton Formation, and at least some of it appears to have come from "feeder stripes" in the forested area on the northwest side of the block stream (Figure 40B).

On the northwest side of the forested extension of the block stream, outcrop is present almost directly upslope from the Appalachian Trail shelter (along the road that crosses the block stream). The rock is well-indurated, light gray metaquartzite with bedding partings generally between 20 and 50 cm, but some float slabs are over 1 m thick. Etched surfaces normal to bedding on outcrop show delicate crossbedding and closer spaced partings. The rock has much secondary silica in a variety of gash fillings and other fractures, but the main joint system does not seem to have had secondary silica filling. Bedding orientation is N17^{o}E, 58^{o}SE; joint orientation is N75^{o}W, 90^{o}+/-10^{o}. The main spacing of joints in the outcrop is 2-3 m, but there is a strong suggestion that closer spacing may occur in a discontinuous fashion rather than the continuous nature of the wider spacing. The general size of blocks which occur on the lower slope

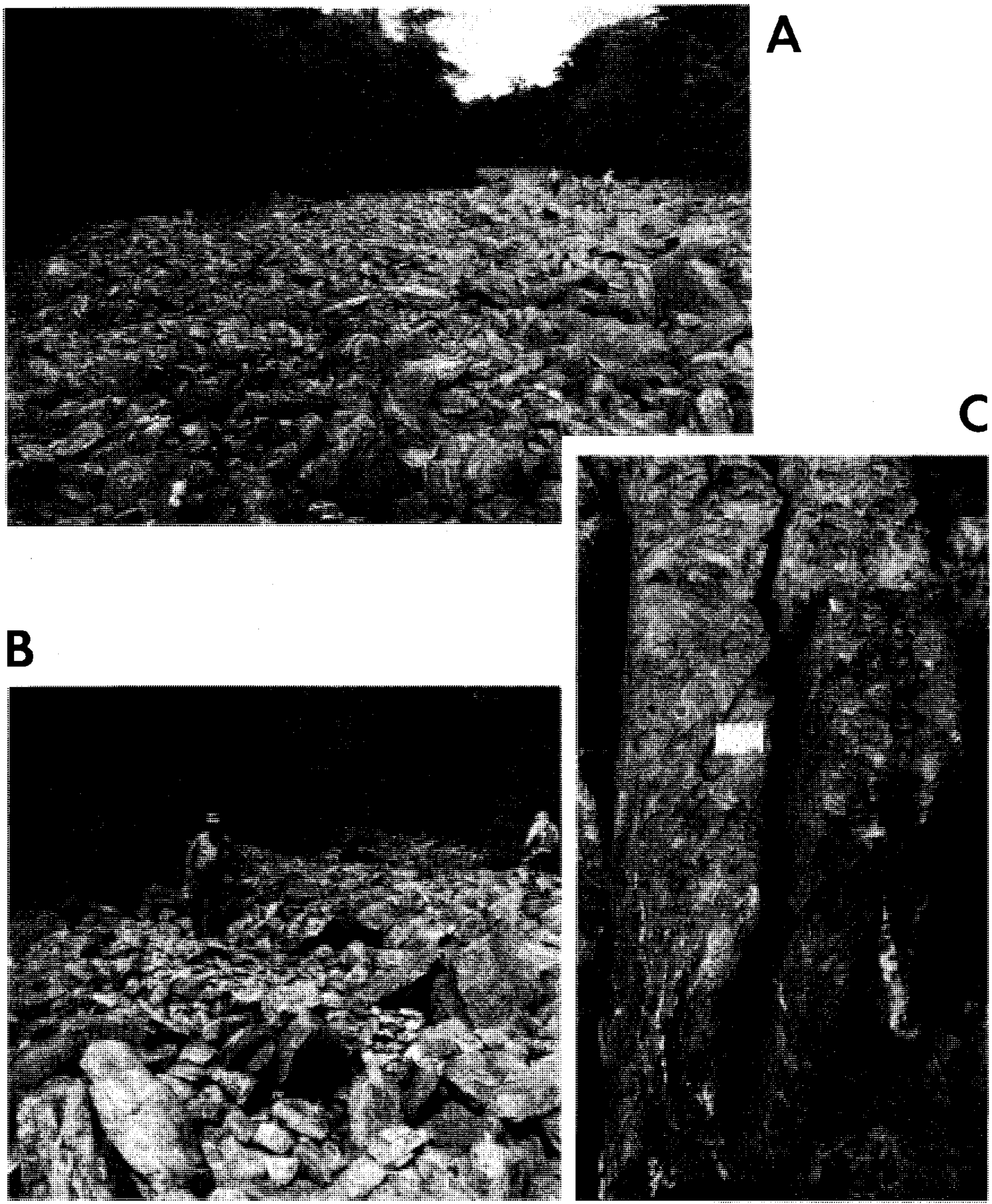


Figure 40. Features on Raven Rock Hollow block stream. A. Central area, looking upslope, showing variation in width of treeless area; B. Area near northwest edge displaying area of smaller, subequant boulders (center figure) with larger tabular blocks to right; C. Block weathering and breakup in place, scale divided into inches and centimeters.

and in the block stream suggest that a closer spacing than 2-3 m is present, but not obvious in outcrop. There is a very subtle hint of a fracture tendency normal to bedding and the primary fracture system, but it is not obvious enough to measure and is probably quite irregular in trace.

Some very large slabs have detached from the outcrop and moved a small distance downslope. The largest is about 7x3x2 m, and is now breaking up in situ. These large slabs are on the slope just below the outcrop. Downslope from these large slabs to an area where slope angle decreases, there are smaller blocks covered partly to completely by vegetation. On the lower gradient slope to the edge of the treeless block stream the blocks are nearly totally covered with vegetation and matrix with only small pieces of blocks showing above the surface. On the treeless area blocks occur which are broken in place but not separated (Figure 40C) and which display solution pits (Opferkessel) where water collects today. Some of these solution pits are 20-30 cm in diameter and about 5 cm deep.

Block fabric was measured along two short transects (downstream from the road that crosses the block stream) by plotting poles to the maximum projection area (ab-planes) of blocks (Figure 41). These measurements suggest that at least late-stage formative processes within the block stream were capable of orienting tabular blocks on edge. There are a number of surfaces with have been ground by movement against another block, but almost none are in close contact with the block against which the carving took place. This may suggest that there has been some settlement in the field after the main movement occurred. Most of the sorted patterned ground is in the form of circles, and this suggests that they developed after block stream movement ceased, at least in these sorted patterned ground area. Also, many of the natural topographic depressions tend to be circular to slightly elongate in the upstream-downstream directions as well.

After periods of high precipitation or rapid snowmelt, loud sounds of subsurface running water can be heard beneath the central part of the block stream. These waters emerge downslope as a surface stream, and could have provided a mechanism for the removal, over time, of any interstitial clastic matrix that may have existed between the blocks. On the other hand, the very presence of running water below the surface would have also provided an abundant source of water to form interstitial ice that might have acted as both a matrix and a mechanism for block transport.

Interpretation

No evidence of downslope movement has been observed to date on the Raven Rock Hollow block stream. If movement is present, it probably is restricted to very slow and essentially vertical settling that would occur if fine material is being removed from beneath the block stream by running water. The presence of vegetation mats at the edge of the treeless area, and islands of arboreal vegetation supported by mats within the treeless area

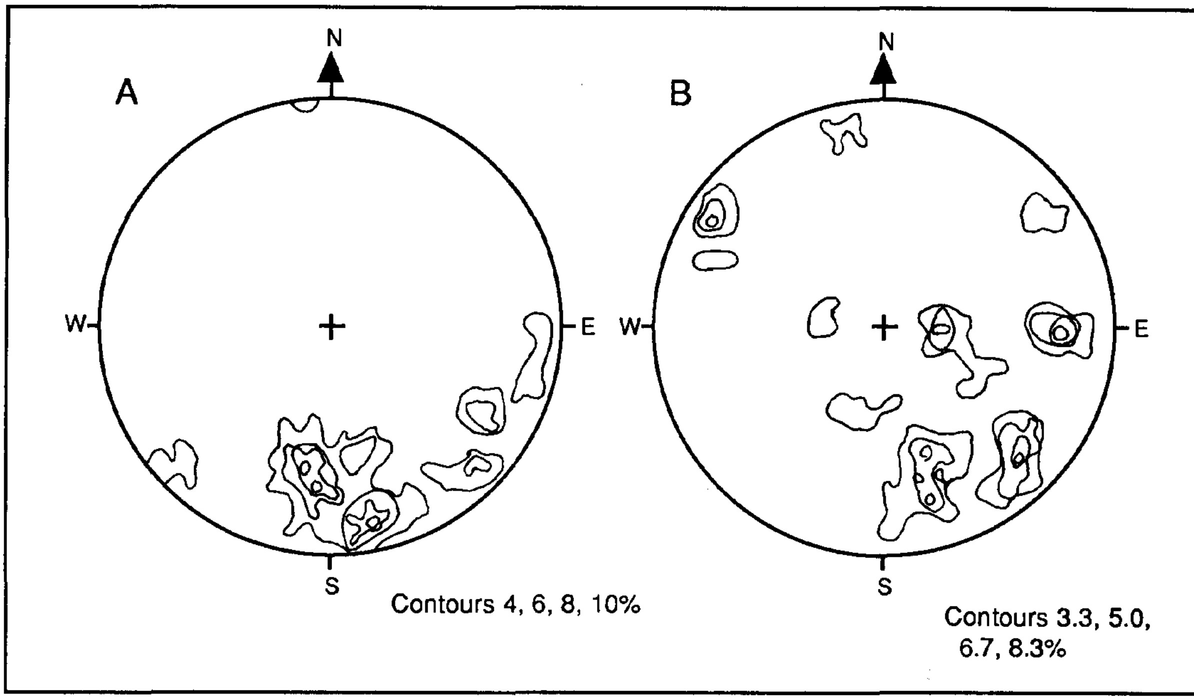


Figure 41. Equal-area nets of poles to maximum projection planes (ab planes) for two sites on Raven Rock Hollow block stream. A. Sample traverse with 50 blocks with long axes >75 cm. B. Sample traverse with 60 blocks >75 cm long.

suggests that the present-day trend is one of forest encroachment. Thus this feature is most logically interpreted as a fossil periglacial landform, although the ages and mechanisms of emplacement are unknown and remain to be determined.

Features to examine:

1. Orientation of tabular blocks, some of which appear to define linear trends on the block stream, or form arcuate "snow plow" orientations.
2. Sorting of large from small boulders to form sorted circles, and rarely, sorted stripes.
3. Microtopography composed mainly of circular pits, but also some elongate pits and a few vague low mounds.
4. Weathering features, including the breakup of large blocks in place, development of solution pits (Opferkessel), and perhaps a few areas where blocks have formed "ball and socket" features due to rotational movement at points of contact.

Questions to ponder

1. Did the blocks in the blockfield move downslope with a matrix of finer material that was subsequently removed, or did it move matrix-free substantially as we see the blocks today?
2. Assuming a periglacial origin, was the blockfield emplaced during just the last of the several cold phases of the Pleistocene, or over several? If the former, why isn't there evidence

of more than one phase, and what happened during earlier cold phases? If the latter, is there evidence of more than one "generation" or pulse of the blockfield?
3. You may wish to ponder the mechanism of formation of the solution pits on a few blocks. This has occurred on quartzite, which we generally think of as among the more insoluble rocks in this climate. How long might it take for one of these solution pits to form?

		LEAVE STOP 1. PROCEED STRAIGHT AHEAD on Fort Ritchie Road.
0.3	61.0	Rock fence on right has blocks of metabasalt which come from upslope but do not show up on the surface of the of the lower part of the treeless part of the block stream. Up to within 0.5 from the drainage divide ahead, the treeless part of the block stream can be seen from the road, especially in leaf off.
1.9	62.9	On left are buildings of Fort Ritchie Military Reservation.
0.9	63.8	**STOP SIGN. TURN LEFT** onto Camp Ritchie Road.
0.2	64.0	**STOP SIGN. PROCEED STRAIGHT AHEAD** on MD Route 550 N.
1.1	65.1	Pennsylvania state line. Enter Franklin County.
1.0	66.1	**STOP SIGN. TURN LEFT** onto PA Route 16 W.
1.2	67.3	**TURN RIGHT** onto Mentzer Gap Road in Micheux State Forest.
0.3	67.6	**TURN RIGHT** onto Rattlesnake Run Road towards Old Forge.
0.4	68.0	Outcrop and float of sandstone from the Weverton Formation on the right.
1.6	69.6	Pipeline crossing.
0.8	70.4	**TURN RIGHT** onto High Rock Road.
1.0	71.4	Scenic view and turn out on left.
0.3	71.7	Enter Adams County.
1.2	72.6	**STOP 2.** Easy site to miss. Stopping place at crest of rise, on curve, near a small, circular, grassed pull out on left.

STOP 2. HIGH ROCK ROAD TOR

Discussant: G. Michael Clark

NOTE: High Rock Road is labelled Three Springs Road on the USGS Iron Springs, PA 7.5-minute quadrangle.

The nearly flat upland here (Figure 42) is typical of a number of "flats" on the crests of ridges in South Mountain (for a discussion of these surfaces, see Clark, this guidebook). The flats are most common on the resistant Weverton Formation and Mont Alto Member of the Harpers Formation. This flat is underlain by Weverton.

Proceed southeast into the forest about 450 feet to a 15-foot-high bedrock knob that projects above the flat you stopped upon. As you walk through the woods, note how flat the surface

Tors are free-standing, tower-like masses of rocks which

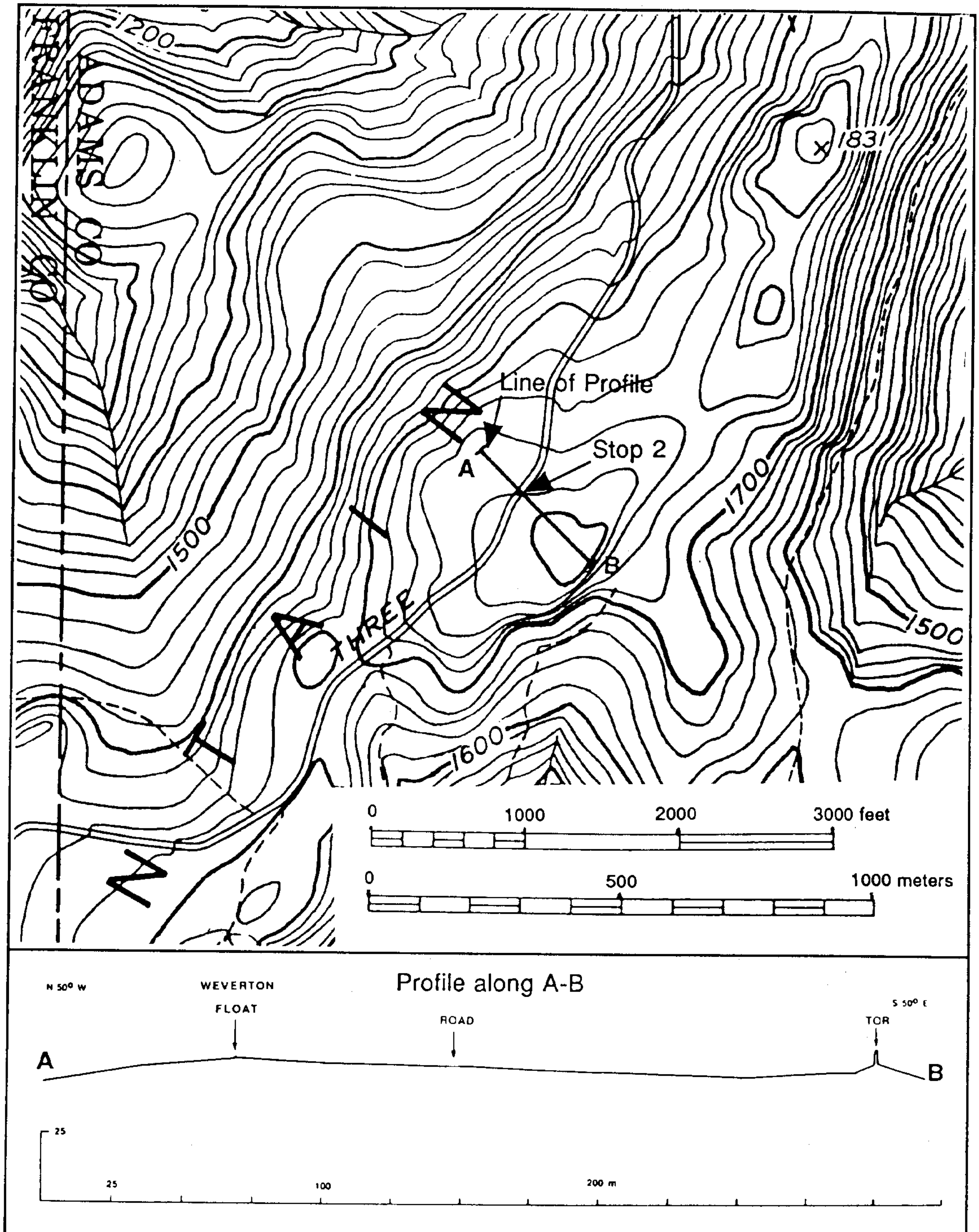


Figure 42. Location map for Stop 2 with topographic profile across the flat. Profile was measured by tape and Jacob's staff. Map from Iron Springs, PA 7.5-minute quadrangle.

is until you reach the outcrop. The flat is very near the crest of a drainage divide with gentle downward slopes both northeast and southwest from the line of profile.
range the gamut from firm bedrock to thoroughly shattered piles of blocks or boulders. Tors develop in thick-bedded to massive resistant rocks. The High Rock Road Tor is composed of Weverton Formation. The original bedrock has been disrupted by massive and pervasive quartz veins, which may account for a greater resistance to subaerial erosion. Exposures around the base of the tor are lacking, however, so that it can not be determined whether or not quartz veins are present on the apron surrounding the tor. Crossbedding can be found in the tor, indicating that the beds are near vertical with original stratigraphic top toward the northwest.

About 300 feet along the low ridge N50°E of the tor is another outcrop of Weverton Formation. This exposure has very few, thin quartz veins so that bedding orientation is preserved. Bedding strike is about N28°E; bedding is overturned with dips of 52° to 70°SE. Crossbeds with tops to the northwest confirm overturning. Surrounding both the tor and the bedrock exposure to the northeast is a mantle of regolith. Around the tor, block cover is best developed on the southeast side of the rock, where block cover on a 20° slope is composed of blocks with a-axes predominantly between 3-6 feet. Their long axes are oriented predominantly in a north-south direction. There are also numerous blocks on either side and to the northwest. The outcrop northeast of the tor is surrounded by a small summit-level block accumulation displaying some interesting microtopographic depressions.

Return along the profile line to the road and cross the road. Go about 300 feet northwest along the profile to another (low) knob. Can you find any bedrock in place? We could not. Just northwest (downslope) from this knob are several interesting microtopographic features developed on relatively stone-free soil. These features include lobe-like forms and a topographic hollow. Are these relatively young features? Or might they be relics of a Pleistocene periglacial climate? If they are older features, how have these forms survived tree root growth, tree throw, and the effects of slope processes since the time of their development?

What processes could have produced the large flat and left the tor and bedrock knob so prominently projecting above it? This site lacks visual evidence of active slope processes, and the ongoing geomorphic processes are suggested to be physical and chemical weathering, occasional block fall, and rare slopewash. The suggested interpretation is that of a periglacial feature that has been little modified by Holocene geomorphic processes. Tors surrounded by flat or gentle slopes are common in periglacial regions such as Alaska today.

Tors have been proposed as ideal sampling sites for specimens of quartzite for cosmogenic exposure age dating methods using accelerator mass spectrometry. This methodology would allow calculation of the length of time that the top of the tor has been exposed to cosmic bombardment, and, if several isotopes of different half-lives are employed, some information can be determined about the rate of erosion at the top of the tor.

		LEAVE STOP 2. PROCEEED STRAIGHT AHEAD.
1.4	74.0	Views ahead on both sides of block debris. Mechanized logging operations have disrupted the surface blocks derived from the Weverton Formation.
1.0	75.0	**TURN LEFT** onto paved Cold Spring Road. Route is now in the headwater drainage area of Antietam Creek. To the north is Carbaugh Run which is tributary to Conococheague Creek. The three cold springs are downslope on the left (south) side of the road. Why are they so cold?
0.4	75.4	Pipeline crossing.
0.2	75.6	**BEAR LEFT** following paved road. Gravel road goes straight ahead.
1.1	76.7	**STOP SIGN. PROCEED STRAIGHT AHEAD.**
0.4	77.1	Franklin County line.
0.6	77.7	**TURN LEFT** onto Old Forge Road towards the Waynesboro Reservoir. Brick building on right corner, brown sign on left corner.
1.2	78.9	Pipeline crossing.
0.2	79.1	**TURN LEFT** onto gravel road leading to PA Fish Commission Waynesboro Reservoir.
0.4	79.5	Road enters from left, bear right.
0.6	80.1	**STOP 3.** Park in parking lot. Site is back up road about 300 feet, then uphill on left.

STOP 3. WAYNESBORO RESERVOIR. CATOCTIN METABASALT AND METARHYOLITE

Discussants: R. C. Smith, II and S. W. Berkheiser, Jr.

Latest Precambrian Catoctin metabasalt and metarhyolite form a crystalline core to the Cambrian clastic sediments that most clearly define the South Mountain anticlinorium. "The Catoctin Metabasalt Story" (this guidebook) describes some of the geochemical and geologic implications of the Catoctin event, but the Waynesboro Reservoir stop provides an opportunity to see how much information can be derived from macroscopic features at the millimeter to tens-of-meter scale. In particular, you are encouraged to contemplate the fluidity, cooling history, and present facing direction of the metabasalt. Venturing to the base of the metarhyolite hill to the west is also not discouraged.

Caution is advised as to your footing on both the slippery turtleback metabasalt outcrop, especially when wet, and the sharp metarhyolite to the west.

Please do not hammer on the metabasalt outcrop or any float blocks greater than 1 m across. Accessible, teaching-quality metabasalt outcrops like this are an endangered species. Metarhyolite outcrops are more common and sampling of the metarhyolite outcrops or float is not discouraged. (If you must have an outcrop piece of metabasalt, please contact one of the discussants to see if they can make arrangements.)

The stop is located in the Iron Springs quadrangle, west of the Waynesboro Reservoir which dams Antietam Creek, Hamilton Township, Adams County, Pennsylvania. The area can be located more precisely as being 1.2 km east of Buzzard Peak or as being at latitude 39°49'11" N, longitude 77°27'28" W. The metabasalt outcrop is at an elevation of about 1350±25 feet and the metarhyolite outcrops at about 1400 feet.

The geology of the area of Stops 3 and 4 is shown in Figure 43, based on Fauth (1978) and recent reconnaissance. It shows the to-be-visited BUZZE metabasalt body to be slightly overturned and to extend from at least the "Loop Road" on the SW nearly to Cold Springs on the NE, and to have a rather linear strike. The eastern contact of the BUZZE metabasalt is covered by colluvium and alluvium and the western contact, also covered, is likely transitional with slightly overturned metarhyolite. Fauth's data also indicate overturning in many nearby areas of clastics, which pose some unresolved structural problems (See 'The Catoctin Metabasalt Story' which sheds additional darkness on the subject.)

Today, however, you are mainly encouraged to observe the BUZZE metabasalt outcrop, which in 15 m of section appears to expose a minimum of 108 basalt flows (Figure 44). This yields an average thickness of merely 14 cm per flow! Even if one discounts many of the flows as being single pahoehoe toes, one must still account for an extremely fluid magma. Figure 44, a stratgraphic section for the exposed metabasalt section could be compared to your interpretations of the outcrop to see if you observe a lesser or greater number of flows.

In addition to flow counting, note the abundance, size, and shape of albite and, to a lesser extent, epidote-filled amygdules. An ideal flow cycle comprises three parts: a few-mm-thick lower chill zone derived from the flow spreading on insulating, and perhaps still hot underlying basalt; a central, several centimeters-thick vesicular basalt zone; and an approximately 2-cm-thick chilled top that developed rapidly in contact with air.

Using the chilled or vesicular zones as indicators, attempt to estimate S_o (primary bedding). Using the incipient but most predominant cleavage, best observed in outcrop in the thicker chilled zones, attempt to estimate cleavage.

Sample BUZZE of the upper 7 cm (including 1 cm of the chilled zone) of a 14 cm flow was analyzed by J. Niemitz of Dickinson College for the Field Conference (personal communication, 4/25/91) as follows: 2.18% TiO_2, 113 ppm Ba, 2 ppm Be, 40 ppm Sc, 108 ppm Sr, 33 ppm Y, 330 ppm V, 160 ppm Zr; 12 ppm La, 40 ppm Ce, and 3 ppm Yb. (For immobile elements, these data are virtually identical to sample SNOMTN from Stop 4).

If interested in the metarhyolite, proceed about 50 m toward the hill and observe the flow banding, presumably due to plastic flow while still partly molten; the subtle, widely spaced fracture cleavage; and 1 to 2 mm subhedral feldspar phenocrysts. A small sample of metarhyolite from the tree-ingrown colluvium block immediately W of the BUZZE outcrop was analyzed by J. Niemitz of Dickinson College for the Field Conference. He reports (personal communication, 4/25/91): 0.46% TiO_2, 505 ppm Ba, 66 ppm Be, 13 ppm Sc, 36 ppm Sr, 128 ppm Y, 4 ppm V, 648 ppm Zr, 117 ppm La, 341 ppm Ce, and 10 ppm Yb. When compared with the analyses for BUZZE, these data suggest the metarhyolite could be a later differentiate from the same magma chamber that supplied BUZZE. When compared with the few published Catoctin metarhyolite analyses, the relatively high TiO_2 suggests an affinity to the metabasalt. However, these interpretations are by no means certain.

Some questions to ask yourself include:

(1) Why was the magma so fluid?

(2) Does the metabasalt appear to be more felsic toward the metarhyolite hill?

(3) What might account for albite vs. epidote--filling in amygdules in adjacent flows?

(4) In the metabasalt, is bedding or cleavage steeper? What does this suggest about facing direction? In the metarhyolite?

(5) What does the relative continuity of features within an outcrop tell about subaerial vs. subaqueous deposition? Do you see any evidence of substantial brecciation?

(6) What length of time might be represented by the observed metabasalt section?

(7) If the BUZZE and SNOMTN samples are chemically the same, what is the possible nature of the structure between them?

(8) How do you interpret the larger (>10 cm) epidotized bodies?

Waynesboro Reservoir Answer Key

(1) Close proximity to a vent may have resulted in a high temperature and retention of volatiles. Evidence for pyroclastic or cascade devolitilization is lacking at this stop.

(2) The matrix color appears to be a little bit lighter on the west toward the metarhyolite hill, but analyses of the collected samples are lacking.

(3) The Na/Ca ratio of the primary magma or of deuteric solutions in the individual flows are possible causes.

(4) Bedding is steeper than cleavage in both the metabasalt and metarhyolite. The whole section is uniformly overturned based on consistent data along strike. At another outcrop in the same metabasalt body, the thicker chill zones are on the tops of flows and a possible ropey pahoehoe surface tends to confirm S_0. At Stop 3, median S_0 and S_1, in the metabasalt are N23°E, 84°SE and N36°E, 46°SE, respectively. In the metarhyolite they are N21°E, 81°SE and N36°E, 24°SE.

(5) In addition to the well-known use of pillows, subaqueous deposition could be indicated by extensive fragmentation, but neither feature has been observed in this area. Pipe vesicles and well preserved ropey pahoehoe, observed south of Mt. Hope, confirm overturning of subaerial flows.

(6) Many lower chilled contacts developed but the magma chamber doesn't appear to have had time to degas. Differentiation within the metabasalt section appears to have been minimal, suggesting that less than 10 years is a possibility.

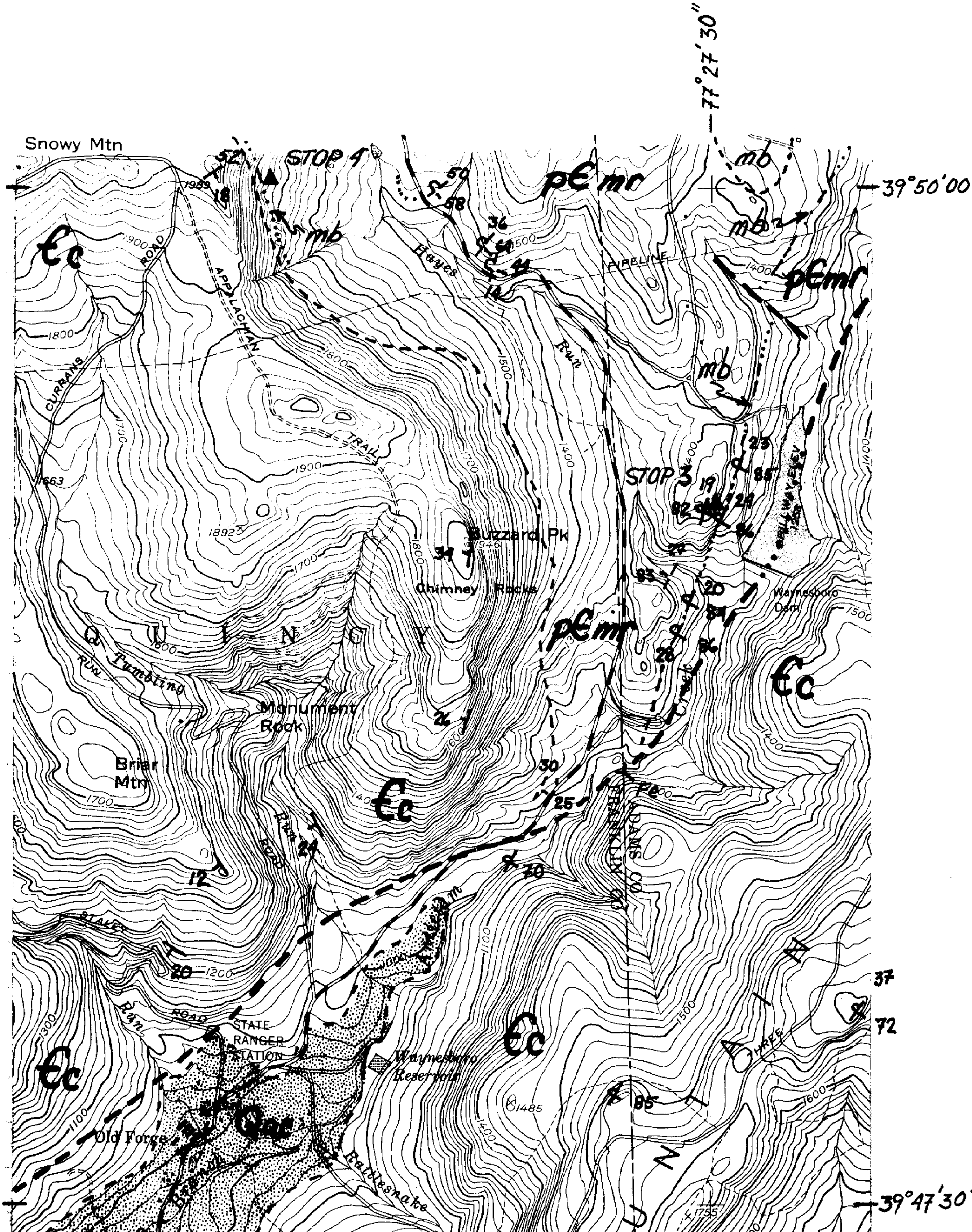
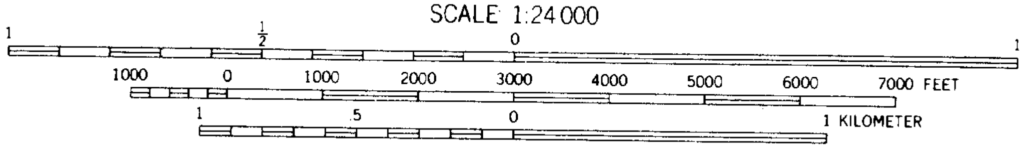
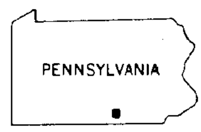
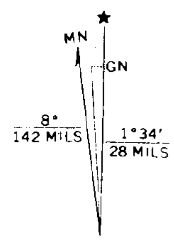


Figure 43. Geologic map of the area of Stops 3 and 4 based on Fauth (1978) and recent reconnaissance.





QUADRANGLE LOCATION



UTM GRID AND 1973 MAGNETIC NORTH DECLINATION AT CENTER OF SHEET

IRON SPRINGS, PA.
SW/4 FAIRFIELD 15′ QUADRANGLE
N3945—W7722.5/7.5
1953
PHOTOREVISED 1968 AND 1973

Qac Cretaceous (?) to Recent alluvium and colluvium debris

Єc Cambrian Chilhowee Group composed of quartzites, sandstones, graywackes, and phyllites

pЄmr Precambrian metarhyolites

mb Precambrian metabasalts

Approximate fault contact; dotted where inferred

Approximate geologic contact; dotted where inferred

Strike and dip of bedding, layering, or flow banding determined by Fauth (1978)

26 inclined

80 overturned

Strike and dip of bedding, layering, or flow banding determined by the authors (median value)

40 35 inclined

40 35 inclined uncertain

40 80 overturned

40 80 overturned uncertain

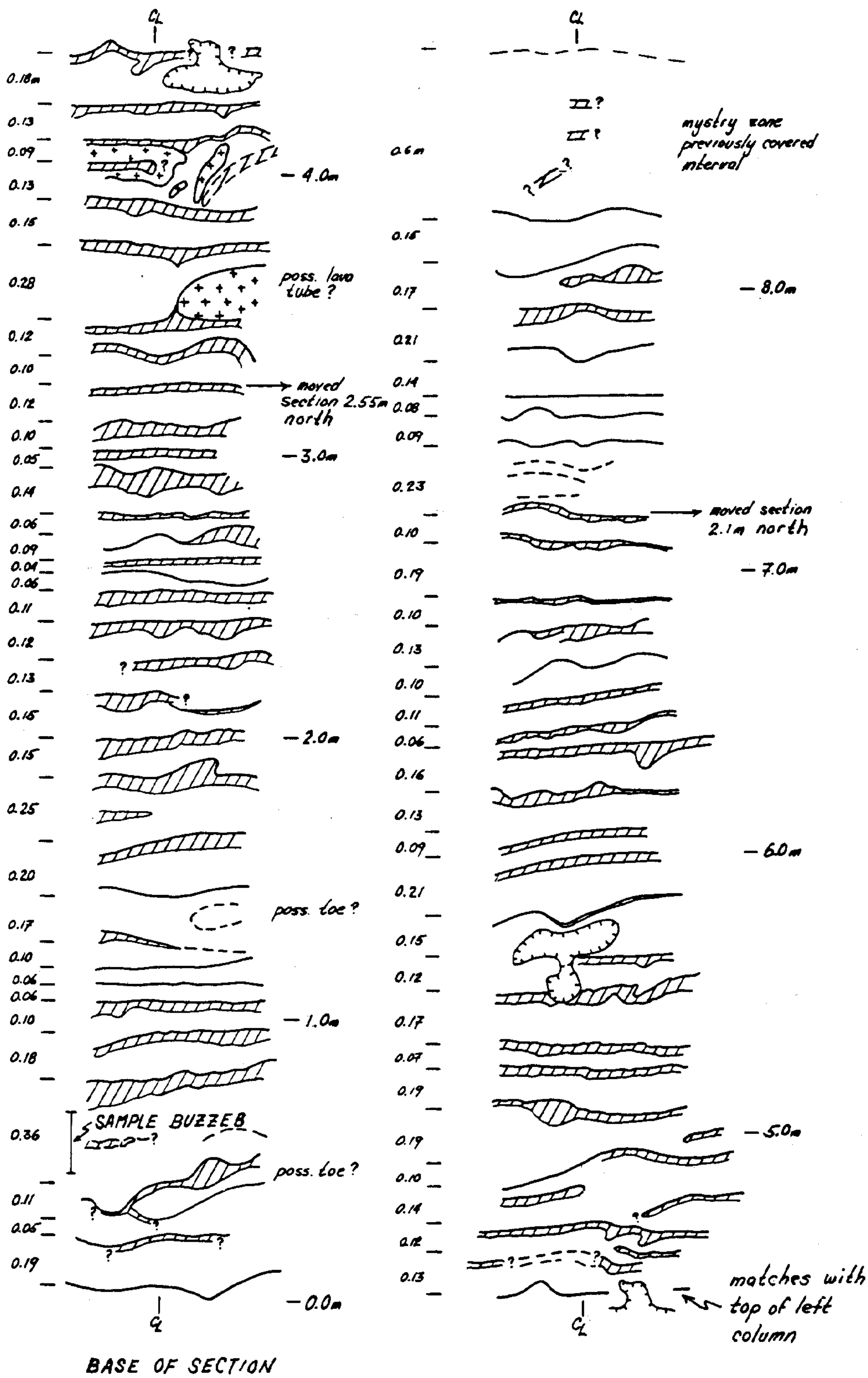
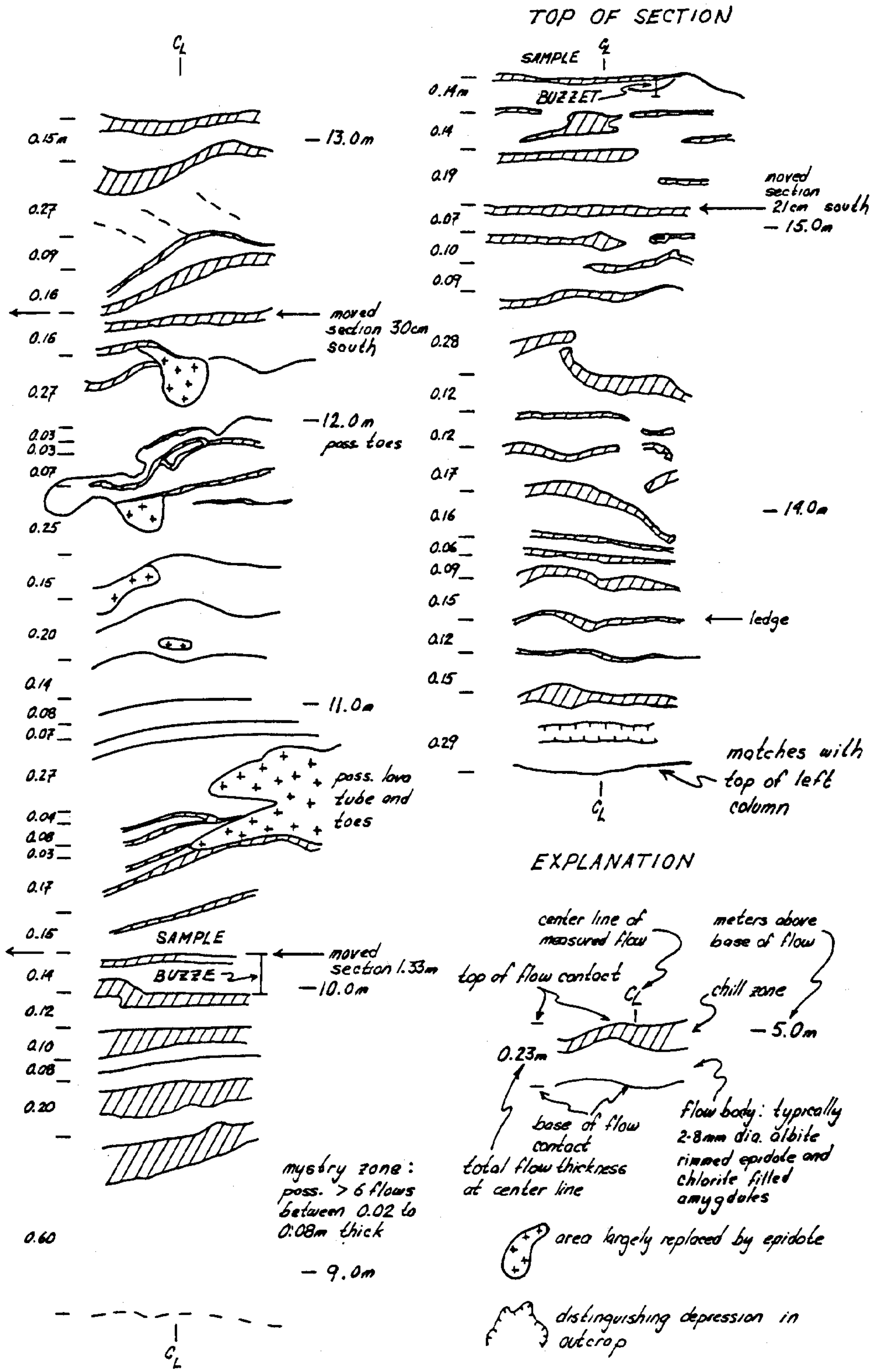


Figure 44. Stratigraphic section of metabasalt at Stop 3, west of Waynesboro Reservoir.



TOP OF SECTION
SAMPLE
BUZZET
0.14m
0.14
0.19
0.07
0.10
0.09
0.28
0.12
0.12
0.17
0.16
0.06
0.09
0.15
0.12
0.15
0.29
moved section 21cm south
– 15.0m
– 14.0m
← ledge
matches with top of left column
0.15m
– 13.0m
0.27
0.09
0.16
0.16
0.27
moved section 30cm south
– 12.0m
poss. toes
0.03
0.03
0.07
0.25
0.15
0.20
0.14
0.08
0.07
– 11.0m
0.27
poss. lava tube and toes
0.04
0.08
0.03
0.17
0.15
SAMPLE
BUZZE
moved section 1.33m
– 10.0m
0.14
0.12
0.10
0.08
0.20
mystry zone: poss. > 6 flows between 0.02 to 0.08m thick
0.60
– 9.0m
EXPLANATION
center line of measured flow
meters above base of flow
top of flow contact
chill zone
– 5.0m
0.23m
base of flow contact
total flow thickness at center line
flow body: typically 2-8mm dia. albite rimmed epidote and chlorite filled amygdules
area largely replaced by epidote
distinguishing depression in outcrop

(7) Wir verstehen nur Bahnhoff. However, it could involve thrusts of the type mapped by S. I. Root and D. B. MacLachlan along the northern margin of the South Mountain anticlinorium. In the area of Stops 3 and 4, thrusts might be concealed within the Loudoun and dips to the west cannot be be ruled out. Second-order folding is also possible.

8) By comparison with better preserved epidote bodies elsewhere in South Mountain some are tentatively interpreted as pahoehoe flow feeder-distributary tubes and some as toes. Epidotization appears to be a replacement phenomenon based on on sparse relict vesicles in some.

		LEAVE STOP 3. RETURN TO OLD FORGE ROAD.
1.0	81.1	**TURN LEFT** onto Old Forge Road.
2.1	83.2	Appalachian Trail crossing.
0.6	83.8	**TURN LEFT** into Old Forge State Forest Picnic Area just past church on right. **LUNCH.**
		LEAVE picnic area. **PROCEED BACK TO OLD FORGE ROAD.**
0.3	84.1	**TURN RIGHT** onto Old Forge Road (back the way came).
0.1	84.2	**BEAR LEFT** following Old Forge Road.
0.3	84.5	Appalachian Trail crossing.
1.9	86.4	**TURN LEFT** onto gated state forest road. Proceed up road about 1 mile to **STOP 4.** This gate is normally closed to all traffic. The alternate route to Stop 4 is to proceed ahead on Old Forge Road to the pipeline crossing, park and proceed on foot up the pipeline to its intersection with the road.

STOP 4. SNOWY MOUNTAIN. CATOCTIN METABASALT, LOUDOUN AND WEVERTON FORMATIONS

Discussants: Volcanics--R. C. Smith, II and S. W. Berkheiser, Jr.
Clastics---W. D. Sevon and N. Potter, Jr.

Although lacking in outcrops of Catoctin Metabasalt, the Snowy Mountain area provides an opportunity to examine metabasalt float. The metabasalt float contains a variety of pyroclastic fragments in what appears to be one of the uppermost preserved volcanic units. Also available for observation, are a few float blocks of Loudoun slate and conspicuous blocks of cross-bedded Weverton sandstone.

Caution is advised as you proceed north from the buses at the sawmill site (Figure 45). Although none of the scaley reptiles observed were longer than the road is wide, the "puff adders" give a memorable performance!

As good teaching localities for metabasalt are rare, you are requested to please not hammer on either the described blocks lettered A through D or those greater than 0.5 m in size unless otherwise noted.

The Snowy Mountain stop area is located on the east flank of Snowy Mountain at an elevation of 1775$\pm$25 feet, latitude 39°50'02"N, longitude 77°28'30"W, or roughly 450 m NNW of the abandoned sawmill site on the south side of the west branch of the headwaters of Hayes Run, Quincy Township, Franklin County, in

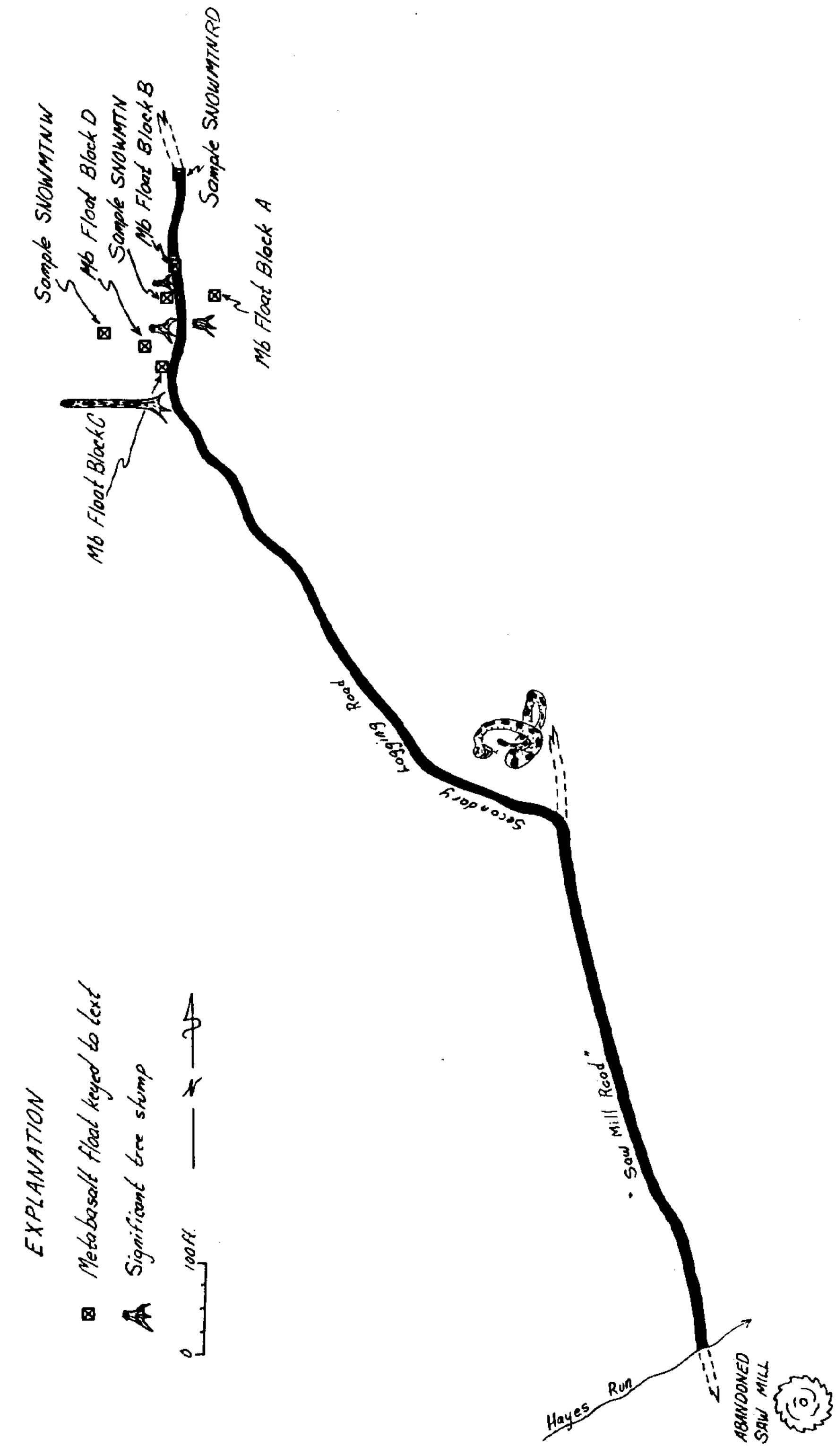


Figure 45. Location map of Stop 4, east of Snowy Mountain, showing the relative locations of the metabasalt blocks to be observed.

the Iron Springs quadrangle.

The geology of the area of Stops 3 and 4 shown in Figure 43 is based on Fauth (1978) and reconnaissance of the volcanics. Because we did not observe evidence of metabasalt being intercepted either in the pipeline located less than 300 m S of the west headwaters of Hayes Run or on the benches farther south, Figure 43 shows less metabasalt than Fauth (1978) did. However, we have not reconnoitered all of the colluvium-covered benches in the area and Fauth's map may yet prove to be correct.

Be that as it may, you are encouraged to examine each of the lettered float blocks (Figure 45) in random order. We trust that all will have a chance to observe amygdules with various mineral fillings and a variety of macroscopic pyroclasts, possibly including some altered glassy fragments. (See Plate I of 'The Catoctin Metabasalt Story,' this guidebook).

Some questions to ask include:

(1) Was the eruption subaerial or subaqueous?
(2) How close was the vent?
(3) Were the last-gasp Catoctin volcanics in Pennsylvania always basaltic?
(4) Why have the primary textures been preserved in this area, when sheared, chloritic metabasalt and epidosite are the norm?

Snowy Mountain Answer Key

(1) The absence of both pillows and extensive fragmentation, presence of traceable vesicular zones, and rounded aspect to most bombs is interpreted to suggest subaerial eruption.
(2) Presumably within ballistic range because of the agglutinate, but a 300 m-high lava fountain wouldn't be unheard of. The unwelded pyroclasts may have been transported after splashdown via conveyor-belt like pahoehoe flows.
(3) To the south of U.S. Route 30 (= Carbaugh-Marsh Creek Fault) perhaps 80 percent of the apparent uppermost Catoctin is metabasalt and 20 percent metarhyolite, but to the north, the percentages are roughly reversed. In the southwestern portion of the Dickinson quadrangle, the uppermost volcanics are bimodal and are intimately intermixed.
(4) Metamorphic grade and structural complexity appear to increase toward the southeast. This answer is a bit circuitous, but it seems likely that the eastern margin has received a Mesozoic structural overprint (Fauth, 1978; S. I. Root, personal communication, 1991). Also, the thickness of the sedimentary and structural cover may have been least on the west side of the anticlinorium.

Descriptions of significant float blocks
Please do not hammer on blocks A, B, C, and D.

Block A contains vesicular pyroclasts in the 1 to 8 cm size range (lapilli to bombs) including a 1 cm black, microvesicular lapillus. Also present are albite-rimmed amygdules, many in the

0.2 to 2 cm size range being spherical, and a few having epidote or chloritic cores. Even though this is a float block, you may wish to identify primary bedding in the flow and one cleavage.

Block B appears to consist of an agglomerate-agglutinate of vesicular-cored cow-pie bombs--spatter which originally had glassy rims but which are now altered to a reddish, jasper-like material. The amygdules in Block B are up to 4 mm in diameter and appear to be filled with pink and white albite. Lying as it does in the road, this block has been abused and lacks aesthetic appeal. However, it serves to represent a lithology which occurs within a few tens of meters of the top of the preserved volcanic section in a metabasalt valley to the east. Is there any evidence of primary plastic flow, i.e., while still partly molten, in the rims of any of the cow-pie bombs?

Block SNOMTNRD is a thin-sectioned but as yet unanalyzed possible outcrop, the texture of which suggests that it might be an intrusive. It contains epidote, chlorite, possible relict olivine, abundant skeletal opaque oxides, and trace titanite in a matrix of predominantly acicular green amphibole. Epidote-rimmed, chlorite-filled amygdules are present but rare. The skeletal opaque oxides suggest an initially medium-grained rock. Could the block be a feeder to one of the presumed lava fountains or cones? You are welcome to sample this rock.

Block SNOMTN is a sampled and analyzed metabasalt that was collected to represent the fresh, relatively non-vesicular fine-grained phase. A few, angular, epidotized autolith(?) or pyroclast blocks and lapilli up to about 10 cm were observed, but not included in the analyzed sample. The block contains sparse epidote and chlorite-filled amygdules up to about 1 cm in diameter. A thin section reveals on the order of 10 percent relict probable olivine in a microlitic to fine-grained diabasic matrix that is partly chloritized and epidotized. Like all of the SNOMTN series samples, it too could be from a hypabyssal intrusive. An analysis shows that the rock contains: 46.2% SiO_2, 14.9% Al_2O_3, 13.6% ΣFe as Fe_2O_3, 6.3% CaO, 7.5% MgO, 3.7% Na_2O, 0.8% K_2O, and 2.2% TiO_2. A C.I.P.W. norm of this sample shows 33 weight percent albite and 14 weight percent olivine (J.H. Barnes, personal communication, 1990) which with other available data suggests an olivine or alkali-olivine basalt. You are free to sample this rock. The loose pieces on top will most closely match the reported analysis.

Block SNOWMTNW is a sampled and thin-sectioned but unanalyzed block of very fine-grained, bluish-gray, fresh basalt-diabase from the highest observed float block. Abundant, very fine-grained skeletal opaque oxides and probable relict olivine microphenocysts occur in a microlitic matrix. You are free to sample this rock.

Block C appears to contain both vesicular and hollow pyroclasts. The hollow pyroclasts include a possible 13 cm spindle bomb and a dumbell pair of hollow lapilli dumbells with a total length of 2 cm. Note the 1 mm to 4 cm quartz-rimmed amygdules and partial epidote and hematite filling in a large, flattened amygdule. The origin of the hematitic streaks-filaments is unknown.

Block D appears to contain a row of lapilli with partial black microvesicular rims and epidote cores. A few other blocks of possible agglomerate-agglutinate occur nearby.

LAUDON FORMATION

In addition to the metabasalt materials present at Stop 4, there are also numerous blocks of Laudon Formation which have moved down hill from outcrop. All Laudon sites are referenced in relation to metabasalt site SNOMTNRD.

Up the road 110 feet from SNOMTNRD is a block in the road bed which displays crossbedding and fining upward from granules to fine grained. A good place to practice determination of right side up.

Fifty feet up the road from SNOMTNRD along the uphill side of the road is a block which has cross bedding and lots of granules. This is a good place to examine the composition of the Laudon Formation. From this block step into the brush on the uphill side of the road and encounter a block stream oriented diagonally to the road.

The block stream has numerous large blocks some of which display well-developed bedding, clasts up to 10 cm long, and some crossbedding. The block stream may be followed downslope to the road near site SNOMTN.

Although the block stream is somewhat obscured by vegetation, it is a good example of the periglacial activity discussed by Clark elsewhere in this guidebook.

		LEAVE STOP 4. RETURN TO OLD FORGE ROAD. TURN LEFT onto Old Forge Road.
0.5	86.9	Pipeline crossing. Alternate route to Stop 4.
1.2	88.1	**STOP SIGN. TURN LEFT** onto SR 2024. Proceed through the village of South Mountain.
1.1	89.2	South Mountain Restoration Center on left.
0.6	89.8	**TURN RIGHT** onto PA Route 233 N.
4.3	94.1	**STOP SIGN. TURN LEFT** onto US Route 30 W.
1.6	95.7	Water treatment plant on right.
0.4	96.1	**TURN LEFT** onto Mt. Cydonia Road to Mt. Cydonia Sand. Red sign on left.
0.5	96.6	Road forks. Stump Run Road goes left. **GO RIGHT.**
0.4	97.0	Pavement ends. **CONTINUE STRAIGHT AHEAD.**
0.1	97.1	Road forks at sand piles on left. **BEAR RIGHT.**
0.2	97.3	**TURN LEFT** onto Kettlespring Road.
0.3	97.6	**STOP** at road intersections. Go left into quarry with mega-ripples. Go right into quarry with weathering and colluviation. Permission to visit this site must be obtained from Valley Quarries, Inc. See mileage 99.8.

STOP 5. MOUNT CYDONIA QUARRY, VALLEY QUARRIES, INC.

Discussant: W. D. Sevon and Randall L. Van Scyoc

This stop has two sites to visit: (1) the mega-ripples and

(2) the saprolite/colluvium. The mega-ripples are to the left at the road intersection and the saprolite/colluvium is to the right. The buses will drive to the mega-ripple site, unload and then return to the road intersection for parking. After participants have examined Site A they will walk to Site B before boarding the buses.

SITE 1: MEGARIPPLES

A magnificent exposure of large sand ripples, megaripples (Figure 46), is presently exposed at this site. These megaripples occur near the middle of the Early Cambrian Antietam Formation (Figure 47).

The Antietam Formation at Mount Cydonia is a clean, coarse-grained, quartzose sandstone. The sand grains are almost entirely subangular to well-rounded quartz grains, accompanied by rare grains of chert. Quartz overgrowth on the grains is common and these, as well as intergranular quartz, serve as the binder for the rock. The formation has two lithologic subdivisions. The lower part is a resistant, bluish to pink quartzite which varies from structureless, to plane bedded, to thin beds with small-scale, multi-directional cross bedding. The upper part consists of white to pinkish sandstone containing abundant *Skolithos* tubes (fossil animal burrows). The prominent megarippled surface (Figure 46) is the approximate boundary between the two rock varieties. The Antietam is about 800 feet thick in the quarry and varies regionally from 700 to 900 feet thick (Stose, 1932).

The Antietam Formation was folded during the post-Ordovician Taconic Orogeny which formed the South Mountain anticlinorium and gave the present 70^{o} northwest dip to the megarippled beds. This deformation also imparted to the Antietam a spaced cleavage which dips perpendicular to the bedding and strikes approximately parallel to bedding.

The trace fossil *Skolithus linearis* (see Figure 2, page 25) occurs in profusion in the upper part of the Antietam and superb examples can be seen in unquarried rock faces and in numerous quarried blocks. These trace fossils represent the vertical borings of an animal whose morphology is unknown, probably because the animal had no hard parts which could be fossilized. In weathered rock *Skolithos linearis* appears as a remarkably straight, sand-filled tube which is circular in cross section if undeformed. However, many of the tubes in this quarry have elliptical shapes because of deformation. The long eliptic axis is parallel with the strike of bedding. *Skolithus* is not readily discernable in fresh rock, but is quite apparent on weathered surfaces. These tubes are thought to have been formed by a worm that lived in the tube and fed at or near the sediment-water interface. *Skolithus linearis* is found almost exclusively in sandstones and is one of a suite of trace fossils interpreted to be associated with dominantly high energy shallow marine depositional environments such as offshore bars and beaches and, to a lesser degree, deeper offshore sediments (Banks, 1970). Its presence, as well as rare marine fossils reported from other localities (Fauth, 1968), establishes a marine origin for the

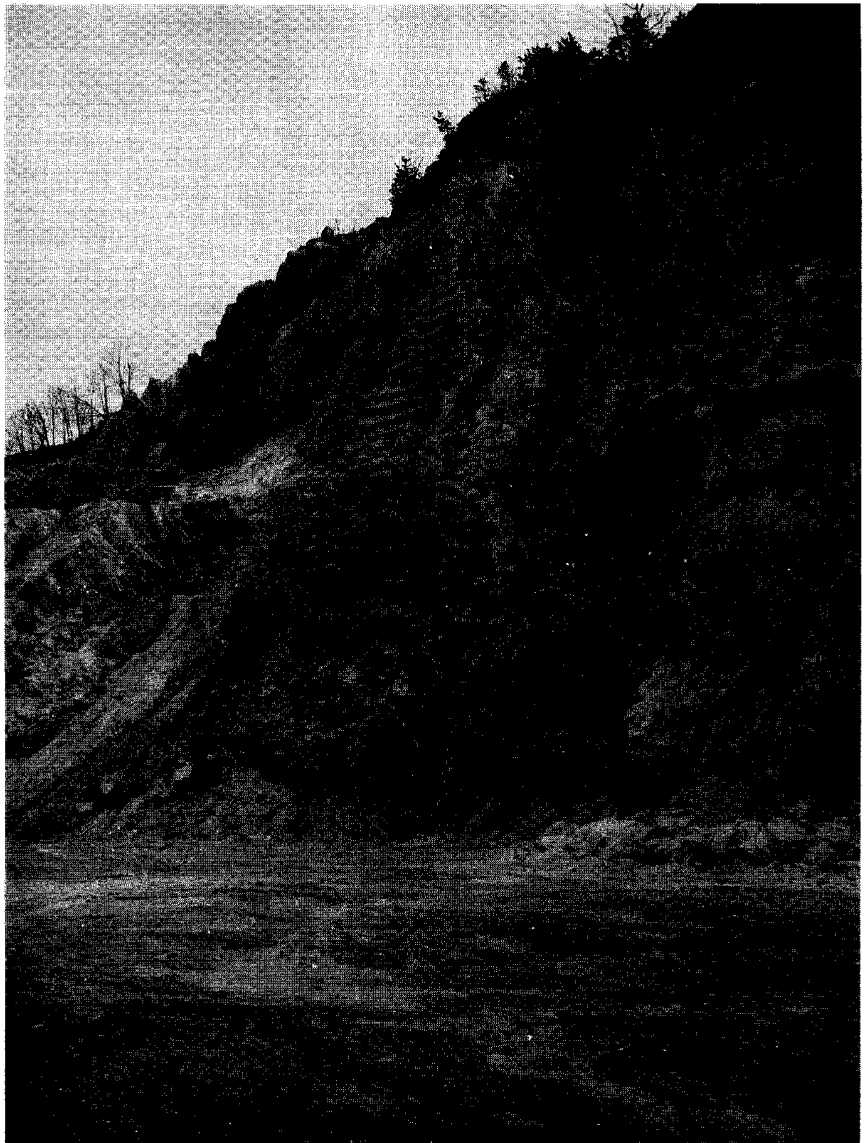


Figure 46. Corrugated surface of megaripples in Lower Cambrian Antietam Quartzite.

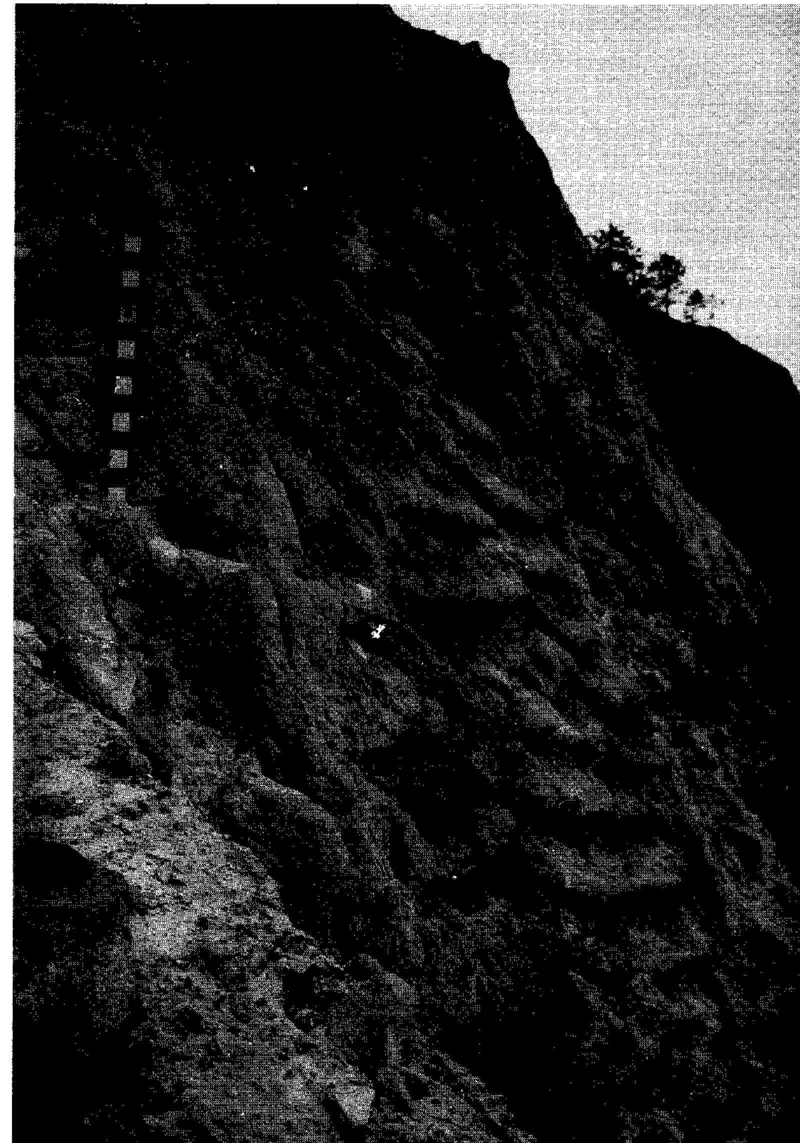


Figure 48. Asymmetrical megaripples in Antietam Quartzite. Scale interval: 10 cm.

Antietam Formation (see also Key, this guidebook).

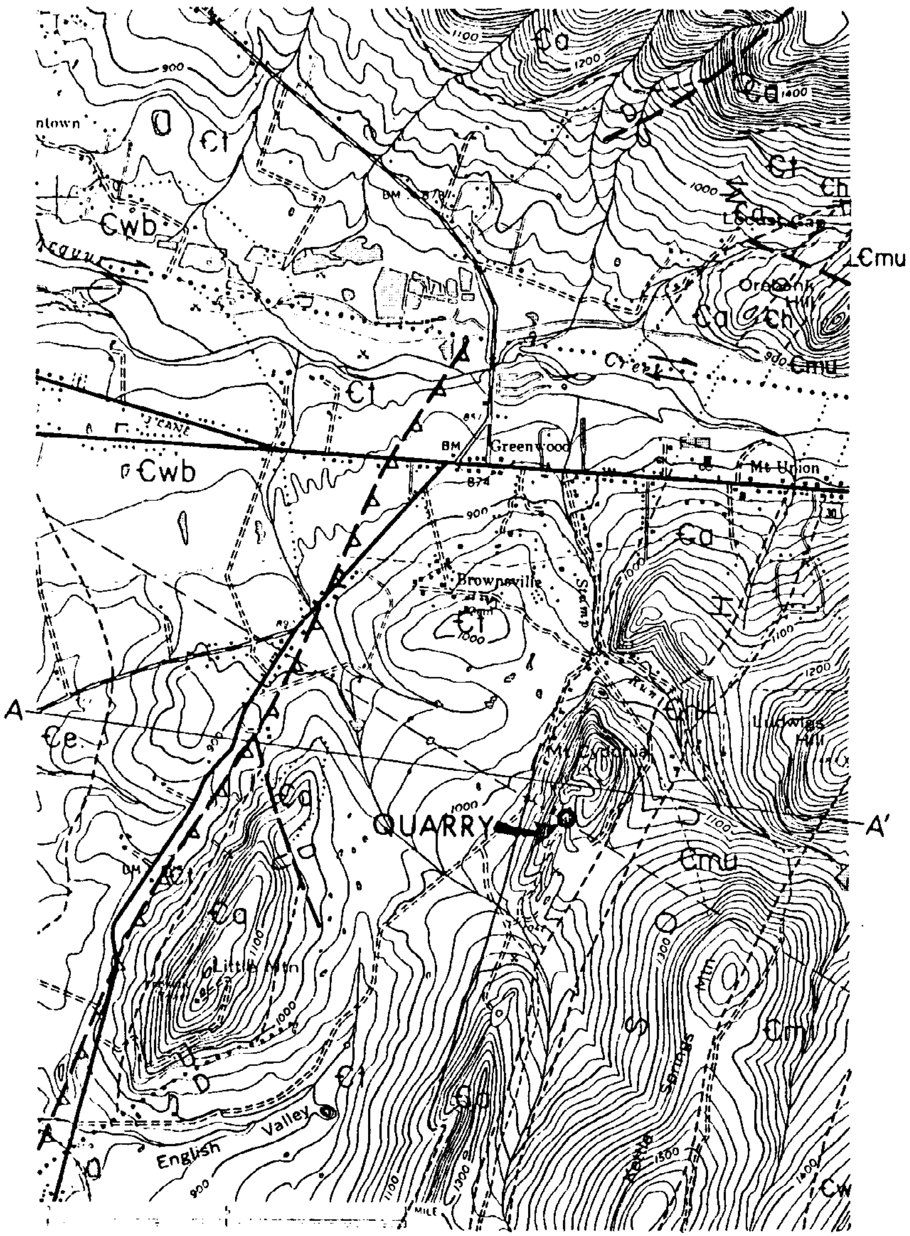
The megaripples which occur on the prominent surface have amplitudes of 12 to 25 cm and wave lengths of 45 to 75 cm (Figure 48). The megaripples are asymmetrical in cross section with steep face to the northwest and shallow face to the southeast. Wilshusen and Sevon (1981) described the megaripples as having unidirectional forset layers which dipped northwest prior to folding. They saw this orientation in a cross section exposure of the megaripples which no longer exists. Similar orientations occur in the present outcrop but are not prominent. Individual megaripples are laterally persistent and vary from linear to slightly sinuous (Figure 46). The crests of the megaripples are rounded.

A cross section of a second layer of megaripples occurs below the spectacular exposure and was not exposed for examination by Wilshusen and Sevon in 1981. This cross section shows the following sequence (from bottom to top): (1) evenly bedded sand layers undulating in what appears to be hummocky cross stratification, (2) a thin layer of sand with unidirectional southeast dipping forset beds with megaripple form on the upper surface, and (3) a thick bed of planar bedded sand which infills the ripples of the underlying bed. The forset beds comprising the megaripples in this lower unit presently dip 22°NW while bedding dips 53°NW; thus the original dip of forset bedding was to the southeast. The overlying planar laminated unit (3) is composed of laminae ranging from fine-grained to very coarse-grained sand with no regularity of variation. Although the interval between the lower and upper megarippled surfaces is not totally clear, it appears that above unit (3) the sequence is repeated by returning to hummocky cross stratification.

Interpretation

Megaripples are formed in sand under water by any of three mechanisms: (1) by unidirectional flow of water, such as in a stream, (2) by oscillatory motion of waves in shallow waters of oceans or lakes, and (3) by a combination of (1) and (2), such as in a tidal channel. A wave-generated origin in which oscillatory flow is stronger in one direction, generally landward, than the other was suggested as a probable origin by Wilshusen and Sevon (1981). However, the new exposure suggests a more complex and different origin.

The presence of stratigraphically close, well-developed megaripples with a 180° difference in orientation argues that they are each the result of unidirectional flow in opposite directions. Kaufman and Frey (1979) suggested that the character and areal distribution of the Antietam Formation is well explained if the original sand was deposited as an offshore ridge (bar). Such coastal features commonly have either tidal or storm-generated inlets/outlets in which currents will flow in opposite directions at different times. The structures present in this quarry appear to fit this offshore bar environment and may provide additional proof of the Kaufman and Frey model.



From Fauth, 1968

LEGEND

Ꞓe
ELBROOK FORMATION
Gray limestone and calcareous sandstone with finely laminated, shaly limestone in lower half.

Ꞓwb
WAYNESBORO FORMATION
Bluish limestone and dolomite with interbedded sandstone and shale.

Ꞓt
TOMSTOWN FORMATION
Gray dolomite with interbeds of white limestone.

Ꞓa
ANTIETAM QUARTZITE
White to gray, medium- to coarse-grained sandstone and quartzite.

Ꞓh, Ꞓmu, Ꞓml
HARPERS FORMATION
Greenish-gray graywacke, Harpers Formation (Ꞓh); light-gray sandstone and quartzite, Upper Montalto Member (Ꞓmu); gray, thin-bedded sandstone and quartzite, Lower Montalto Member, (Ꞓml).

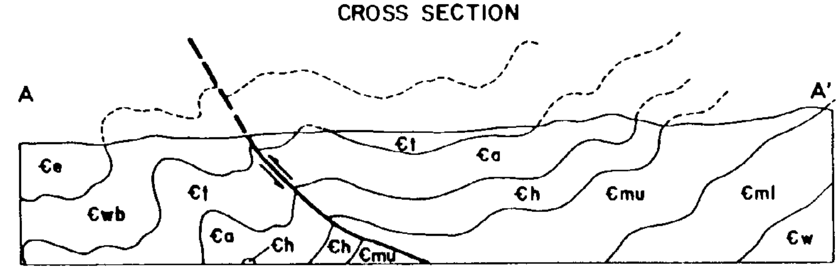
Ꞓw
WEVERTON FORMATION
Interbedded sandstone and graywacke with conglomerate and phyllite intervals.

SYMBOLS

Thrust fault

A —— A'
Line of cross section

CROSS SECTION



Scale: Same as map
No vertical exaggeration

From Fauth, 1968

Figure 47. Geologic map and cross section.

SITE 2: WEATHERING SEQUENCE

A very interesting sequence is exposed on a small vertical face on the east side of north side (below Kettlespring Road) of the southern quarry site. The face is shown in Figure 49A.

The lower part of the sequence comprises saprolite developed from Antietam Quartzite. The saprolite is white and moderately soft because the cement has been removed by weathering. Bedding has a northwest dip and crossbedding is easily seen. Numerous vertical to subvertical fractures and subhorizontal bedding planes are surrounded by zones of slightly harder, reddish yellow (7.5YR6/6), iron-coated, quartz grains. The subhorizontal zones are up to 30 cm thick; the subvertical zones, generally less than 5 cm thick. These zones are only slightly more resistant to disintegration than the white saprolite, but it is enough to allow the development of boxwork structure (Figure 49B). The vertical to subvertical zones generally have within the actual fracture a vein of white kaolinite 1 mm to 1 cm thick. The kaolinite veins are generally brecciated, probably as a result of surface weathering.

The saprolite is overlain by about 1 m of loose to moderately coherent sand which has the form of a number of cyclic beds (Figure 49C and D). The lower part of each bed is loose, yellow (10YR8/8) sand. The grains appear to be slightly coated with iron. The loose sand becomes more coherent upwards as the pore space between grains is filled with white clay, presumably kaolinite. The basal contact between the loose sand and that with clay is fairly sharp, but the amount of clay at the base of the upper zone is small. The amount of clay increases upwards and the interstices at the top of white (10YR8/2) zone are almost totally filled with clay. The upper contact of the bed is sharp and the cycle is repeated. The cyclic beds are 2-15 cm thick and have great lateral discontinuity with some beds being cut out by higher beds. Internally the beds are structureless, planar laminated, and crossbedded. There are occasional small pieces of white kaolinite in the lower part of beds. The total composite of cyclic beds is about 1 m thick. In the uppermost 10-15 cm there are occasional rounded quartzite pebbles and more pieces of kaolinite. The upper contact is sharp.

The upper unit is a diamict composed mainly of quartzite pebbles and cobbles less than 10 cm in diameter with an abundance of clasts less than 5 cm in diameter. There are some clasts up to 25 cm in diameter. The clasts are rounded to subangular in shape. There is no apparent structure in the diamict. A soil zone occurs at the top of the diamict, but it has not been examined.

Interpretation

The Antietam Quartzite was weathered deeply to produce a saprolite. During the process of weathering, minor amounts of iron were produced and migrated downward along the vertical to subvertical fractures and then laterally along the subhorizontal bedding planes. The iron was precipitated on the quartz grains adjacent to the fractures and bedding planes and cemented them together slightly. Intensive weathering also produced kaolinite as an end product clay. This clay was transported down the vertical and subvertical fractures and was deposited there. In the outcrop viewed here the iron-coated grains and the kaolinite appear to be disappearing in the lower part of the outcrop and becoming thicker and better developed in the upper part. This suggests that the outcrop represents the lower part of the downward zone of iron and kaolinite migration which may have been controlled by the groundwater table. Following development of the saprolite the area was eroded to produce the sloping surface on which the overlying materials were deposited.

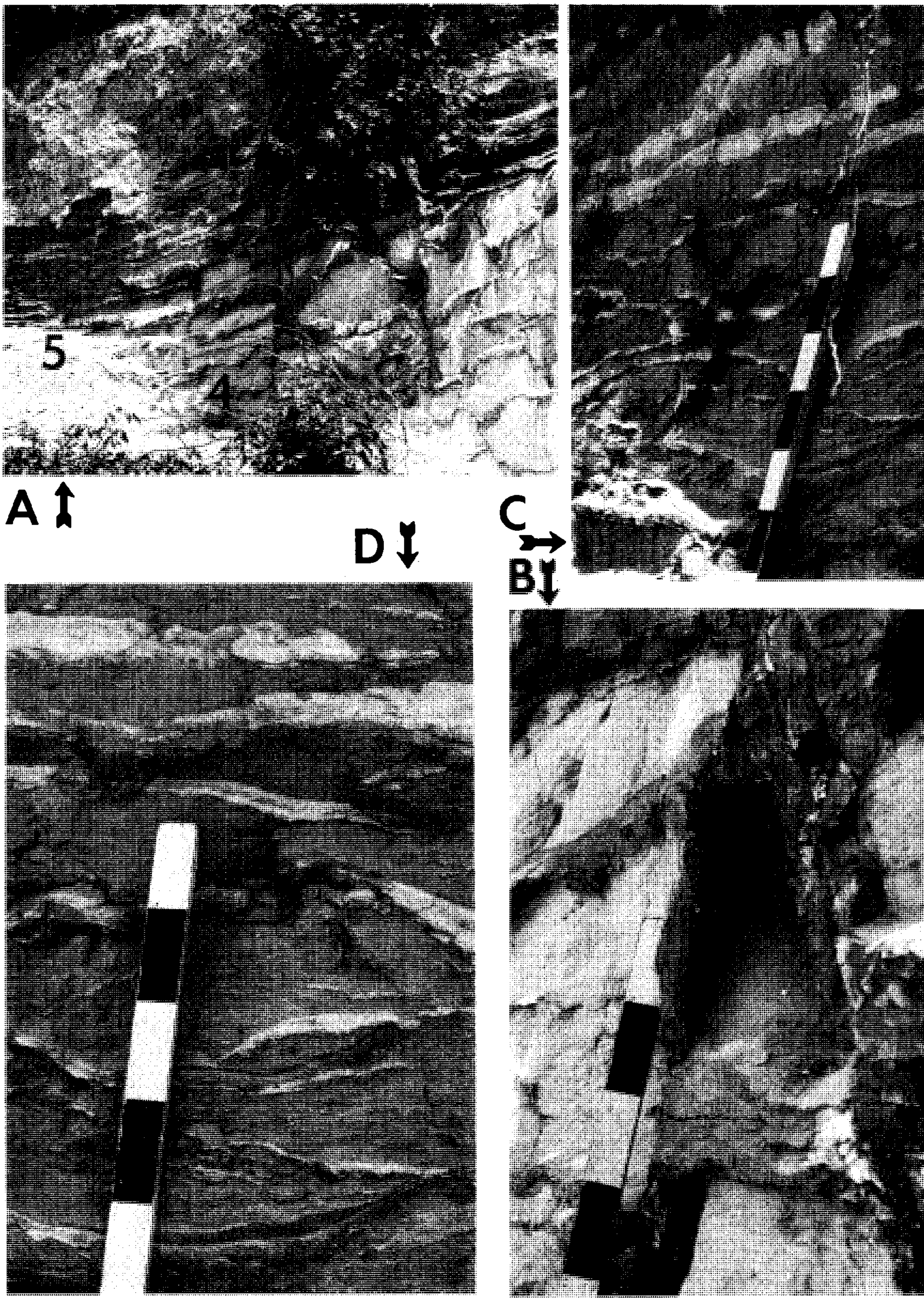
The cyclic sand and clay/sand beds overlying the saprolite may represent surface-wash deposits of eroded saprolite. Each bed presumably represents a significant depositional event. Following the deposition of the sand bed, slope wash for an extended period of time was capable of transporting only clay (kaolinite) which gradually accumulated within the pore space of the sand until the pores were nearly filled. Another significant event transported sand and deposited a new layer of sand as well as often eroding some of the underlying material. This was followed by clay deposition. The cycle was repeated several times.

The diamict represents colluvium transported downslope from rock outcrop exposed up slope after the saprolite had been eroded. The mechanism of colluvial transport is not certain, but solifluction is a possibility.

The timing of this sequence is uncertain, but it could be: saprolite development, middle Tertiary: slope erosion, Middle and Late Miocene; deposition of cyclic sand and clay/sand beds, Pliocene and early Pleistocene; deposition of diamict, Late Pleistocene. It is also possible that everything subsequent to the weathering took place in the Pleistocene.

		LEAVE STOP 5. PROCEED BACK THE WAY CAME on Kettlespring Road.
0.3	97.9	**TURN RIGHT** onto Irishtown Plantation Road.
0.1	98.0	**BEAR LEFT.**
0.2	98.2	**PROCEED STRAIGHT AHEAD** on paved road.
0.9	99.1	**STOP SIGN. TURN LEFT** onto US Route 30 W.
0.2	99.3	**TURN RIGHT** onto PA Route 997 N.
0.5	99.8	Mt. Cydonia Sand processing plant on left. Office here is the place to obtain permission to visit Mt.

Figure 49. Opposite page. **A.** Outcrop of saprolitic Antietam Quartzite with boxwork structure, cyclic surface-wash beds, and colluvium at southern Mount Cydonia quarry (see Figure 47). 4 = site of Figure 49B; 5 = site of Figure 49C and D. **B.** Boxwork structure developed in saprolitic Antietam Quartzite at Mount Cydonia southern quarry (see Figures 47 and 49A for locations). Scale divided into 10 cm intervals. **C.** Cyclic sand and clay/sand beds at Mount Cydonia southern quarry (see Figures 47 and 49A for locations). Scale divided into 10 cm intervals. **D.** Detail of cyclic sand and clay/sand beds shown in Figure 49C. Scale divided into 10 cm intervals and is in same position as scale in Figure 49C.


5
4
A
C
D
B

		Cydonia and Mainsville quarries. Now retracing some of earlier route. Note gravels from South Mountain underlie most of the surface.
3.5	103.3	**TURN RIGHT** onto White Church Road just past brick Mount Pleasant church on left and white house on right. Cemetary on left after turn. There is lots of gravel along this route.
1.3	104.6	Orchard on right just past a white house on a curve. Criders Pond is on the slopes of South Mountain be-this orchard. See Clark, this guidebook for a short discussion of this significant site.
0.6	105.2	**BEAR RIGHT** following White Church Road. Do not go straight.
2.8	108.0	**BEAR RIGHT** following White Church Road. Hill on left ahead is composed of limestone. Some outcrops occur along this road stretch.
1.0	109.0	**STOP SIGN. TURN RIGHT** onto Mainsville Road in Mainsville.
0.1	109.1	**PROCEED STRAIGHT AHEAD** on Sandy Lot Road.
0.8	109.9	**TURN RIGHT** to Sand Bank Orchards and Valley Quarries Mainsville #2. **MAKE AN IMMEDIATE LEFT** through gates which are closed when quarry is not operating.
0.5	110.4	**CONTINUE STRAIGHT AHEAD** through gates.
0.1	110.5	**TURN RIGHT** into quarry. Permission to visit this site must be obtained from Valley Quarries, Inc. at their office at Fayetteville, see mileage 99.8.

STOP 6. MAINSVILLE QUARRY, VALLEY QUARRIES, INC.

Discussant: W. D. Sevon, Randall L. Van Scyoc, and Douglas C. Chichester

The Mainsville quarry of Valley Quarries, Inc. was opened in 1989 in the diamictons which form an apron of unconsolidated sediment along the north margin of South Mountain. See the paper by Van Scyoc for a description of the quarry operation. The site descriptions presented here are based on exposures available in June and July, 1991.

DIAMICTON

The Mainsville quarry is in the proximal part of an apron of unconsolidated sediment which is continuous along the length of the north and west side of South Mountain from Brandtsville east of Boiling Springs (Stop 7) to the Pennsylvania-Maryland state line, a distance of 50 miles (Figure 12, p. 61). The materials continue southward, either continuously or discontinuously through Maryland and along the length of the Blue Ridge in the Shenandoah Valley of Virginia (e.g., Duffy and Whittecar, 1991). The apron has the surface form of a series of coalescing fans each of which has its apex at the mouth of a small drainage basin eroded into the Cambrian clastic rocks of South Mountain (Figure 50). The surface of the apron is generally quite irregular and undrained depressions are common. "Islands" of bedrock occur within the apron (Figure 51), and the distal part of the apron usually terminates at a stream flowing parallel to the mountain front, such as Yellow Breeches Creek between Walnut Bottom and Brandtsville and Antietam Creek north of Waynesboro. Streams emanating from the margin of South Mountain usually traverse a channel which has been cut several meters into the original upper surface of the fan. Streams which cross the fans flow at the surface during the wetter and cooler parts of the year, but generally dry up by mid-summer. Because most of the streams within the drainage basins do have flowing water year around, subsurface drainage in the fan area must be assumed.

The bulk of the material present in the Mainsville quarry is properly termed a diamicton, an unconsolidated, unsorted accumulation of material ranging in size from clay to boulders. Figure 52A and B show a presumably representative part of the deposit. These photographs were used to calculate the area of various sizes of clasts present. The assumption is made that the surface area approximates the volume and the following generalities are made about the size distribution of this part of the deposit:

25 % clasts > 10 cm diameter
10 % clasts 5-10 cm diameter
15 % clasts 2.5-5 cm diameter
10 % clasts 1-2.5 cm diameter
30 % > 200 mesh (0.074 mm) and < 1 cm diameter
20 % < 200 mesh (0.074 mm) (from page 128)

The material < 200 mesh comprises in part, and possibly mainly, reddish brown clay which surrounds most particles. The clay is smectite with a trace of physically mixed kaolinite (x-ray analysis performed at Pennsylvania Geological Survey). There are occasional thin, lighter-colored clay laminae composed of partly dehydrated montmorillonite-chlorite mixed layer clay and physically mixed kaolinite.

The clasts are almost entirely white and weathered to some degree. Many clasts disintegrate with only a light touch, many disintegrate with moderate pressure, and a few are relatively hard. These clasts were presumably all quartzites at one time and were derived primarily from the Mont Alto Member of the Harpers Formation, although some of the clasts are derived from the Weverton Formation, and a few may be from the Antietam Formation. The clasts vary from well rounded to subangular with most clasts showing a moderate to considerable degree of edge abrasion. Clasts are equant to elongate in shape. Clasts in the present streambed of Furnace Creek have shapes and roundness similar to those occurring in the deposit and differ only in that they are hard and not weathered.

The bulk of the Mainsville deposit appears to be structureless, although thin zones of bedded sand and gravel occur locally. There is no apparent sorting of the constituent clasts in the bulk of the deposit. Three lithologic variations occur in the quarry. About 3 m below the top of the deposit in the area of the clastic dike (Figure 53), the diamicton has a change in clast content and lacks larger boulders. There is another change to even smaller-clast diamicton in the uppermost meter which truncates the clastic dike. In the back of the quarry (Figure 53) is a thick deposit which is mainly sand and material < 200 mesh. Pebbles and cobbles occur in this fine-grained deposit, but are not common. Horizontal bedding is present, but details of its character were not observed. At the time of writing the shape and extent of this unit was not known, but it was at least 20 feet thick and appeared to be a channel filling.

The thickness of the diamicton deposits bordering South Mountain is known to be variable and considerable (Becher and Root, 1981). Figure 51 shows an isopach map of the thickness of these materials in the Mainsville area and the outlines of 13 drainage basins which were eroded to form the adjacent apron. The figure also shows the contours of a surface reconstructed to indicate a possible surface of South Mountain prior to the erosion of the 13 drainage basins. This reconstruction is not to be taken as the probable topographic configuration of this part of South Mountain prior to the erosion of the drainage basins. Certainly there must have been valleys and drainage divides such as there are now, not a uniform sloping surface. The purpose of the reconstruction is only to provide a simple mechanism for comparing the volume missing from the present drainage basins with the volume of material in the diamicton apron.

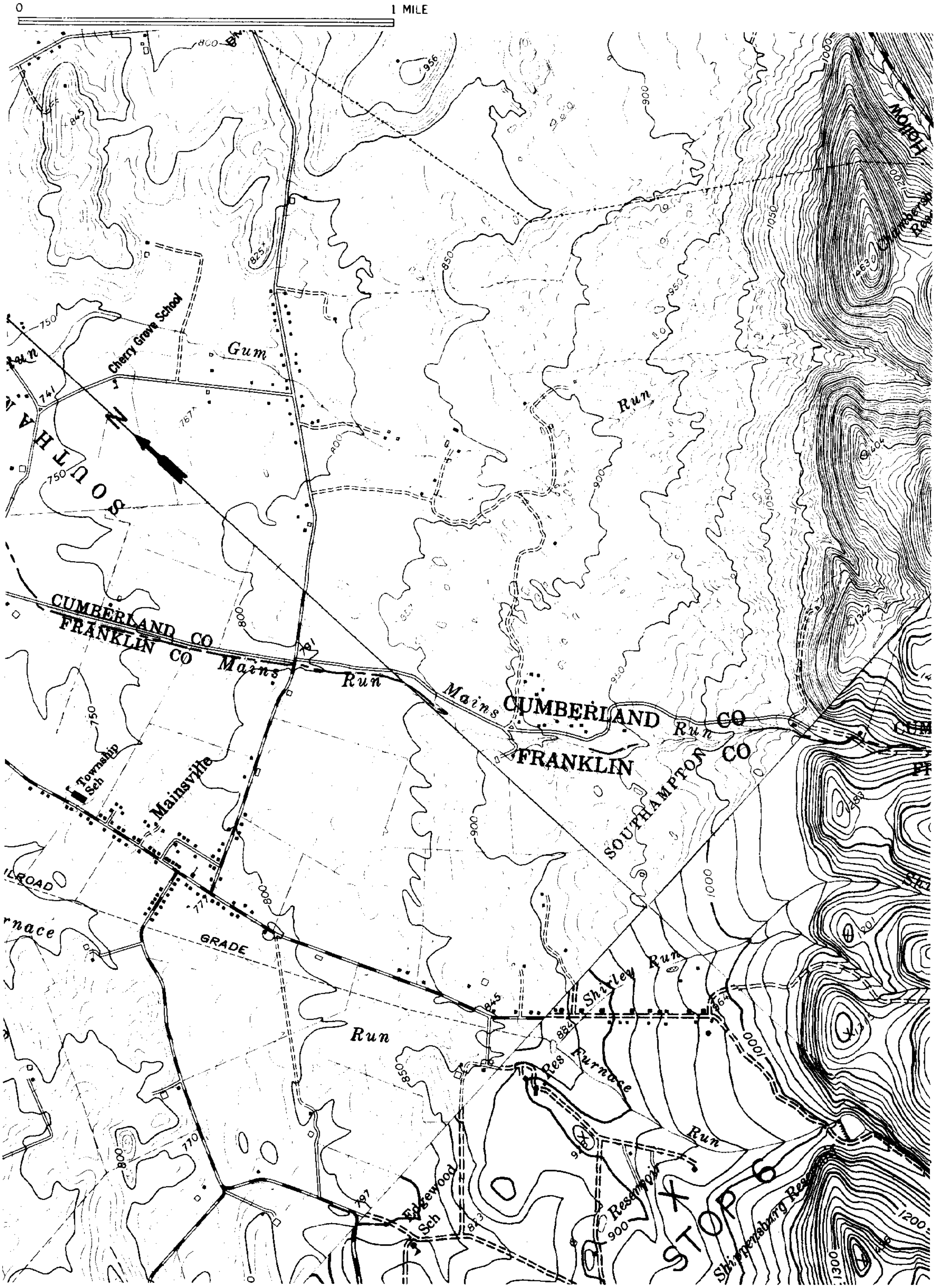
The contours are generalized, but in general correspond with elevations on the present drainage divides. Using these reconstructed contours and the contours for the present topography, the volume of material removed from the 13 drainage basins was calculated to be 0.59 mi^3. The volume of diamict in the apron was calculated from the isopach map to be 0.62 mi^3. Considering the assumptions and potential margin for error, this is remarkable agreement and seems to indicate that the material present in the diamicton apron represents that produced during erosion of the adjacent drainage basins from a level approximated now by their drainage-divide elevations. The difference in the two figures above could represent residual material produced by carbonate dissolution beneath the diamicton. Unfortunately,logs of wells drilled through the diamict are not sufficiently explicit to allow differentiation of diamicton and residuum.

CLASTIC DIKE

An outstanding feature present at the Mainsville Quarry is a clastic dike (Figure 53) which has a vertical orientation and measures 18 m deep and 31 m long of exposed material. The top of the dike has been scalped and the bottom of the dike extends below the quarry floor. The dike is between 1 and 1.5 m wide at the top and narrows to about 10 cm at the base. The narrowing is generally progressive and suggestive of a very narrow wedge. Orientation of the dike changes from N68°W (top) to N57°W to N45°W (base) as measured on the different benches.

The clastic dike is filled predominantly with laminated

Figure 50. Opposite page. Topographic map showing shape of diamict apron in the Mainsville area. Parts of Caledonia Park, Scotland, Shippensburg, and Walnut Bottom 7.5-minute quads.


0
1 MILE
N
SOUTHAMPTON
Cherry Grove School
Gum
Run
CUMBERLAND CO
FRANKLIN CO
Mains
Run
Mains
CUMBERLAND
Run
CO
FRANKLIN
SOUTHAMPTON
CO
Township Sch
Mainsville
GRADE
Run
Shirley Run
Res
Furnace
Run
Reservoir
Edgewood Sch
STOP 6
Shippensburg Res

brown clay. The color varies from strong brown (7.5YR5/8) to yellowish red (5YR5/8). The clay is well laminated with laminations oriented vertically (Figure 52C). The laminae appear nearly parallel in the lower part of the dike, but weather to suggest an irregular, almost braided pattern in the upper part. Laminae surfaces, particularly in the lower part of the dike where the clay is freshest, have slickensided surfaces and sometimes very minute striations. The clay is the same as that present in the diamicton and is smectite with a trace of physically mixed kaolinite. There are rare, thin, lighter-colored clay masses which are composed of partly dehydrated montmorillonite-chlorite mixed layer clay and physically mixed kaolinite.

Sand grains and occasional pebbles and cobbles are scattered throughout the clay. There seems to be less sand in the lowest part of the dike, but the material was not analyzed for texture at any horizon. The sand and larger clasts appear to be identical to that comprising the host diamicton.

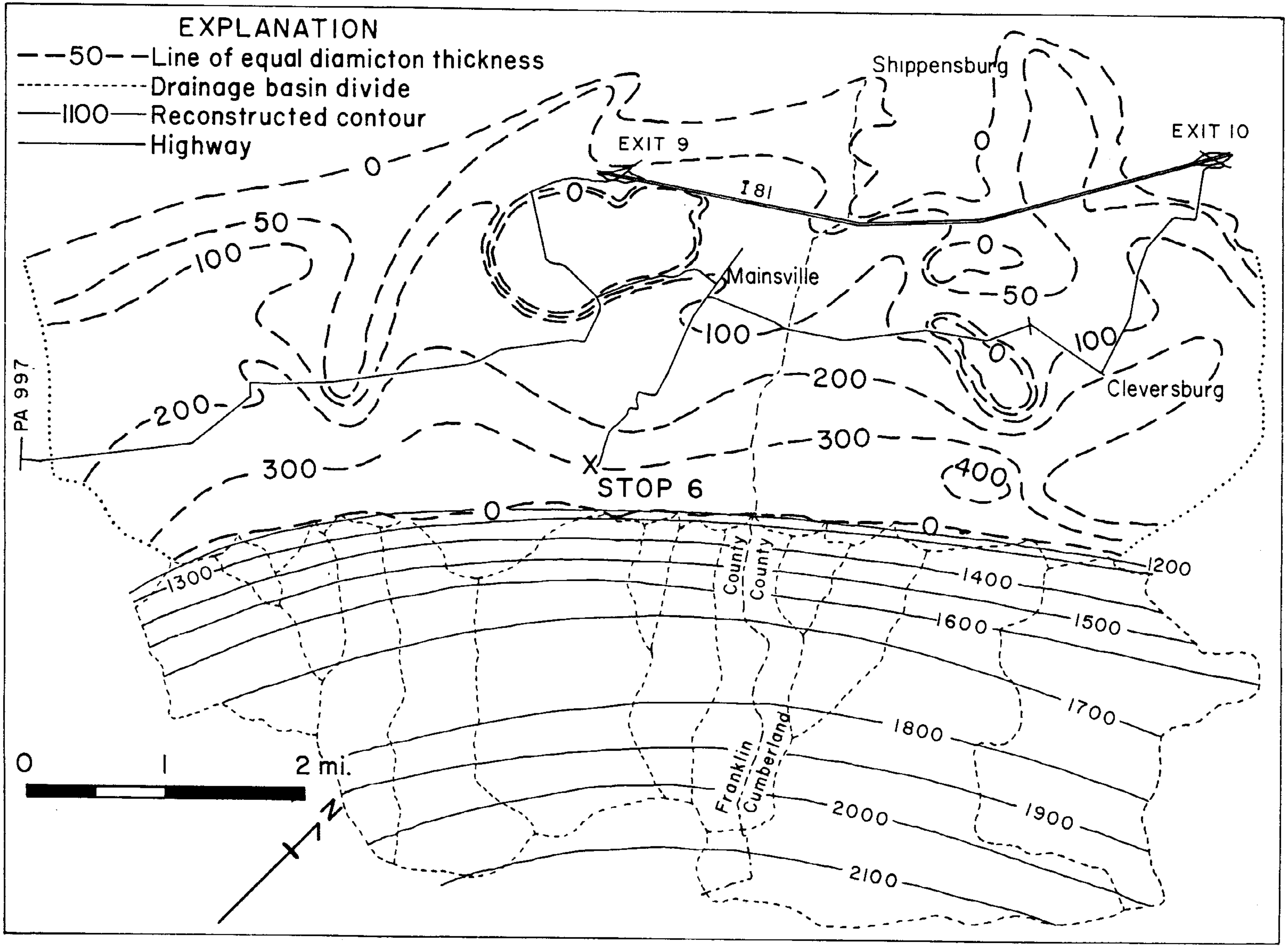
The margins of the dike appear very sharp from a distance, but are less sharp when viewed closely. Margins in the lower part are relatively sharp and there seems to be little diffusion of clay into the surrounding host. In the upper part of the dike, there is considerable diffusion of clay into the surrounding host and near the uppermost part of the dike the actual margin of the dike is vague. There appears to be no distortion either upwards or downwards of the host at the margin. The clay appears to wrap around larger clasts which protrude into the main body of the dike. At one place about 4-5 m below the top, a vertically oriented slab of host about 5 cm thick and 50 cm long is separated from the main wall and surrounded by clay.

The uppermost 1.5 m is a paleosol in which laminations have been totally destroyed by burrows and roots whose small holes riddle and homogenize the clay. The uppermost 1.5 m contains lots of sand, but almost no pebbles or cobbles. The margins are very diffuse probably because of bioturbation. This paleosol is described below. The paleosol is truncated and overlain by a meter of diamicton which has a modern soil.

Description of soil profile at top of clastic dike, Mainsville quarry, Stop 6. Roots are not noted in the description, but they are present in small numbers throughout the described unit. Acidity of the materials was not determined.

Depth (cm)	Horizon	Description
0-22	A	Very pale brown (10YR8/3) coarse to very coarse sand; structureless; loose, nonsticky to slightlysticky, nonplastic; 35-50 percent rounded to subrounded sandstone clasts generally with long

Figure 51. Opposite page. Isopach map of diamicton apron in the Mainsville area, outlines of adjacent South Mountain drainage basins, and contours of reconstructed surface (see text). Cumberland County part of isopach map modified from Becher and Root 1981.


EXPLANATION
50 Line of equal diamicton thickness
Drainage basin divide
1100 Reconstructed contour
Highway
Shippensburg
EXIT 9
EXIT 10
I 81
Mainsville
Cleversburg
PA 997
STOP 6
Franklin County
Cumberland County
0
1
2 mi.
N

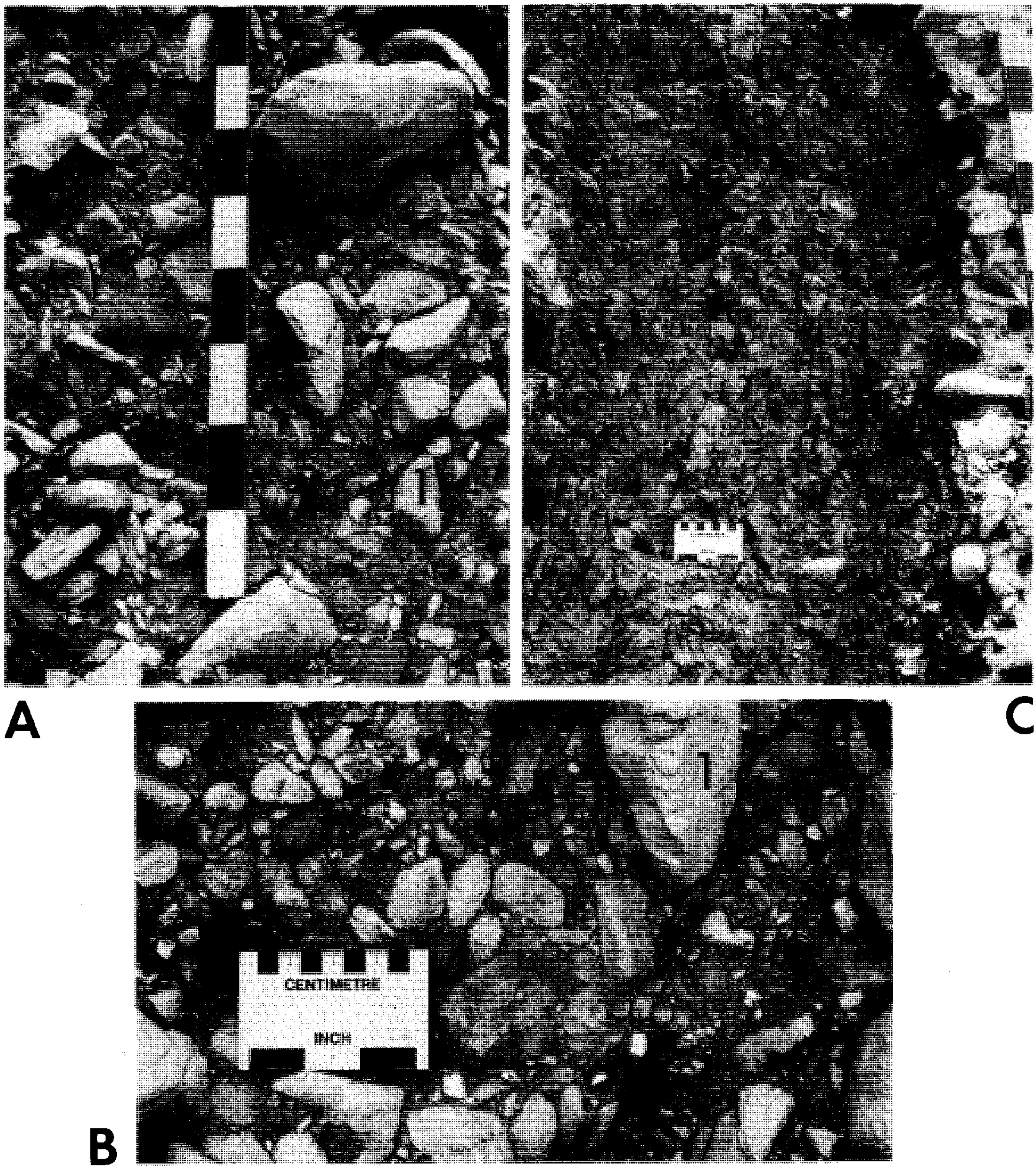


Figure 52. **A**. Photograph of diamict at Mainsville quarry. Scale divided into 10 cm units. Numbered (1) cobble same as (1) in Figure 4. **B**. Photograph of diamict at Mainsville quarry. Scale divided into centimeters (top) and inches. Numbered (1) cobble same as (1) in Figure 52A. **C**. Photograph of clastic dike at Mainsville quarry showing vertical laminations and sharp contact with host diamict. Vertical scale divided into 10 cm units. Small scale divided into centimeters (top) and inches.

		axis less than 4 cm but some up to 9 cm; top surface scalped but upper 1-2 cm has some black organic material; abrupt boundary.
22-63	Bt	Strong brown (7.5YR5/6) coarse to very coarse loamy sand; very pale brown (10YR8/3) mottles are common, distinct as subhorizontal 1/2-1 cm thick zones with less abundant clay films; structureless; slightly hard to hard, sticky, slightly plastic; 35-50 percent rounded to subrounded sandstone clasts generally with long axis less than 4cm but some up to 9 cm; abrupt boundary.
63-81	IIC1	Reddish yellow (7.5YR6/8) coarse sand; structureless; soft, nonsticky to slightly sticky, nonplastic; abrupt boundary.
81-98	IIC2	Reddish yellow (5YR6/6) loamy sand; structureless; hard, slightly sticky to sticky, nonplastic; abrupt boundary has lateral relief of 25-30 cm because of erosional nature of boundary.
98-120	IIIBtb	Red (2.5YR4/6) to yellowish red (5YR4/6) sandy clay loam; coarse prismatic structure strongly developed; hard, sticky to very sticky, plastic; rare clast up to 1 cm in diameter; clear boundary.
120-129	IIIBCb1	Yellowish red (5YR5/8) very coarse sand; some 2-5 mm diameter spots of redder or yellower colors; structureless; slightly hard, slightly sticky, slightly plastic; some clay adhering to sand grains, but real contrast to horizon above; clear boundary.
129-189	IIIBCb2	Red (2.5YR4/8) sandy silty clay loam; 5-10 percent olive yellow (2.5Y6/8) mottles are discontinuous, subhorizontal, 2-5 mm thick clay layers; structureless; slightly hard, sticky, slightly plastic to plastic; moderate clay content adheres grain to grain; prominent subvertical partings with clay films; rare clasts up to 1 cm in diameter; diffuse boundary.
189-246	IIIBCb3	Red (2.5YR4/6) to yellowish red (5YR4/6) silty clay loam; structureless; slightly hard to hard, sticky, plastic; some rounded clasts up to 10 cm in diamter; some very pale brown (5YR4/6) pieces of clay; diffuse boundary.
246-286	IIIBCb4	Yellowish red (5YR5/6) silty clay loam; structureless; slightly hard, sticky, plastic; a few discontinuous, subhorizontal, 1-2 mm thick layers of yellowish red (5YR4/6) clay; abrupt boundary.
286-306	IIIC	Yellowish red (5YR5/6) silty clay loam; vertical to subvertical laminae 5-10 mm thick; slightly hard, sticky, plastic; scattered sand grains and rare clasts; this is basically similar to material present in remainder of clastic dike; no boundary, continuous into similar ma-

terial.

INTERPRETATION

Diamicton

The surface of the diamicton apron bordering South Mountain comprises many coalescing, fan-shaped surfaces whose apexes are at the mouths of streams emerging from small drainage basins eroded into the north side of South Mountain. The morphology suggests that the apron is the result of an alluvial fan system which developed as material was eroded from South Mountain and deposited on the carbonate lowland adjacent to the mountain. The character of the diamicton suggests that the bulk of deposition occurred by debris-flow mechanisms with occasional intervals of fluvial transport and deposition. The deposits may have similarity to alluvial fans described elsewhere in the Appalachians by Kochel (1990).

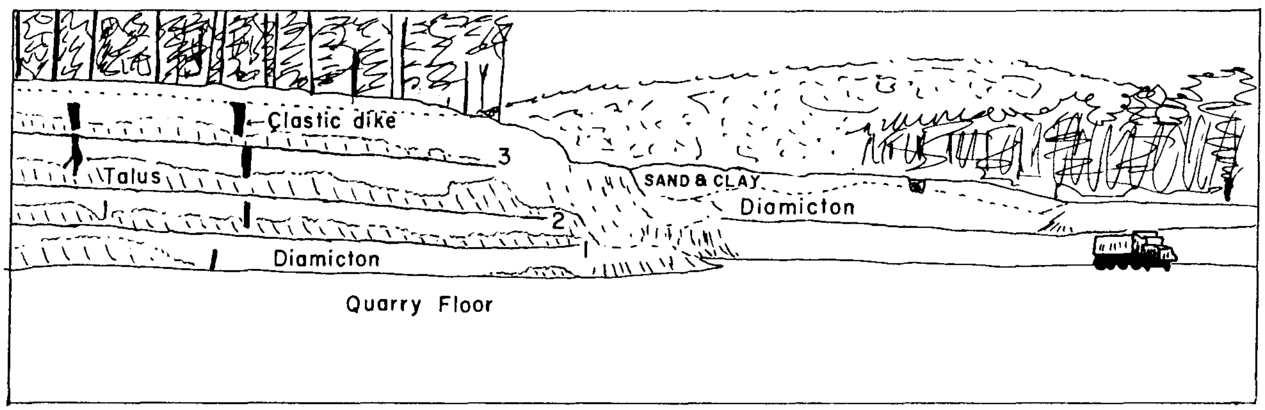


Figure 53. Sketch showing general aspect of Mainsville quarry in late June, 1991.

It is not known how much of the clay now present in the deposit was transported and deposited with the sand and clasts and how much of it was produced in situ by weathering of the clasts. Fauth (1968) reports that the Mont Alto Member contains (based on 10 thin sections) 1-13 percent "micromicas", presumably sericite, and also that the Harpers Formation contains interbedded phyllites. It seems probable that much of the clay was transported to the depositional site after being formed by weathering in the source area and that some of the clay formed in situ as the clasts were weathered after deposition. This is supported by the large amount of clay in the clay and sand deposit.

The clay itself is a problem because it is mainly smectite, a relatively transitional form, and not an end member of weathering which would be expected under current climate or some of the hypothesized past climates. It is possible that the smectite represents clay formed at the bottom of a very deep (and very old?) weathering zone developed under much more intense climatic conditions such as those which presumably occurred during the early part of the Cenozoic (see Clark, this guidebook). The smectite would then represent a weathering product which has not yet attained its most stable form. Another possibility is that the smectite formed during an interglacial period which was sufficiently intense climatically to do considerable weathering, but was not of long enough duration to produce kaolinite as an end member.

The diamicton deposit appears to have undergone several phases of deposition. The bulk of the deposit exposed to date is quite uniform in texture and appearance and would appear to have been deposited either during a single interval or during multiple intervals of sufficient similarity that discrete units have not been detected. There then may have been an period of erosion during which a wide channel was cut in the back part of the present quarry. This channel was subsequently filled by sand and clay. A later episode of diamict deposition occurred after the formation of the clastic dike.

The north flank of South Mountain is not an active fault and therefore one should reflect upon the diamicton apron and mountain relationship. The surface of the mountain has been going down as a result of erosion while the valley floor has been building up as a result of deposition. With a diamicton fill of 300 feet or more adjacent to the mountain it would seem that the valleys of the drainage basins should have been very deep at one time and should have filled with sediment as the valley floor rose to adjust its level to that of the accumulating diamict apron. We have no data on the depth to bedrock at the mouths of the drainage basins, but the topographic shape of the valleys suggests that the depth to bedrock is not great.

The above raises the question, what happened to compensate for the continued accumulation of diamicton without filling the valley. The obvious answer is that the valley floor was lowered by solution as the diamicton accumulated. Pierce (1965) speculated on the amount of solutional lowering in the Pond Bank area (near Stop 5) and suggested that the surface on the carbonates may have been lowered by as much as 1400 feet since the Late Cretaceous. Potter (1985) discussed probable solutional rates for the Cumberland Valley and concluded that a rate of 30 m/Ma (average) was reasonable. If that figure is applied to the Mainsville area, the carbonate surface underlying the diamict apron could be lowered by solution about 50 feet in 0.5 Ma or 1500 in 15 Ma (Middle Miocene). Solution of the underlying carbonate rock would thus seem a probable mechanism to account for the lack of discordance between the diamicton apron and the drainage basin valley. It does not resolve the problem of timing of deposition.

Clastic Dike

Although there is an initial tendency to call this clastic dike an ice wedge cast, a cursory examination of the literature (e.g., Black, 1976, 1983; Leffingwell, 1915; Mackay, 1974; Pewe and others, 1969; Taber, 1943; Washburn, 1980) indicates that this clastic dike is not an ice wedge cast. Black (1976, 1983) lists several criteria required for a feature to be a true ice wedge cast and this dike fails several items. In particular, the host adjacent to the dike lacks upturned pressure features.

It is apparent that the dike was filled from the top; therefore, a crack in the host must have developed and then been filled. The crack was almost certainly not a desiccation crack because they are not known to form in materials with the texture of the diamict (Wilden and Maber, 1961; Neal and others, 1968). Heron and others (1971) reported on clastic dikes in North and South Carolina which fill cracks formed either by slump or hillside creep, but this mechanism does not seem probable at the Mainsville site. It is possible that a crack formed as the result of subsurface collapse of carbonate material, but this does not seem a likely explanation. Yehle (1954) described soil tongues developed by differential solution in carbonate bearing gravels, but the lack of carbonate in the Mainsville diamict and the overall character of the clastic dike argue against this origin.

I suggest, for the sake of argument, that the crack was the result of frost cracking during a period of glacial maximum when the diamicton was frozen as part of continuous or discontinuous permafrost. The diamicton is an ideal material to maintain a vertical face even when unfrozen, as evidenced by mined faces in the quarry, and would be even better when frozen. The initial crack was probably not more than a few meters deep. During summer thaw clay was washed into the open crack along with some sand and occasional clasts which fell from the upper margins of the crack. During the winter the smectite froze and expanded thus widening and deepening the crack. Frost cracking occurred again and more material was subsequently added to the deeper crack. The process was repeated many times until the crack and its filling reached its present width and depth. This hypothesis is not totally satisfactory. Although frost cracks are known to reoccur at the same place many times and are often repeatedly filled or partly filled with wind-blown sand and silt, filling with water-transported sediment is either uncommon or unknown because it is not cited in the literature. Water-transported sediment is also a problem if the climate was dry as suggested earlier (p. 67). It is also not known whether or not the frost crack, once initiated, can be propagated downward by the pressure of freezing clays. Nor is there certainty about the expansive character of the clays when frozen. Other realistic hypotheses are welcome.

After the glacial maximum was past and the ground became unfrozen, the smectite was subjected to seasonal wetting and drying from the vertical margins of the dike. This caused the clay to orient parallel with the crack margins and to develop laminations. This process is rapid as indicated by the development at a bench surface of horizontal laminations from clay which had vertical laminations prior to cutting of the bench. The reorientation of laminae took place within one year.

Subsequent to the formation of the clastic dike, the area was subjected to an intense interglacial and a thick and well-structured soil was developed on the dike material. This soil was subsequently truncated by erosion and another layer of diamicton was deposited. Another soil was developed on the uppermost diamicton. The uppermost soil has development which could be expected for soil formed since the last glaciation (about 20,000 years). That makes the age of the paleosol problematic. It could have devloped during the post-Illinoian interglacial, the Sangamon, or during a pre-Illinoian interglacial. The paleosol is similar to paleosol development of presumed pre-Illinoian age elsewhere in Pennsylvania and is probably associated with those soils even though their age is also unkown. Braun (1988, p. 21-23) argues that the pre-Illinoian glaciation in Pennsylvania could have occurred at either 440 Ka or 640 Ka.

It appears reasonable that the main body of diamicton at the Mainsville site was deposited prior to the Illinoian glacial maximum. Just when is a good question. Because the clasts were hard and unweathered when transported and deposited, they could not have been derived from saprolitized rock. Rather it would appear that the clasts were prepared for transport by physical weathering and that the chemical weathering which produced the clays lacked either sufficient intensity or time, or both, to produce an end member clay, kaolinite. Thus, the clasts and clay could have been prepared during climatic fluctuations sometime in the early Pleistocene and subsequently transported and deposited.

On the other hand, the clasts and clay might represent the bottom of a deep weathering zone which had not yet saprolitized all of the rock or altered the clays to kaolinite. In this scenario the weathering could be a middle Cenozoic event and the erosion and deposition an event as old as Middle Miocene. There is no soil evidence to support this hypothesis, although such could have been destroyed by surface stripping during the Pleistocene. The deeply weathered clasts add credibility to the antiquity of the diamicton, but the thickness of the deposit does not compare favorably with the probable amount of carbonate surface lowering since the Middle Miocene. Good hypotheses are welcome.

HYDROGEOLOGY

The deposits and physical setting at the Mainsville Quarry can provide some insight into the hydrogeologic framework of the Cumberland Valley. The exposed deposits consist primarily of poorly sorted material ranging in size from fine sands to large, weathered cobbles. In areas proximal to the source material (quartzite, phyllite, and schist that form South Mountain), the alluvial and colluvial deposits can be several hundreds of feet thick. Precipitation and streams that flow over this material are able to infiltrate water quickly and easily into the porous, unconsolidated material. The water then infiltrates downward to recharge the ground-water system. Within the unconsolidated aquifer, ground-water flow is downward to the underlying carbonate rocks and laterally to streams and springs. Once in the carbonate rock, the chemically aggressive water is able to weather and dissolve calcitic material from the bedrock creating enlarged openings and cavities. Some of the cavities are large enough to cause solution sinkholes to be reflected on the land surface. The weathered carbonate rock and the overlaying alluvium and colluvium can be as thick as 450 feet. The saturated part of this regolith, up to 240 feet thick, represents a large reservoir of stored water. Through the summer months and during dry periods, this stored water can be slowly and steadily released to the carbonate-rock aquifer, and to streams and springs lower in the valley.

		LEAVE STOP 6 AND RETURN TO MAINSVILLE THE WAY CAME.
1.4	111.9	**STOP SIGN. TURN RIGHT** onto Mainsville Road.
0.7	112.6	Cumberland County line. Note gravel ahead.
1.0	113.6	Hill on left is composed of limestone which continues to the hilltop. There is an outcrop at 113.9.
0.5	114.1	Route is now on gravel again.
0.2	114.3	**STOP SIGN. PROCEED STRAIGHT AHEAD** onto Cleversburg Road.
0.7	115.0	**TURN LEFT** onto Neil Road just after a sign on the right for the Village of Cleversburg.
1.1	116.1	**STOP SIGN. PROCEED STRAIGHT AHEAD** on Hershey Road.
0.8	116.9	**STOP SIGN. TURN LEFT** onto Walnut Bottom Road and immediately **TURN RIGHT** to Interstate 81 N (almost straight across).
4.5	121.4	Floodplain of Yellow Breeches Creek on right to the southeast.
0.7	122.1	Leave floodplain of Yellow Breeches Creek and ascend onto topography with greater relief.
6.0	128.1	Leave belt of more rugged topography.
3.2	131.3	Cross Alexander's Spring Creek. On right is a dry valley that extends for 1.25 miles to the southwest.
5.8	137.1	Interchange 16 which leads to PA Route 74 S and the first stop of second day.
3.0	140.1	**EXIT** Interstate 81 at Exit 17 E to US Route 11 New Kingston.
0.3	140.4	**YIELD SIGN. BEAR RIGHT** onto US Route 11 N.
0.3	140.7	**TURN RIGHT** into The Embers Inn.
		END OF DAY 1 FIELD TRIP. SEE YOU AT THE BANQUET!

ROAD LOG AND STOP DESCRIPTIONS - DAY 2

Mileage Inc	Cum	
0.0	0.0	**LEAVE** parking lot of The Embers Inn. **TURN LEFT** onto US Route 11.
0.4	0.4	**TURN RIGHT** onto Interstate 81 South.
3.3	3.7	**EXIT RIGHT** at Exit 16, PA Route 641, High St.
0.1	3.8	**STOP SIGN. TURN RIGHT** onto PA Route 641. **GET IN LEFT LANE.**
0.1	3.9	**TURN LEFT** onto Fairfield Street.
0.5	4.4	**STOP SIGN. TURN LEFT** onto PA Route 74 S.
0.2	4.6	**TURN RIGHT** onto Forge Road. Note that between here and Boiling Springs the route does not cross a single stream. Yet relief is low and mapped sinkholes are rare.
3.5	8.1	**STOP SIGN. TURN LEFT** onto First Street at the Boiling Springs Tavern on the left. **TURN LEFT IMMEDIATELY** into the parking lot behind the tavern.

STOP 7. HYDROGEOLOGY AND THE SOURCE OF BOILING SPRINGS, SPRINGS, CUMBERLAND COUNTY, PENNSYLVANIA

Discussant: Albert E. Becher

Boiling Springs is one of the three largest springs in the Cumberland Valley and ranks as the seventh largest in the state. Average daily discharge is about 16.5 million gallons based on six instantaneous measurements made between 1944 and 1971 by the U. S. Geological Survey (Flippo, 1974). The water is used solely for recreational purposes in Boiling Springs Lake and the world-renowned trout fishing waters of Yellow Breeches Creek. Until recently, the sequence of carbonate rock formations that floor the southern part of the Cumberland Valley was considered to be the source of water for the springs.

Flow from Boiling Springs is through openings in strongly folded limestone of the Elbrook Formation about 1 or 2 miles north of South Mountain (SM) (Figure 54). The Elbrook Formation is composed of interbedded calcareous shale, argillaceous limestone, and medium-bedded to massive limestone. Rocks typical of the formation are well-exposed north of Boiling Springs Tavern. A diabase dike of Triassic age extends across the valley from the north and just north of Boiling Springs, splits into two branches that enclose the spring area. To the south, on the flank of SM, alluvial, colluvial, terrace, and residual deposits overlie the carbonate rocks to depths up to several hundred feet. Deposits are thickest near the contact between quartzites of SM and the overlying carbonate rocks.

There are two major areas of ground-water discharge from Boiling Springs; one in the walled basin north of the tavern; the other near the southwestern shore in the northwest corner of the lake (Figure 54). In both areas the discharges can be seen as boils that rise several inches above the water surface. Head differences between water in the openings and in the basin may be

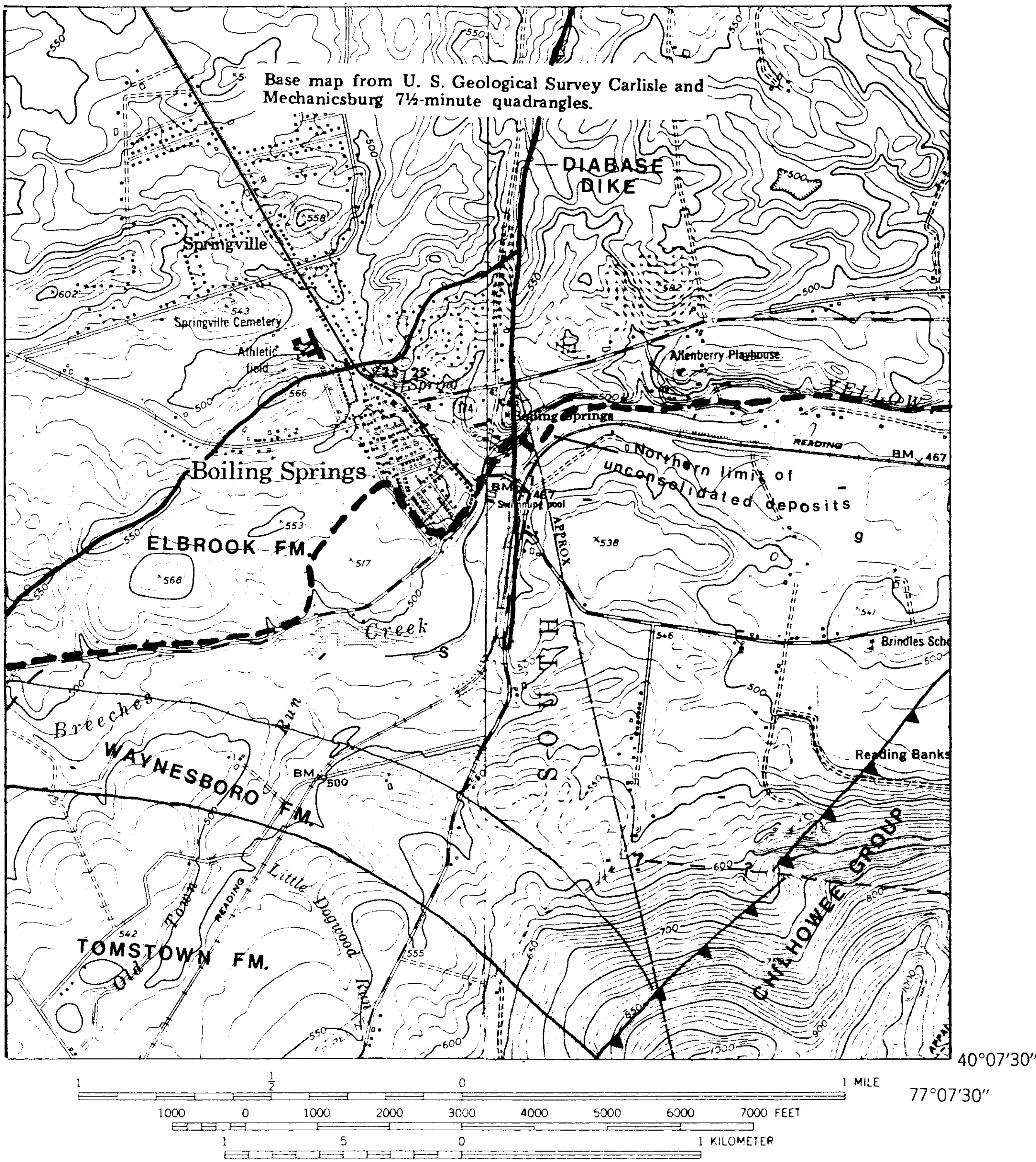


Figure 54. Map of Boiling Springs area.

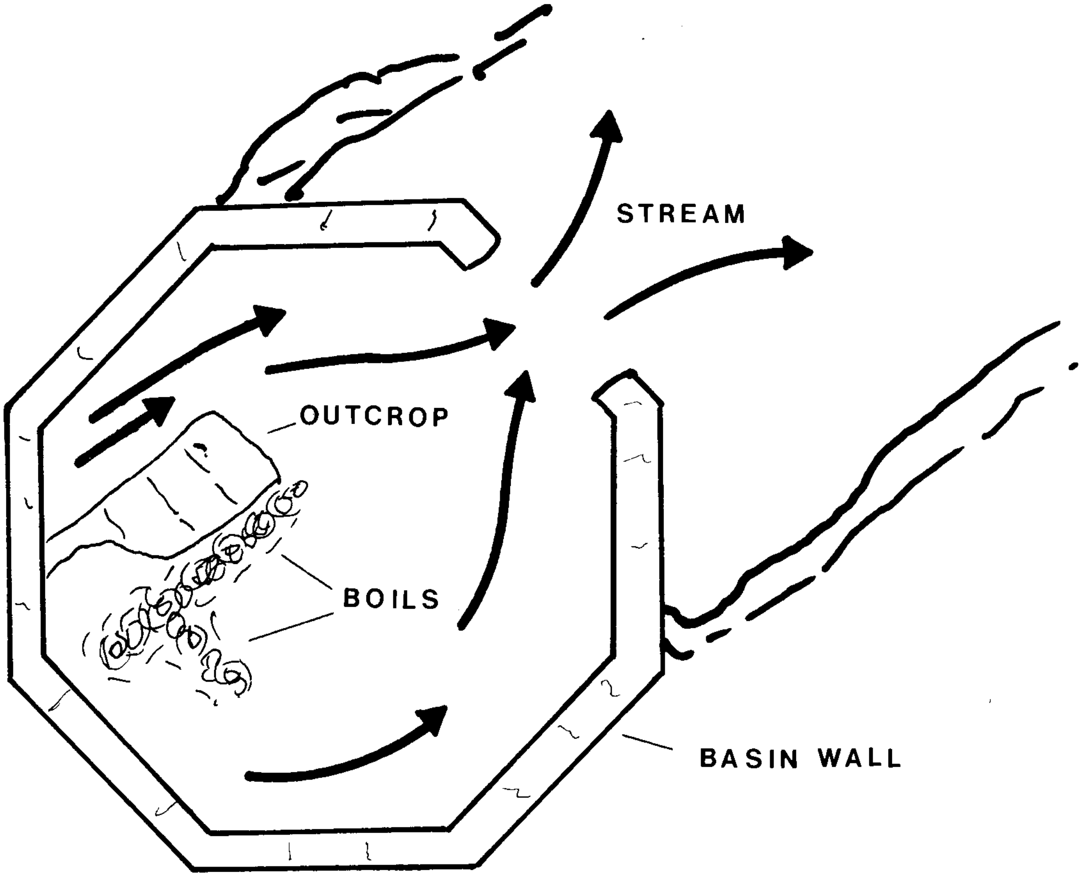


Figure 55. Diagram of walled spring basin behind Boiling Springs Tavern.

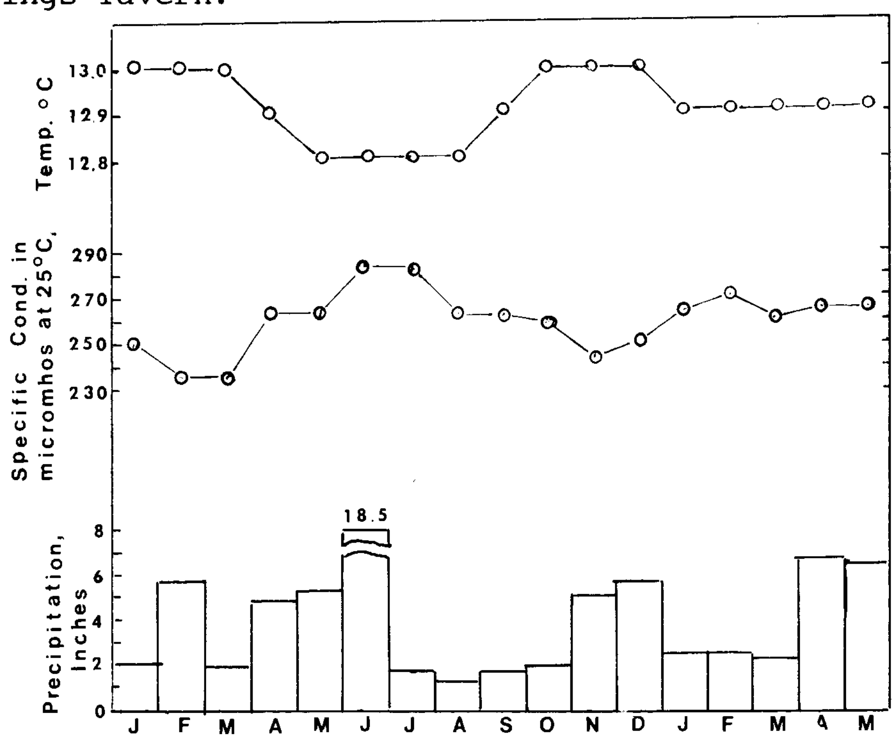


Figure 56. Monthly temperature and specific conductance of water from Boiling Springs and precipitation at Carlisle.

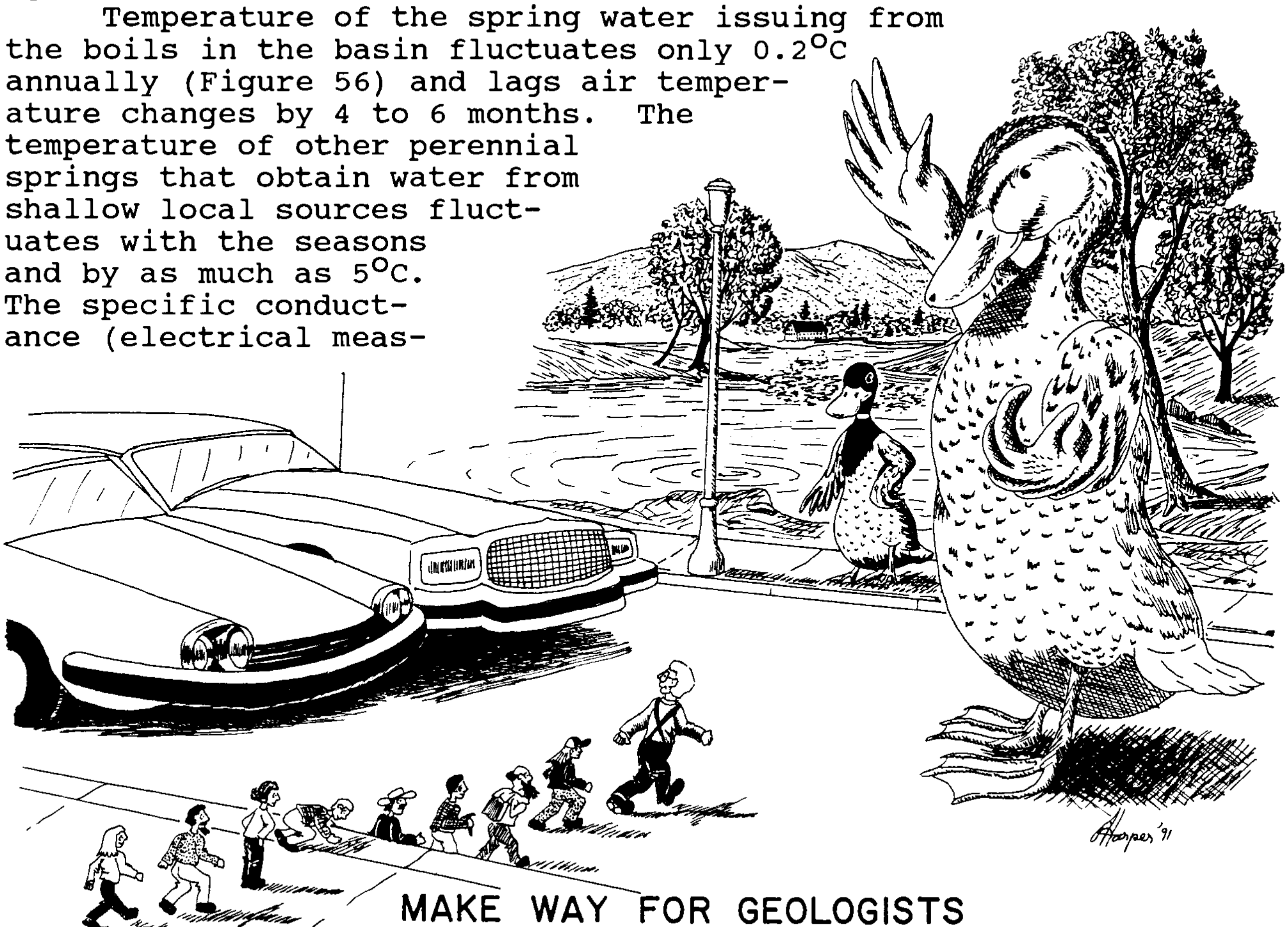
several tens of feet. The water level in the basin is 10 to 15 feet above Yellow Breeches Creek.

Characteristics of the spring discharge openings can be seen best in the basin (Figure 55). Here two intersecting linear zones of discharge that parallel local joint directions are visible. Additional flows may be present under the north wall.

Three alternative sources of water were considered for Boiling Springs. The magnitude and degree of fluctuation of the flow, water quality and its seasonal variability, and geologic factors, however, support SM as the major source. Many stream channels on the flank of SM lose water to the colluvium and some often are dry before reaching Yellow Breeches Creek. It is probable that precipitation and runoff from SM infiltrates the colluvium and moves downward through permeable zones into solutionally-enlarged openings in the carbonate bedrock, then under Yellow Breeches Creek and finally discharges, under pressure, through openings at the narrow end of the funnel-like area created by the branching diabase dike.

A drainage area of more than 20 mi^2 is needed to collect the amount of water discharged by Boiling Springs based on the average basin-wide groundwater discharge of 0.81 million gallons per day per square mile. From the ground-water divide, separating drainage to Conodoguinet and Yellow Breeches Creeks, south to the springs, the drainage area is only about 3 mi^2. Ample drainage area exists only on the flank of SM.

Temperature of the spring water issuing from the boils in the basin fluctuates only 0.2°C annually (Figure 56) and lags air temperature changes by 4 to 6 months. The temperature of other perennial springs that obtain water from shallow local sources fluctuates with the seasons and by as much as 5°C. The specific conductance (electrical meas-



MAKE WAY FOR GEOLOGISTS

With apologies to Robert McCloskey

ure of dissolved ionic species) of water from Boiling Springs is about half that of ground water in nearby wells but is like water from wells in carbonate rock on the flank of SM. Wells on SM are completed as open casings that penetrate only a few feet into bedrock below the consolidated overburden. The lag in spring water temperature fluctuation suggests a longer residence time; the slight fluctuation suggests deep, well-mixed water; but the lower specific conductance means lesser contact with carbonate rock. A SM source fits these interpretations.

Spring and well data suggest that the movement of water under Yellow Breeches Creek occurs along the entire flank of SM. Much of this water is then discharged into the Yellow Breeches but some moves under the apparent ground-water divide into the Conodoguinet Creek drainage basin (Becher and Root, 1981).

		LEAVE STOP 7. TURN RIGHT onto First Street (PA Route 174 W).
0.2	8.3	**TURN LEFT** onto Walnut Street to Mt. Holly Springs.
1.6	9.9	Cross Yellow Breeches Creek. Park on the left illustrates an environmentally sound use of a floodplain. The low ridge ahead which is crossed at 10.5 miles is a branch of the diabase dike discussed at Stop 7.
1.3	11.2	PPG Industries on the left is a float-glass plant which extrudes molten glass onto a bed of molten tin, which then cools in two ribbons continuous for the half-mile length of the plant. Test borings made when the plant was built show depths to bedrock typically 50-75 feet in the area. One boring did not reach bedrock at the 230-foot completion depth. Well logs indicate gravels overlying silty clays interpreted as residuum.
0.1	11.3	Well tower adjacent to Pump House Road on the right is part of the South Middletown Township water supply. Depth to bedrock when drilled was 107 feet. During the pumping test, a sink hole opened in the adjacent field which was 20 feet in diameter and 20 feet deep. It exposed gravels in the walls. There may be some large cavities below. Note the karst topography in the area.
1.4	12.7	**STOP SIGN. TURN LEFT** onto PA Route 34 S in Mt. Holly Springs borough.
0.4	13.1	Amilia Given Library on the left is made of Triassic sandstone and is a classic example of Richardson Romanesque architecture.
0.4	13.5	Four feet of water covered the road here during the Agnes flood of 1972. On August 7, 1892 Walcott collected *Olenellus* fragments from Antietam sandstones exposed along the railroad across the creek on the left (see p. 29-30). Deer Lodge on right.
0.2	13.7	**TURN RIGHT** onto Ridge Road, a gravel road after Deer Lodge parking lot and before the bridge.
0.3	14.0	**SHARP RIGHT TURN** up hill. Buses need wide swing. Do not go straight ahead.
1.3	15.3	Scenic view on right (ignore dump).

The Agnes Flood in Mount Holly Gap, looking South. Mountain Creek is on the left (and in the road!). June 22, 1972, at about 5 pm.

1.7	17.0	**ROAD FORK. BEAR RIGHT.**
0.6	17.6	Turnout and scenic view.
1.4	19.0	On left are float blocks of Weverton Formation derived from Hammond's Rocks.
0.2	19.2	**PARK** in parking area on left.

STOP 8. HAMMOND'S ROCKS

Work by: Noel Potter, Jr., Henry Hanson, Rolf Ackermann, Gretchen Dockter, Steven Lev, Laura Pezzoli, and Thomas Troy
Discussants: Noel Potter and Henry Hanson

INTRODUCTION

Hammond's Rocks is a 100 by 400-foot outcrop of Lower Cambrian Weverton Quartzite along Ridge Road on the crest of South Mountain 4.5 miles southwest of Mount Holly Springs, Pennsylvania (Figure 57). It displays a variety of sedimentary, structural, and geomorphic features unrivaled nearby in the Weverton for close study.

The outcrop is composed of interbedded quartz sandstone and conglomerate. Approximately 1/3 of the exposed 150-foot section

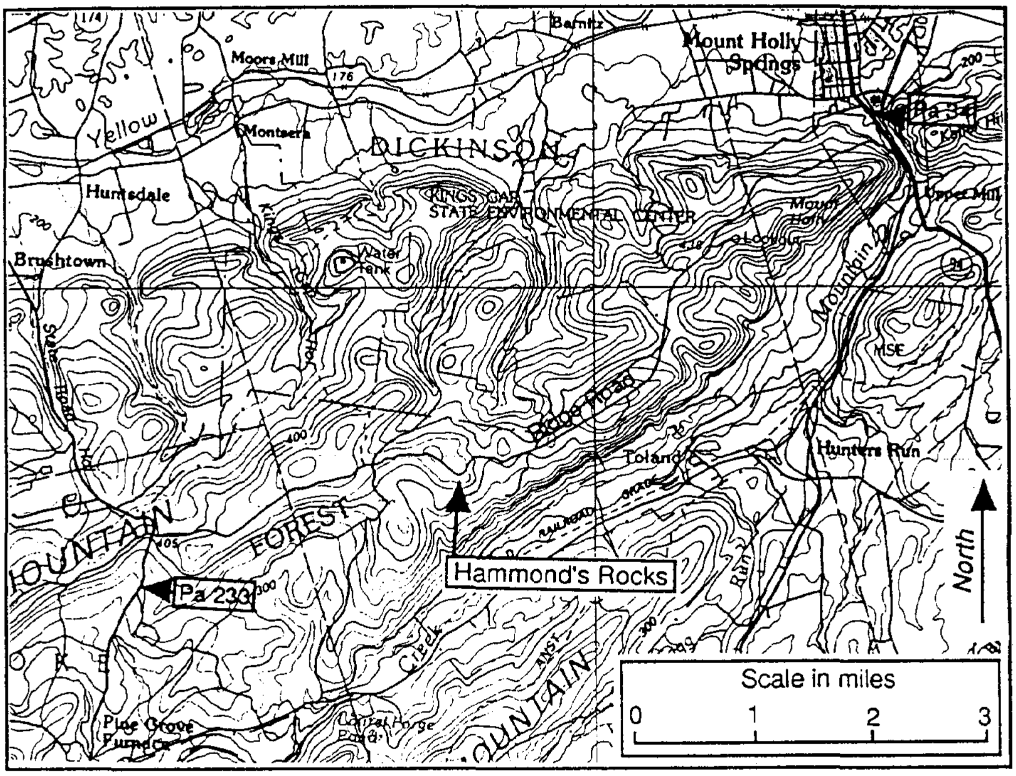


Figure 57. Location of Hammond's Rocks. From USGS Carlisle, PA. 1:100,000 sheet.

is composed of pebbly beds on the order of 2-3 feet thick. The pebbles have been deformed into elliptical markers that can be used as measures of strain. In general pebbly beds stand out in relief, whereas some sandy beds tend to weather more rapidly. Some of the sandy beds are cross-bedded. For a discussion of Weverton stratigraphy, see Key (this guidebook).

Good outcrops of the Weverton are not common, and this exposure has the potential to yield considerable information. To help locate features in a confusing jumble of boulders, two of us (Hanson and Pezzoli) mapped the topography of the outcrop on a grid with 1-meter spacing for over 7000 points and prepared a computer topographic map from that data (Figure 58). The version of the map used here is unedited, so vertical slopes are spread over 1 meter, and the bias of gathering the data has produced a somewhat rectangular pattern to the contours.

Hammond's Rocks lies within the Mount Holly Springs quadrangle mapped by Freedman (1967). The crest of his "Hammonds Rocks Anticline" trends about N50°E and passes just north of the outcrop and Ridge Road. Deformation of the rocks follows the structural plan of South Mountain described by Cloos (1947; 1971), Fauth (1968), Root (1970), and Root and Smith (this guidebook). Hammond's Rocks consists essentially of the south limb and trough of a slightly overturned syncline with its axial surface dipping southeast (Figure 59). Most of the eastern 2/3 of the outcrop consists of near vertical or slightly overturned beds with strikes about N60-75°E. The very northern edge of the

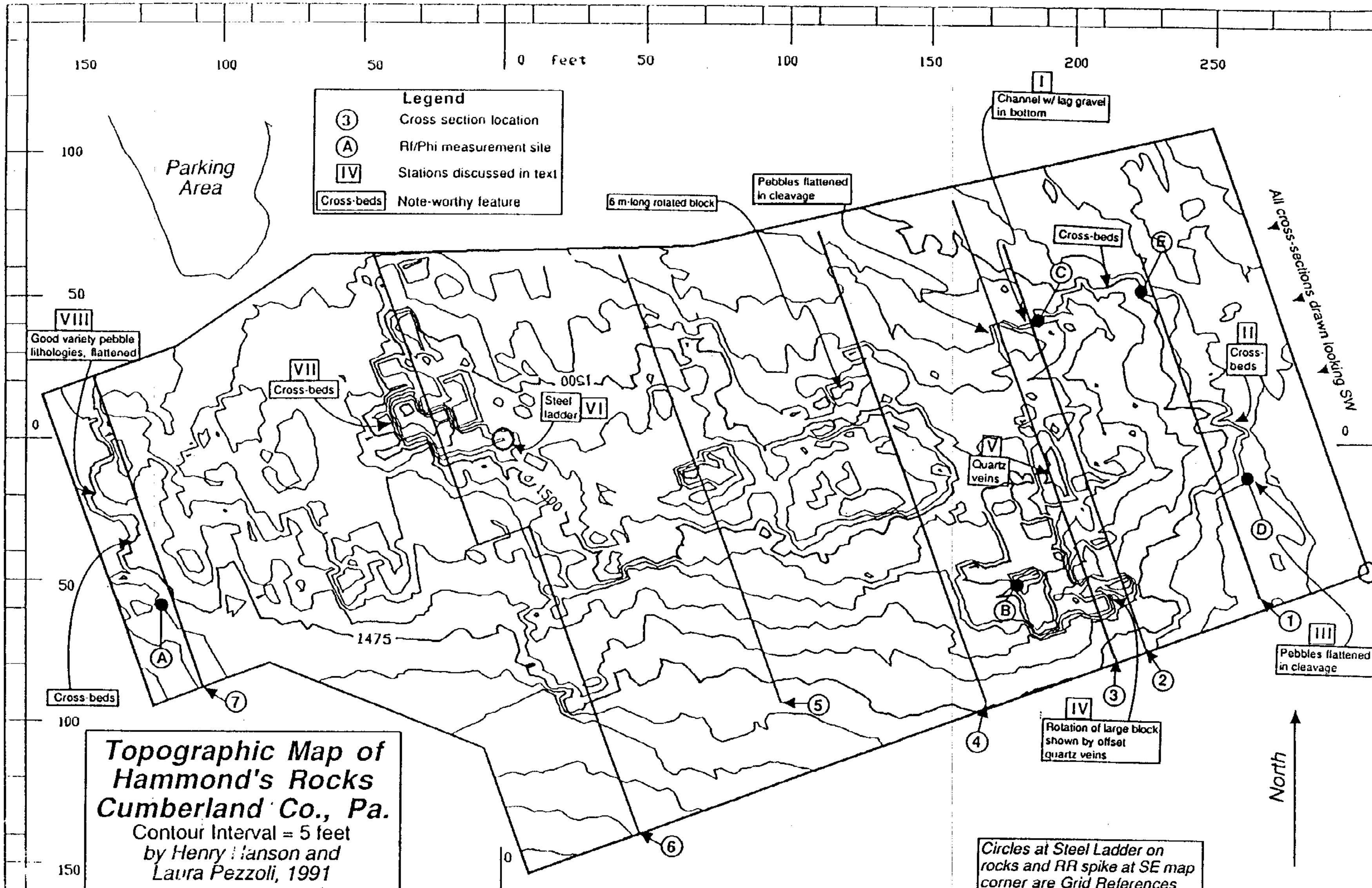


Figure 58. Topographic map of Hammond's Rocks showing locations of features discussed in text.

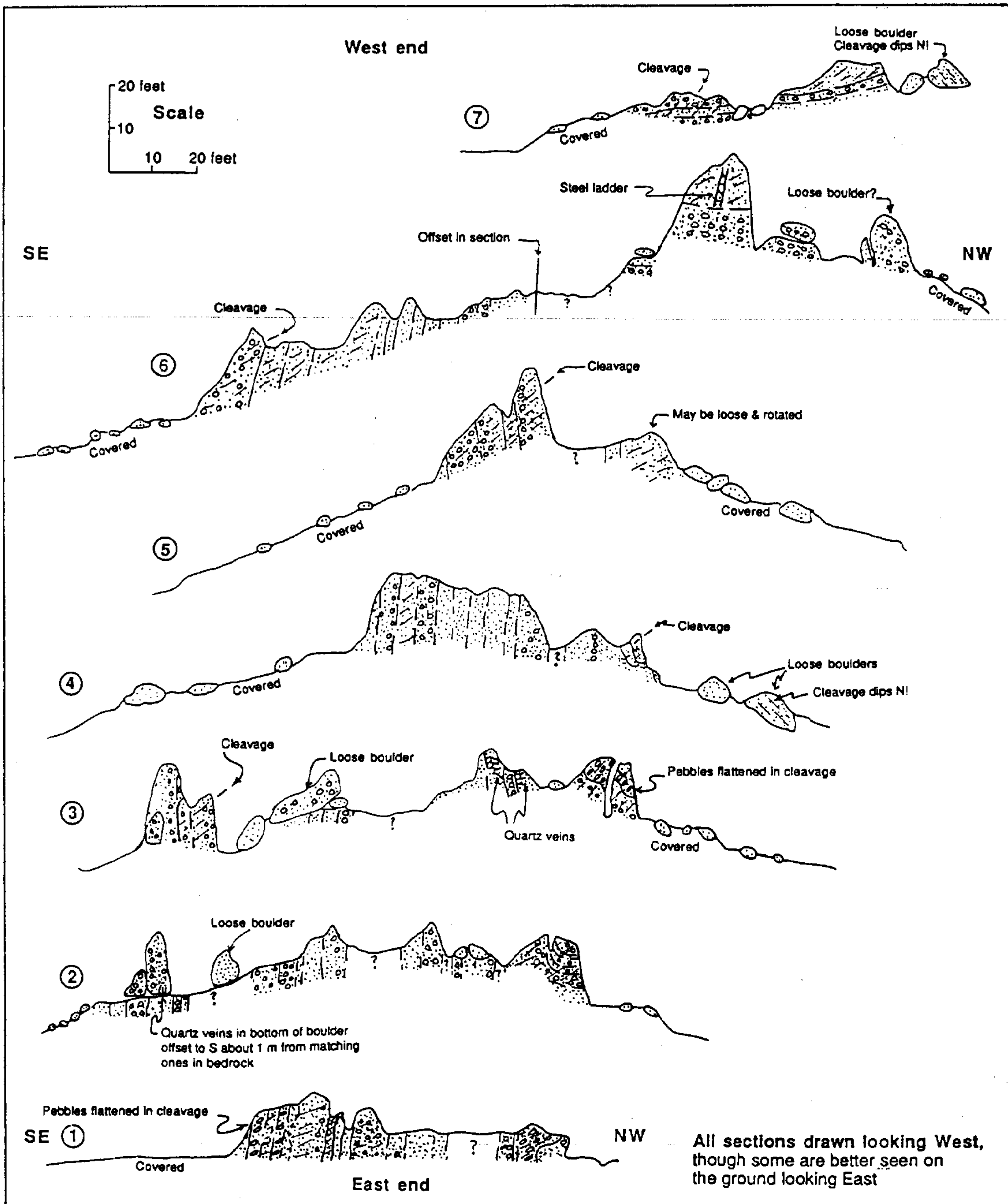


Figure 59. Cross sections of Hammond's Rocks. All sections are drawn looking WEST for ease of comparison, despite the fact that some are better seen at the outcrop looking in the opposite direction. Locations of sections are shown on Figure 58.

eastern part of the outcrop and most of the western part of the outcrop lie near the trough of the syncline, where dips are gentle, often with northwest strikes and dips of 10-20° toward the east. An equal-area net of poles to beds shows a girdle with concentrations of poles expected from the syncline (Figure 60A). The pole to the plane defined by that girdle should approximate the hinge of the syncline--it trends approximately N65°E and plunges 10-20°NE.

Cleavage consistently strikes about N45°E and dips about 45°SE (Figure 60B), approximating both the axial surface of the syncline at Hammond's Rocks and of Alleghanian folds throughout South Mountain and the Great Valley.

We determined bedding/cleavage intersections on sterographic nets for each of the locations where we measured both. These intersections should approximate the fold axis for Hammond's Rocks. The equal-area net of bedding/cleavage intersections (Figure 60C) shows maximum concentration at about N65°E, 20°NE, which is quite consistent with that derived from the girdle of poles to beds. Most intersections plunge northeast, but a few plunge gently southwest.

Because we were interested in the strain recorded in the deformed pebbles in the conglomeratic units, we searched for pebbles in the outcrop that were sufficiently exposed by weathering for us to see, and to be certain of the orientation of all 3 axes. We placed a wooden dowel parallel to the long axes of each pebble, and measured its bearing and plunge. The orientations of the long axes of 44 of these pebbles were plotted on an equal-area net (Figure 60D). The pebble long axes form a broad girdle across the southeast quadrant of the net, approximating the plane of cleavage, but with considerable scatter about that plane.

We had naively expected the pebbles to be oriented similar to deformed oolites in the Great Valley in the classic studies of Cloos (1947, 1971). Cloos' oolites typically have their longest axes in the plane of cleavage and perpendicular to fold axes, their intermediate axes parallel to fold axes (bedding/cleavage intersection), and their shortest axes perpendicular to cleavage (Figure 61A). The long axes of the pebbles in Hammond's Rocks do not follow this simple plan (Figure 61B). Why the difference, when both the younger carbonates enclosing the oolites and the sandstone enclosing the pebbles at Hammond's Rocks have been subject to the same deformation? The answer lies in a simple but not immediately obvious fact. Whereas it is reasonable to assume that most oolites were nearly spherical before deformation, and thus were deformed the same amount in a particular direction at a given location, the pebbles can not be assumed to have originally been spherical, and therefore their pre-deformation shape has influenced their post-deformation shape. Indeed, a careful examination of pebbles in a modern gravel deposit in most cases will show that most are ellipsoidal. We shall argue the details of what happened to the pebbles in our discussion for Station III. For the moment, suffice it to say that we believe that the initially ellipsoidal pebbles had random orientations, and were deformed mainly by being squashed perpendicular to and rotated

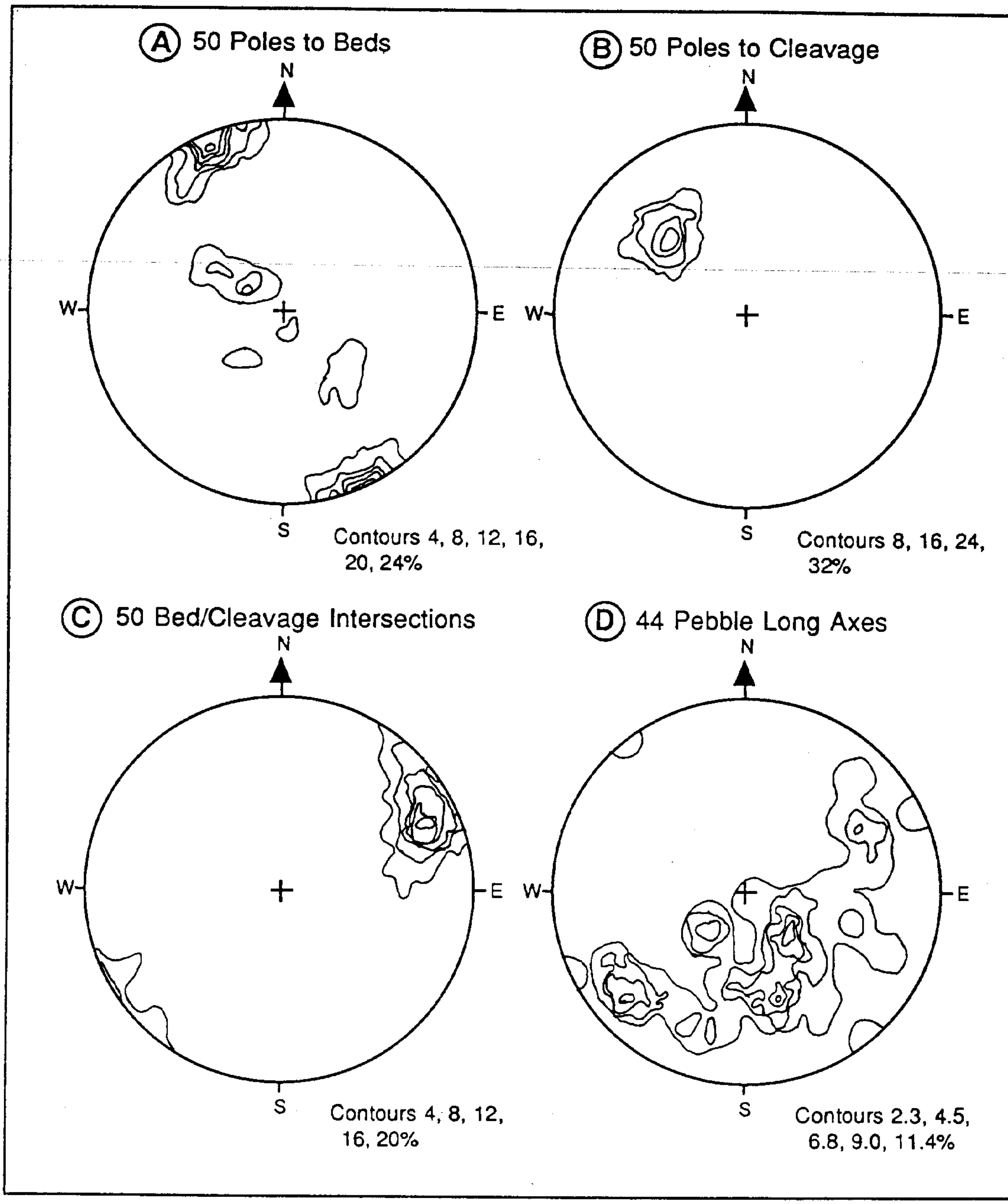


Figure 60. Equal-area nets showing orientation of (A) bedding, (B) cleavage, (C) bedding/cleavage intersections, and (D) pebble long axes at Hammond's Rocks.

toward cleavage to give them their observed orientation (Figure 61B).

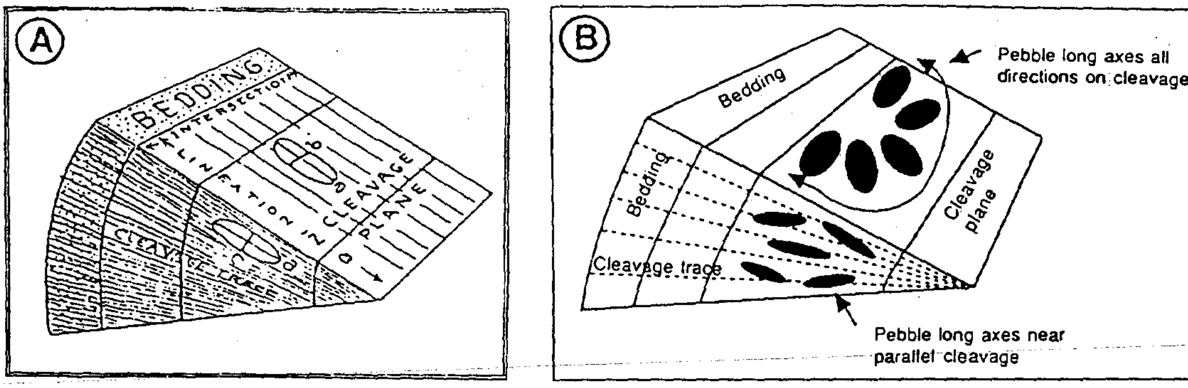


Figure 61. Comparison of orientation of deformed oolites in Cambro-Ordovician carbonates of the Great Valley and of deformed pebbles at Hammond's Rocks. (A) Deformed oolites from Cloos (1947, p. 895), (B) Deformed pebbles at Hammond's Rocks.

We measured the orientation of 81 joints in Hammond's Rocks. These were plotted on an equal-area net, not presented here. The only preferred orientation gleaned from the data is that most joints are near vertical with a 75-90^{o} dip. Strikes span the compass. There were perhaps slightly greater concentrations of near-vertical joints trending about N40-50^{o}W and N60-80^{o}E. These directions may be reflected in the main trends of large faces in the outcrop. Fauth (1968, p. 59) found the majority of joints he measured in the Caledonia Park area to be near-vertical.

Clark (this guidebook) has suggested that Hammond's Rocks is a tor--a larger version of the one seen at Stop 2--formed under periglacial conditions during the cold phases of the Pleistocene. The outcrop projects 20-30 feet above a surface that slopes gently away from it. Many of the blocks on the periphery of the outcrop are loose and can be demonstrated by anomalous orientations of structural features to have rotated or slid away from the outcrop. Several of these blocks are shown on the cross-sections (Figure 59). Numerous 1-2 meter-long blocks mantle the surfaces gently sloping away from the outcrop. Many of the sharp inflections in contours on the topographic map (Figure 58) are caused by these large blocks.

There is no indication of substantial movement of these boulders today, so we infer, but can not prove, that these blocks were moved under the more severe conditions of a periglacial climate. The gently sloping surfaces resemble cryoplanation terraces that are common in periglacial regions such as Alaska today. Clark (this guidebook) points out the discordance of topography and structure in Hammond's Rocks and nearby outcrops to the north and east mapped by Freedman (1967, Figure 21 and Plate 1), and suggests that the gentle surfaces around Hammond's Rocks are surfaces of reduction formed by blocks moving away from the tor during cold climate.

THE OUTCROP

It is convenient to discuss many of the details of Hammond's Rocks at stations (identified by Roman numerals on Figure 58) that you can visit at the outcrop. These stations are arranged in a clockwise direction, beginning near the northeast corner of the outcrop. Begin by turning left from the parking area through the woods along the north (near) side of the outcrop to near its east end.

As you walk toward Station I, note that there are numerous large boulders north of the outcrop. There is no evidence that they are moving substantially today. We suggest that periglacial processes moved them during the colder parts of the Pleistocene.

Station I

We begin by examining some aspects of the sedimentology of the Weverton. The long 10-12-foot-high outcrop is best viewed from the northwest. Although you are looking perpendicular to strike, the variety of lithologies represented in Hammond's Rocks is displayed here. Several crossbedded units are visible at the lower-left end of the face. Some of these can be seen in cross section at the right (west) end of the face. In the upper-middle section of the face is a prominent channel (Figure 62). The channel floor is scoured into a conglomerate bed and the channel is filled with crossbedded sandstone. A prominent quartz pebble that protrudes into the channel is 12 x 5 cm.

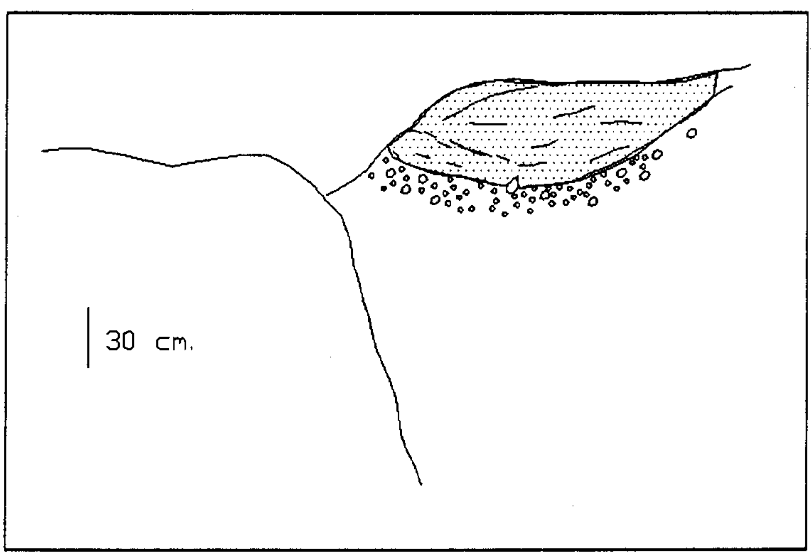


Figure 62. Cross section of channel with lag gravel at Station I.

What was the nature of the current that created the channel? The critical shear velocity required to move such a pebble in traction can be estimated from Shield's diagram (Blatt, et. al., 1980, p. 102) and is on the order of 20 cm/sec. That velocity would put into suspension quartz grains with a diameter up to 2 mm. Shear velocity is equal to the square root of gravity x depth x slope. A stream flowing over its own sediment has a slope on the order of 0.001. Were a stream to produce such a shear velocity, it would be about 4 meters deep. The channel that the cobble protrudes into is much smaller, suggesting that the channel is either a scour mark in deeper water or that much of the channel was removed by subsequent erosion.

A cross-sectional view of the channel can be seen by climbing into the notch next to the channel. Access is easiest from the east end of the outcrop and above. Although beds are nearly horizontal in the face, they almost immediately turn vertical into the south limb of a syncline just to the south of the face (see Figure 59, Cross section 2).

Between Stations II and III

The east end of the outcrop is composed almost entirely of near-vertical to slightly overturned beds with tops toward the north (Figure 59, Cross section 1). A walk along this end gives a sense of the proportions of pebbly sandstone, conglomerate, and crossbedded sandstone, particularly between Stations II and III. It is our impression that in the oldest (left) bed of pebbly sandstone, the size of the pebbles in the lower part of the bed are larger, but that there are more pebbles in the upper part. Just below the first sandstone bed, the pebbly sandstone grades into a conglomerate. The same sequence occurs higher in the section where there are small troughs filled with crossbedded sandstone and several conglomeratic beds. One crossbedded, sand-filled trough (Figure 63, at Station II) can be seen in both plan view and in cross section. The cross-beds are arcuate in plan view and outlined by dark heavy minerals.

The sedimentological history of these beds appears to consist of rapid deposition of poorly-sorted, pebbly sand, followed by armoring (Allen, 1982, p. 158) of the surface by winnowing of sand, cutting of channels, and transport and deposition of the sand to fill channels and other topographic lows.

Station III

You are at the left end of Cross Section 1 (Figure 59). We have chosen this locality to discuss the deformed pebbles in Hammond's Rocks for ease of access, but the method can be applied in any of the better exposures of conglomerate in the outcrop.

An Overview

Examine the southernmost exposure of conglomerate, looking southwest parallel to bedding and cleavage. Bedding dips about 80°SE and cleavage dips about 53°SE. Bedding is steeper than cleavage, and both dip SE, indicating that the beds are overturned with tops toward the north. The pebbles in the outcrop have a distinct fabric with most of their longest axes (at least as seen in this cross-section) nearly parallel to cleavage, and their shortest axes nearly perpendicular to cleavage. It is important to note, however, that pebbles can be found with their

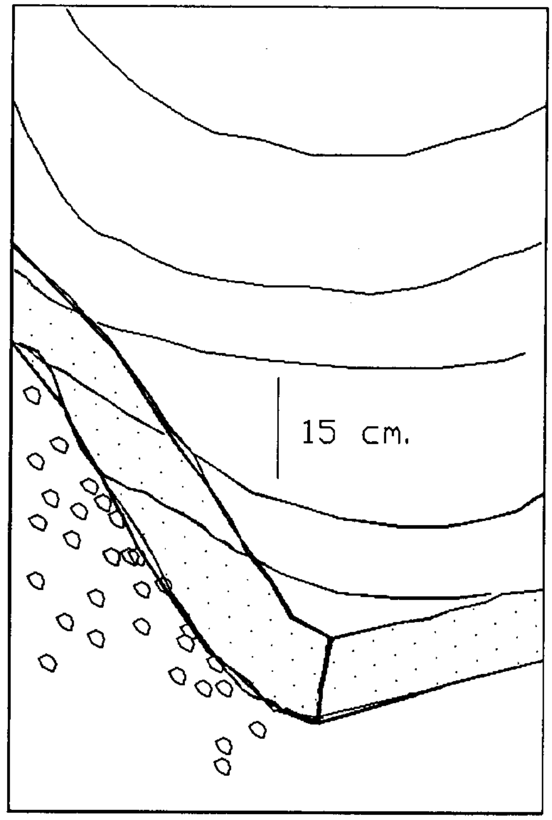


Figure 63. Arcuate cross-beds in plan and section at Station II.

long axes at high angles to cleavage. We have shown that the longest axes of pebbles lie in or near the plane of cleavage, but have a wide range of directions of plunge from near horizontal to directly down dip (Figure 60D).

Pebble Shape

It is difficult to measure all 3 axes of many of the pebbles in the outcrop because few are sufficiently exposed. However, after reading of a solution to the problem for a similar situation in the Piedmont of South Carolina (Hatcher, 1990, p. 94-97), we discovered that we could sample a large number of pebbles by collecting them from the surface where they had weathered from the outcrop. We collected and measured 3 suites of pebbles from: (1) Site HRW on the western half of Hammond's Rocks, (2) Site HRE on the eastern half of Hammond's Rocks, and (3) HRC from another outcrop of conglomerate with deformed pebbles, informally known as the "Chinese Wall," about 1000 feet north of the Hafmmond's Rocks parking area.

We selected for measurement only pebbles that were not broken and were clean enough of matrix to determine all 3 axes. Any excess matrix was gently removed in the lab. We measured long (X), intermediate (Y), and short (Z) axes to the nearest millimeter for each pebble (Table 16), and calculated the ratios of the axes: $a = X/Y$ and $b = Y/Z$. From the two ratios, k, the

Table 16. Shape of pebbles weathered from Weverton Formation at Hammond's Rocks. Size of pebbles in cm. SD = Standard Deviation. Ave = Average. Int = Intermediate.

Site	Number of Pebbles	X=Long Axis Ave (SD)	Y=Int Axis Ave (SD)	Z=Short Axis Ave (SD)	a=X/Y Ave (SD)	b=Y/Z Ave (SD)	k(see text) Ave (SD)
HRW	170	3.9(0.8)	2.8(0.6)	1.6(0.4)	1.4(0.2)	1.8(0.4)	0.9
HRE	69	2.8(1.0)	1.9(0.6)	1.0(0.3)	1.5(0.3)	2.0(0.6)	0.8
HRC	134	2.6(0.9)	1.9(0.7)	1.2(0.5)	1.4(0.2)	1.7(0.4)	0.8

ratio of the principal axes, can be determined as k = (a-1)/(b-1) (Hatcher, 1990, p. 85).

The ratio of the principal axes, k, is a measure of the shape of the pebbles, and it can be represented in Flinn diagrams (Flinn, 1962). Our data is shown in Figure 64. There are two

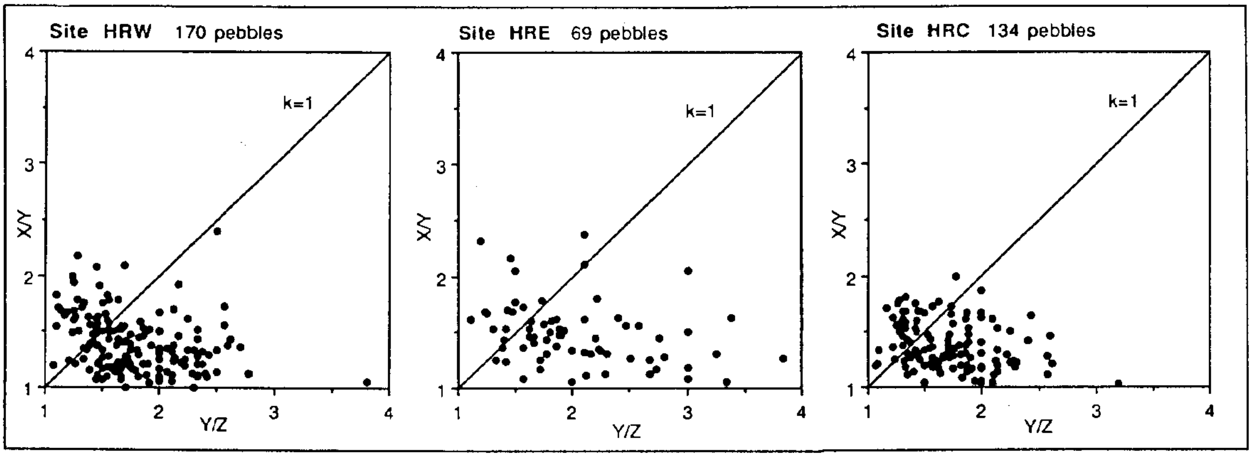


Figure 64. Flinn Diagrams showing shape of loose pebbles weathered from Weverton conglomerate at Hammond's Rocks. X = longest axis, Y = intermediate axis, Z = shortest axis. See text for explanation.

regions on the Flinn diagram separated by k = 1, which has a slope of 1, and would represent a spheroidal pebble. Hence the line subdivides prolate spheroids, which plot in the upper-left part of the diagram, from oblate spheroids, which plot in the lower-right part of the diagram. Hatcher (1990, p. 87) terms prolate pebbles "hot dogs," and oblate ones "hamburgers." Clearly the majority of our pebbles at all 3 sites are "hamburgers," though there are some "hot dogs." We observed that many of the loose pebbles have sharp edges along the periphery of their maximum projection plane. The "hamburgers" exhibit this more often than the "hot dogs." These sharp edges may be analogous to crushed quartz "tails" seen in thin sections of the Chilhowee Group rocks by Fauth (1968, p. 64), but we have not yet examined thin sections from Hammond's Rocks. Some pebbles were striated. It would be interesting to determine striation direction with respect to pebble axes, particularly in the outcrop, but they are difficult to see without proper lighting angle (flashlights anyone?).

It would be ideal to be able to measure all 3 axes of pebbles in the outcrop, but this is rarely possible, so a two-dimensional technique known in the arcane world of structural geology as the Rf/Phi technique is used. A short course follows before we return to the outcrop.

The Rf/Phi Technique--A Short Course

The shape and orientation of initially spherical markers deformed into ellipsoids has long been used to determine strain in rocks. The studies of Cloos (1947, 1971) are classics of the genre. But the technique of determining strain from initially ellipsoidal markers, such as pebbles, has only become common in the last 20 years. The Rf/Phi technique was developed by Ramsay (1967) to deal with these initially ellipsoidal markers. The technique is described in Ramsay and Huber (1983, p. 75-87) and in Lisle (1985), probably the best thorough introduction to the method. There are many variations on the theme, but the basic method will be used here.

The method involves two-dimensional measurements in a plane. It is convenient to choose that plane perpendicular to cleavage and fold axes, but the method can be employed in any plane. Once the plane is chosen on an outcrop, the direction of a reference line in the plane is chosen. It is convenient to make the reference line parallel to cleavage. Then for each pebble, two things are determined (Figure 65): (1) the long and short axis are measured, and Rf = ratio of long axis/short axis is calculated, and (2) the angle Phi (f) is measured between the long axis of the pebble and the reference line (+ if the long axis is clockwise from the reference line and - if counterclockwise). The results for each pebble are plotted on a Rf/Phi diagram, with Rf on a logarithmic scale (see Figure 66 for examples). A sample of at least 30 pebbles should be used.

The method and its results are best demonstrated by showing what happens to ellipsoidal pebbles with progressively greater strain (Figure 66). This was done on a Macintosh computer with the paint program Superpaint (see Bjornerud, 1991 for the technique). A cross section of 37 undeformed pebbles of two different shapes (Rf) was created and their long axes were turned in a variety of directions to simulate "random" orientations. A vertical reference line was chosen, and for each "pebble" the two axes were measured, Rf calculated, and Phi measured with respect to the reference line. Then the original diagram was deformed on the computer, keeping area constant to simulate finite strain. Rf and Phi were determined for each successive deformation. An open circle in the upper left corner of the undeformed sample (Figure 66) becomes an ellipse and shows the strain as deformation progresses. Rs is the strain ratio, which is the ratio of the long to short axis of the open ellipse, and represents the

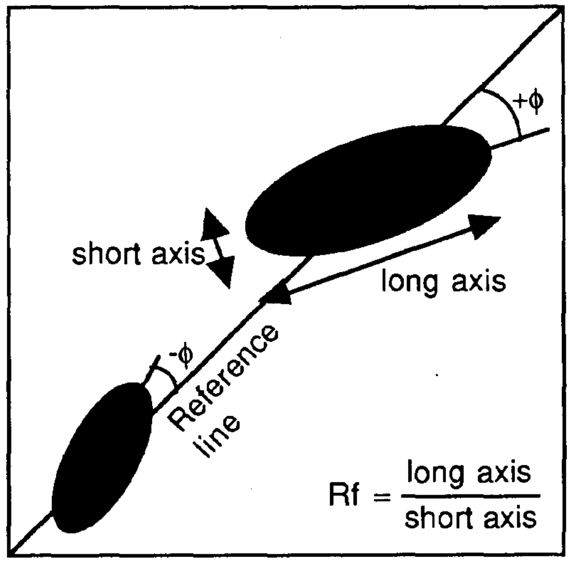


Figure 65. Measurements for the Rf/Phi method.

equivalent of Rf for the open ellipses. Note that as strain increases two things happen: (1) the pebbles become longer and skinnier--that is, Rf increases (except pebbles whose long axes are parallel to the maximum compression axis--they have to become spheres before their long axes switch directions), and (2) the pebbles rotate toward the reference line (cleavage)--that is, Phi decreases and converges toward 0^{o} (except those pebbles with long axes parallel to the maximum compression axis, which finally switch axis direction between Rs = 3.00 and Rs = 7.20).

Rf/Phi diagrams from the resulting measurements are plotted below each set of deformed pebbles in Figure 66. Note that as strain increases, Rf increases and Phi decreases. With progressive deformation, the data cluster becomes longer and skinnier, and plots closer to the center line for the Rf/Phi diagram.

Our introductory "squeeze" was for elliptical pebbles of "random" orientation. Now consider what a fabric and the resultant Rf/Phi diagram would look like if we squeezed initially spherical markers like oolites (circles in two dimensions). Figure 67 (left half) shows the result. All of the circles become ellipses. Because Rf is a ratio of axis lengths, all have the same Rf after deformation (Note: the small scatter in the Rf/Phi diagram is the result of measurement error on small objects), and Rf plots as a tight cluster at 0^{o} Phi.

Now consider what would happen if we deformed initially elliptical pebbles with a strong depositional fabric (Figure 67, right half). Pebbles of different shapes were created and oriented at 45^{o} to the imaginary vertical reference line (and to the maximum compression direction). Note that the Rf/Phi diagram for the undeformed pebbles is asymmetrical with respect to Phi (all plot at $+45^{o}$). The deformed sample shows all of the pebbles rotated toward the reference line and Rf increased. This asymmetry gives a test of whether there was a strong initial depositional fabric. It is not foolproof, for if we had squeezed par-

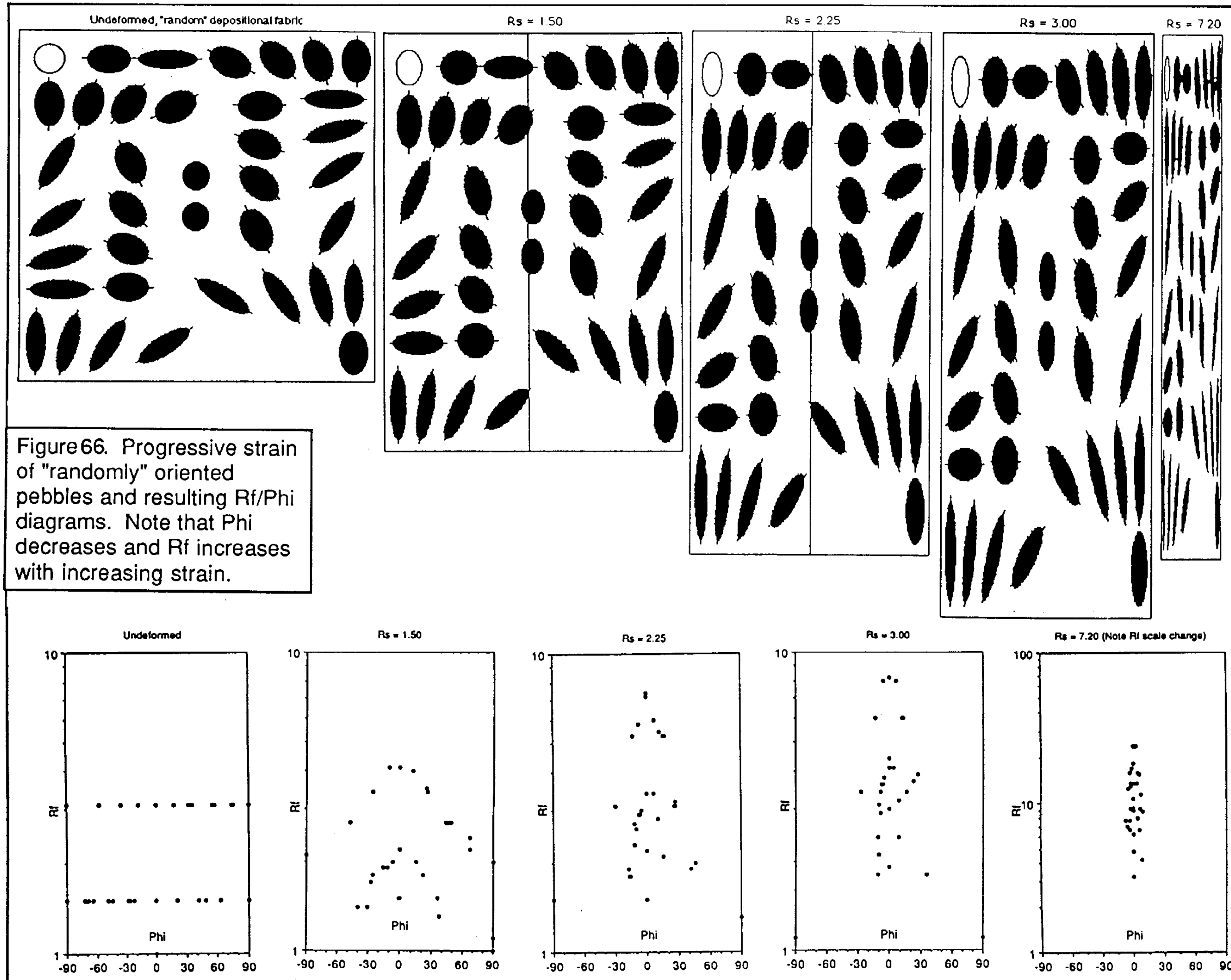
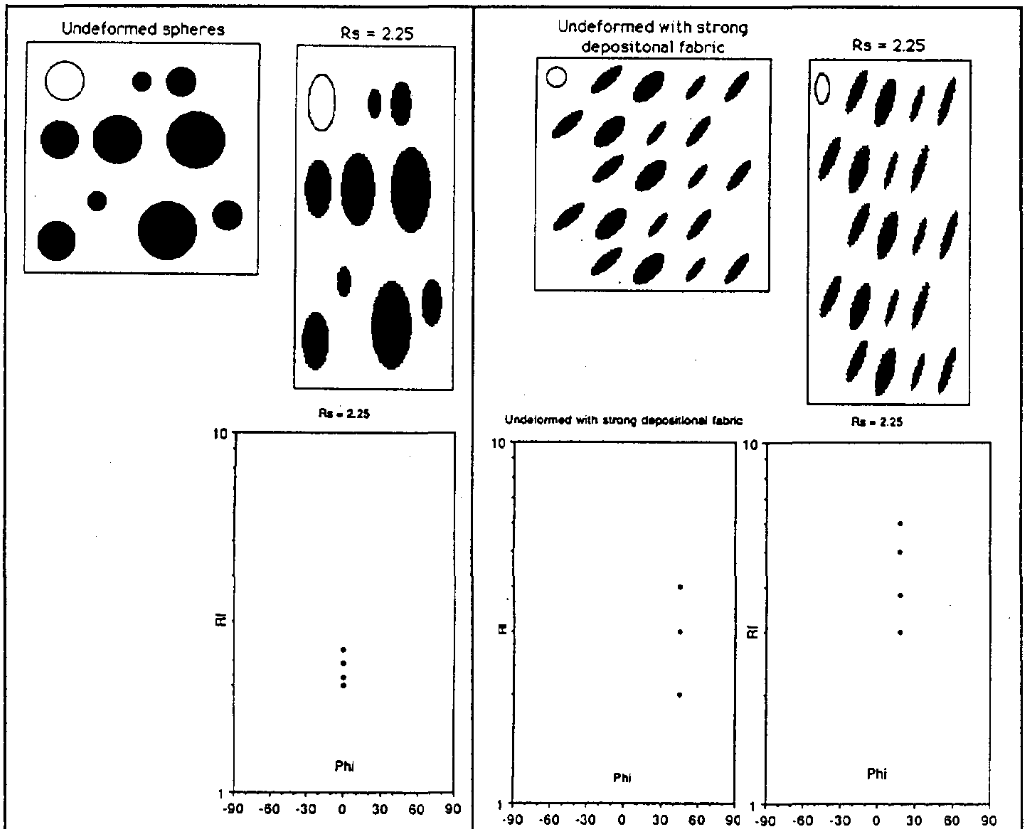


Figure 66. Progressive strain of "randomly" oriented pebbles and resulting Rf/Phi diagrams. Note that Phi decreases and Rf increases with increasing strain.



Deformed spheres produce a Rf/Phi diagram with both Rf and Phi clustered. In this case, the small scatter in Rf is the result of measurement error.	Deformed ellipsoidal pebbles with an original "strong" depositional fabric produce an asymmetrical Rf/Phi diagram with average Phi offset from zero. With greater the strain, however, Phi will approach zero.

Figure 67. Effect of strain on undeformed spheres and on ellipsoids with a "strong" depositional fabric.

allel to either the short or long axes of the undeformed pebbles, the diagram would be symmetrical about 0^{o} Phi. But by making Rf/Phi measurements at several sites on an outcrop (particularly one with variable dips of bedding) one can increase confidence that symmetrical Rf/Phi diagrams represent no strong initial depositional fabric.

We have now considered qualitatively what can be done with Rf/Phi diagrams. They can also be treated quantitatively, by making assumptions about the style of deformation. The process is somewhat tedious, but rewarding (see Lisle, 1985 for details). Briefly, a series of symmetry tests can be applied to the data, and if these are passed, the data can be fitted to a series of templates (Lisle, 1985) from which one can estimate Rs, the strain ratio.

We can now return to Station III where we applied the Rf/Phi method. We chose the vertical face perpendicular to cleavage, and chose the intersection of cleavage with the face as the reference line to measure Phi angles from. For each pebble, we measured the long and short axes in the vertical plane, then placed a dowel parallel to the long axis and measured Phi with a protractor. For this site (Figure 68, Site D), we measured 40 pebbles. The data is visually symmetrical about 0o Phi, but there are some high Phi angles. We applied the same technique at 4 other sites on Hammond's Rocks (Figure 68, locations on Figure 58) with similar results. Summary data for all 5 sites is given in Table 17).

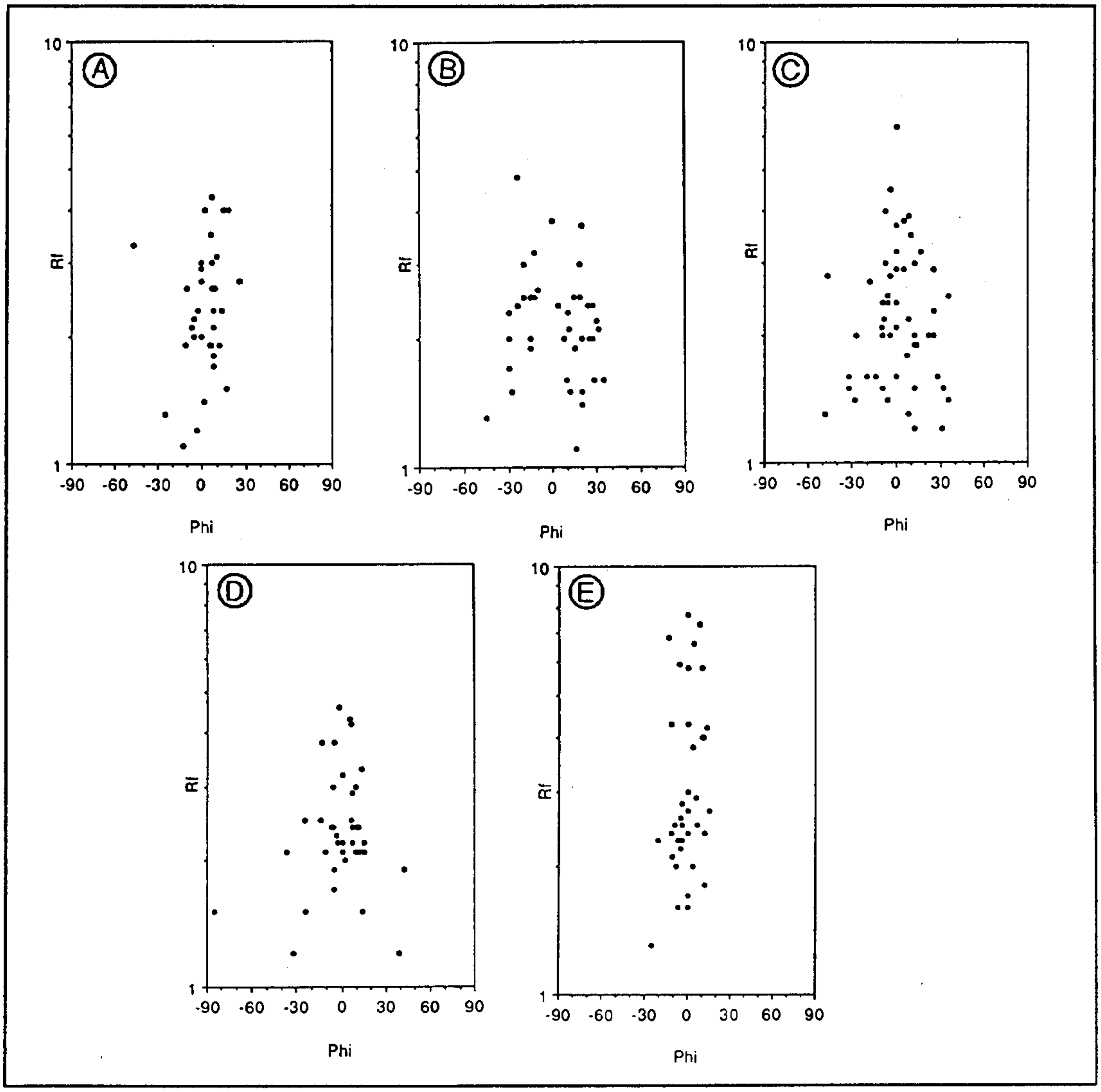


Figure 68. Rf/Phi diagrams for deformed pebbles from 5 sites at Hammond's Rocks. All measurements were made in a vertical plane perpendicular to cleavage. Site locations (A through E) are on accompanying map (Figure 58). Summary data for sites is in Table 17.

Table 17. Hammond's Rocks Rf/Phi data summary.

Site	Number of Pebbles	Vector Mean Phi	Harmonic Mean H	I_{SYM}	Estimated RS
A	35	+0.7	2.2	0.80	2.1
B	40	+0.7	2.1	0.80	1.9
C	53	+0.6	2.1	0.83	1.9
D	40	+0.5	2.2	1.00	2.0
E	40	-0.9	2.7	0.75	2.5

At all 5 sites mean Phi is within 1^{o} of zero Phi. The harmonic mean is a measure of average Rf, and according to Lisle (1985, p. 15) H "approximates, and slightly overestimates the strain ratio" (Rs). ISYM is the index of symmetry. High values (approaching 1) indicate high symmetry. Data from all 5 sites passed the symmetry test for the sample size. The estimated strain ratio, Rs, was determined by using the templates in Lisle (1985). The average strain ratio is somewhat over 2.0. It is worth noting that Site E has a higher strain ratio. Note that the Rf/Phi diagram for this site (Figure 68) has higher Rf values than the other sites, and the data clusters closer 0^{o} Phi. The greater deformation can be seen in that part of the outcrop (Figure 58).

We conclude that most of the deformed pebbles at the 5 sites where we applied the Rf/Phi technique were initially ellipsoidal and had no discernable depositional fabric before deformation. A simple view of the deformation would have the pebbles squashed perpendicular to cleavage and elongated parallel to cleavage. The compression perpendicular to cleavage rotated the pebbles toward cleavage, producing low Phi angles. Pebbles with initial random orientations of their long axes deformed in this manner would produce the deformational fabric with pebble long axes lying in a variety of orientations in or near the plane of cleavage (Figures 60 and 61). Pebbles with their long axes initially oriented at high angles to cleavage have not been sufficiently rotated or squashed so that a few pebbles still have high Phi angles. (By now all but the die-hards are ready to become geomorphologists!)

Station IV

Here one of the largest blocks in Hammond's Rocks can be demonstrated to have moved about 1 meter downslope. In the outcrop beneath the east end of the block are 3 quartz veins with distinctive spacing. The matching 3 quartz veins can be found in the bottom of the block. Walk around the block and consider how it has moved. We think it rotated, with the east end moving downslope more than the west end. Consider climbing atop this rock and making some structural measurements when you don't know that it has moved--movement is not obvious unless you observe carefully. There are many blocks like this one in the "outcrops" here, and the lesson is that caution should be used to be sure that one's outcrops are "in place."

Station V

A series of deformed extension quartz veins here could be used to decipher the deformational history, but we have not worked on them yet. There are several other groups on the outcrop.

Station VI

On a clear day climb the steel ladder for a fine view over the top of the trees. To the north you can see 14 miles across the Great Valley to the first ridge in the Appalachian Mountain Section of the Ridge and Valley Province. Immediately to the south is the valley of Mountain Creek (location of Stop 10), and beyond to the Mesozoic Gettysburg Basin.

Station VII

Is the boulder seen here (Figure 69), looking east just beneath the highest point of the rocks, in place? It shows

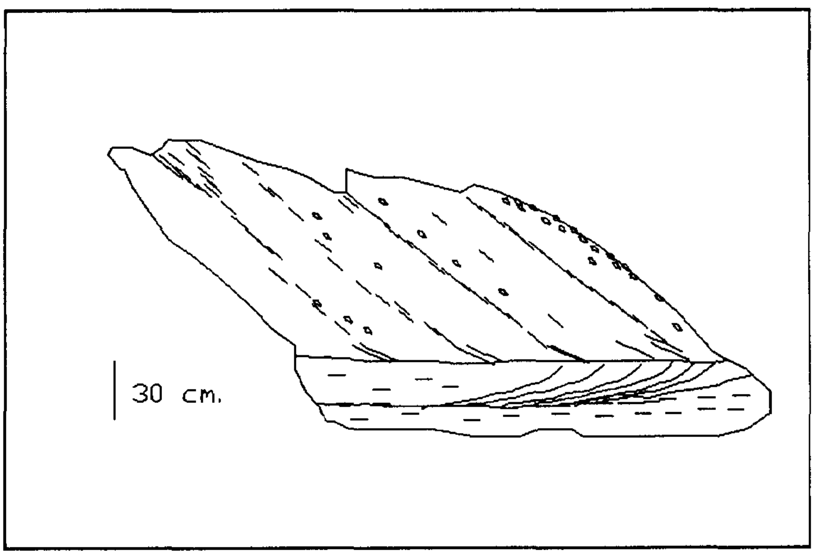


Figure 69. Crossbedded boulder at Station VII.

several well-defined beds. An upper pebbly sandstone bed, about 1.5 m thick, overlies a sandstone lens and pebbly sandstone about 30 cm thick. In the upper bed, pebbles have been rotated into the plane of cleavage, which cuts the upper bed at about 40°. The pebbles seem to be concentrated in crossbeds. Crossbeds in the sand lens at the bottom suggest an opposite transport direction, but the arcuate crossbeds in the sand lenses that are exposed in three-dimensions at Station III could, in a similar section here, produce a similar contrast between two partly viewed lenses without the apparent large difference in transport direction. The lesson is to use caution inferring crossbed orientation until you can see the third dimension.

If this boulder is crossbedded, then it suggests a minimum depth of flow for 1.5 meters. What bedforms do these crossbeds represent--migrating dunes, sand waves, point bar?

At this point you may wish to consider what depositional environment these rocks represent. There is both the direct evidence of small channels and a suggestion of bedforms on the order of a meter in height. These features, found in what appears to be an isolated section through coarser than normal Weverton Formation, suggest that the rocks in the entire outcrop here might represent a large channel deposit. Such features, if they exist, should be interpretable from detailed mapping of the few hundred beds that appear to be exposed here.

Large channels occur in a variety of depositional environments from continental slope to mountain stream. In the instance of this possible channel in the entire Weverton Formation, which is basically a dirty sand body on the order of hundreds of feet thick and occurring over an area of several thousand square miles, some depositional environments are more probable than others. Were the sand a clean quartz sand, the channel would probably represent a channel through a moderately high energy beach. If the interstitial materials, the "dirty stuff," was derived from sand-grade rock fragments by cataclasis or other processes during deformation, which seems possible, then the channel could still represent a channel across a beach or barrier. If, on the other hand, the "dirty stuff" was deposited as silt and clay, then the depositional environment would seem more likely to have been a channel through a sandy delta, an estuary, or across a lagoon, or even a low energy beach.

We measured the orientation of about a dozen topset and foreset beds, and rotated the topset beds to horizontal on a stereographic net to determine paleocurrent directions. This yielded a wide variety of foreset dip directions with implied source of current from every direction except the east. It also yielded a number of foreset bed dips that were much higher than a reasonable angle of repose for sand, some as high as 50-60°. We do not yet know the reason for the high "initial" dips, but one possibility is that the angles between topset and foreset beds have been changed by subsequent deformation. Whitaker (1965) found an "overwhelming number" of the 136 crossbeds in the Weverton that he studied in the Blue Ridge in Maryland showed sources from the west.

Station VIII

This is a good exposure without graffiti in which to consider the variety of lithologies represented by the pebbles here for comparison to those in the underlying Loudoun Formation that we will see at Stop 9. Quartz pebbles are dominant. The small percentage of slate chips and perhaps rhyolite fragments here appear to be similar to those in the Loudoun (Stop 9). Thus there may be a "local" source for some of the material in the underlying Precambrian rocks. It should be noted however that despite the dominance of metabasalt in the Precambrian of South Mountain, no metabasalt pebbles have been found here. The non-opaque heavy mineral suite from the Weverton sand is dominated by zircon with rare tourmaline and rutile. Leucoxene-coated sphene, reported by Fauth (1968, p. 106) to be common in metarhyolite, is not obvious in the Weverton here.

Future work

Much remains to be done at Hammond's Rocks. Future work should include: (1) a detailed sedimentological description, bed-by-bed, of the section of Weverton exposed here and analysis of the depositional sequence, (2) an analysis of cross-beds and their usefulness as paleocurrent indicators, (3) examination of oriented thin-sections to discover the how the matrix surrounding pebbles has behaved during deformation, (4) analysis of the quartz veins for the contribution that they surely can make to the deformational history, and (5) detailed geomorphic analysis of the blocks that appear to have moved away from the main outcrop under periglacial conditions. In addition, some of the techniques we have used could be applied to several other good outcrops to the north and east of Hammond's Rocks.

		LEAVE STOP 8. PROCEED STRAIGHT AHEAD on Ridge Road. Be alert for flattish topographic surfaces that don't make sense if the underlying bedrock is composed of highly-deformed lithologies of varying relative resistance to weathering and also have steeply dipping parting planes. They could be the result of cryoplanation (See Stop 2 and Clark, this guidebook).
0.8	20.0	**CROSS ROADS. PROCEED STRAIGHT AHEAD.**
1.0	21.0	Note flat surface. Outcrops would allow us to determine whether of not this surface might be structurally controlled or if cryoplanation is a possible explanation for the surface.
1.0	22.0	**STOP SIGN. TURN LEFT** onto PA Route 233 S.
2.0	24.0	**STOP SIGN. TURN RIGHT** following PA Route 233 S at entrance to Pine Grove Furnace State Park.
0.1	24.1	**TURN LEFT** onto Bendersville Road.
0.1	24.2	**TURN LEFT** following Bendersville Road at road fork below brick ironmaster's mansion. The route ahead crosses Piney Mountain which is underlain by Weverton Formation.
4.2	28.4	**STOP SIGN. TURN RIGHT** onto Coon Road (SR 4012).
2.6	31.0	**STOP SIGN. TURN RIGHT** onto Wenksville Road. Route ahead and rugged terrain to left is on metavolcanics. The Mesozoic Lowland is about 3 miles to the southeast.
3.4	34.4	**PARK** on right side of road.

STOP 9. LOUDON FORMATION

Discussant: Henry Hanson

Walk from the road northwest beneath power power line to base of steep slope. Most of the flat surface is underlain by Precambrian metabasalt. A few float boulders on the flat are metabasalt, but most are pieces of the Loudoun Formation which have moved downslope over the metabasalt. The lowermost Cambrian Loudoun Formation underlies the steep slope beneath the power line, approximately to the poles at the crest of the steep rise. Beyond the poles, there are some exposures of the Weverton Quartzite.

Fauth (1968) recognized two subdivisions of the Loudoun in this area:(1) a basal phyllite, which is typically reddish-purple to gray, and (2) an upper conglomerate member composed of interbedded sandstone and conglomerate with rounded pebbles 1/2 to 2 inches in diameter. The phyllite member is not exposed here, but a few float fragments at the base of the steep slope beneath the power line may be from the phyllite member. The phyllite can be found as bedrock along Old Railroad Road, about 2 miles east of the power line. The outcrops and boulders on the steep slope beneath the power line nicely display the range of lithologies in the upper conglomerate member.

For those who don't wish to climb the hill, the major lithologies of the conglomerate member can be seen in float boulders at the base of the slope. For those who are more adventurous, a climb up the first ridge (Figure 70) will provide opportunity to examine sedimentary features at several in situ bedrock exposures.

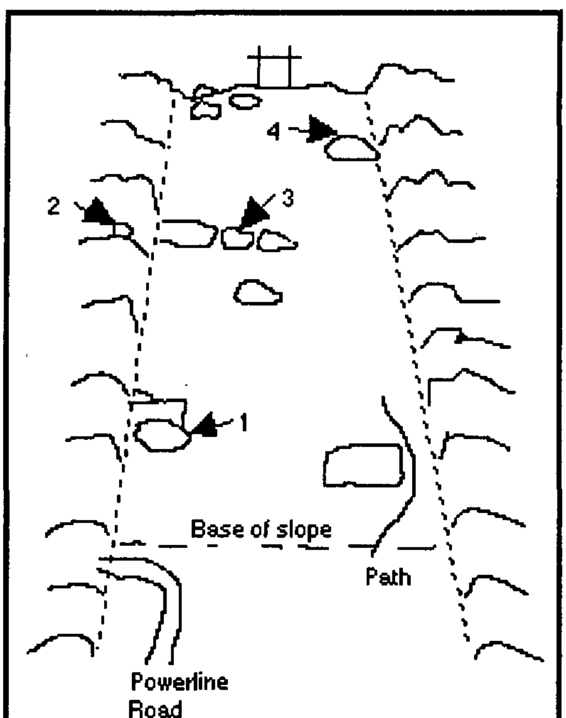


Figure 70. Stop 9 looking NW along powerline. Numbers are stations referred to in text.

Beds in the exposures on the steep slope are overturned. They strike about N50°E and dip about 80° SE. Cleavage strikes about N45°E and dips about 35°SE. Cross-bedding in several exposures confirms that bed tops are toward the northwest.

Compare the lithology of the Loudoun Formation here with that of the Weverton Quartzite seen at Stop 8 (Hammond's Rocks). In particular, consider the composition of the pebbles in the conglomerate here. Pebbles in the Loudoun here are dominantly quartz, but fragments of shale chips, pink quartz, and jasper and/or rhyolite do occur. Fauth (1968, p. 107-108) reports that thin sections of the conglomerate "show pebbles of quartz, quartzite, rhyolite, and phyllite plus smaller grains of quartz in an aligned matrix of quartz, muscovite, and sericite." The Weverton at Hammond's Rocks contains mostly quartz pebbles, and other lithologies are rare. It is interesting to note both here and at Snowy Mountain (Stop 4) where the Loudoun overlies metabasalt, that metabasalt does not occur as pebbles, but metarhyolite does.

FEATURES TO OBSERVE (refer to Figure 70 for locations):

Station 1. A relatively fresh 6 x 10-foot face on a float boulder (Figure 71) shows two crossbedded sandstone layers, one about a foot thick and the other half as thick. These beds are more uniform in thickness than the sandstones at Hammond's Rocks. Both crossbed sets show a similar transport direction.

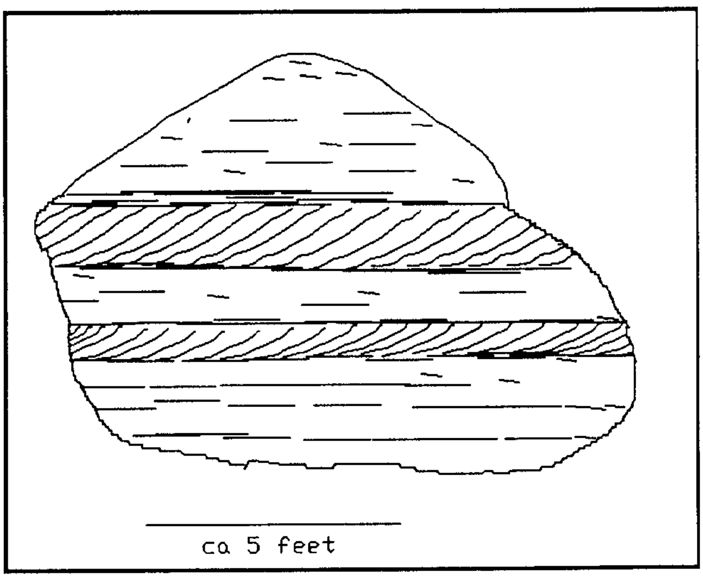


Figure 71. Crossbedding in Loudon Formation at Station 1.

Station 2. A float boulder about 50 feet into the woods is about 8 feet above ground level on top of an outcrop. It shows a sandstone lens cut into a conglomerate bed (Figure 72). The channel is reminiscent of those at Hammond's Rocks, but the conglomerate here contains a much greater abundance of rhyolite and shale chips.

Station 3. Straight northeast along contour from Station 2 are outcrops beneath power line. Here there are outcrops of interbedded conglomerate and sandstone. One conglomerate bed is at least 3 feet thick. You can: (a) examine pebble lithologies,(b)

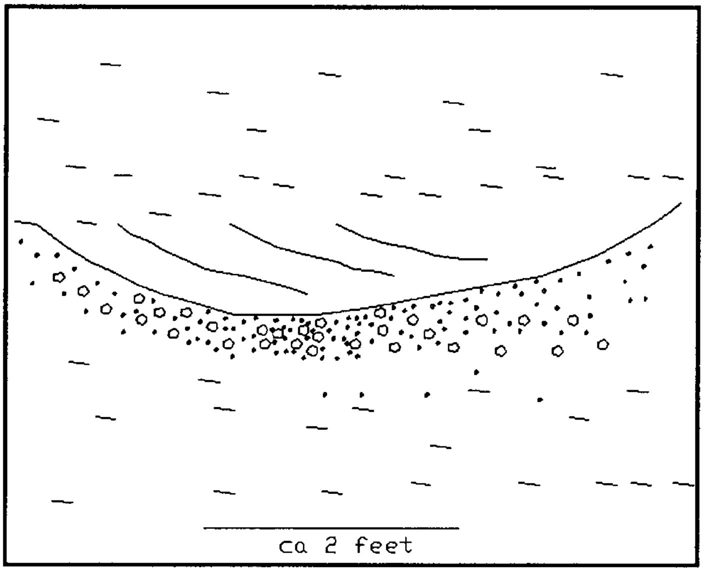


Figure 72. Channel in Loudon Formation at Station 2.

see the relationship of bedding (overturned) to cleavage, and (c) see pebbles flattened in the plane of cleavage (are some lithologies more deformed than others?).

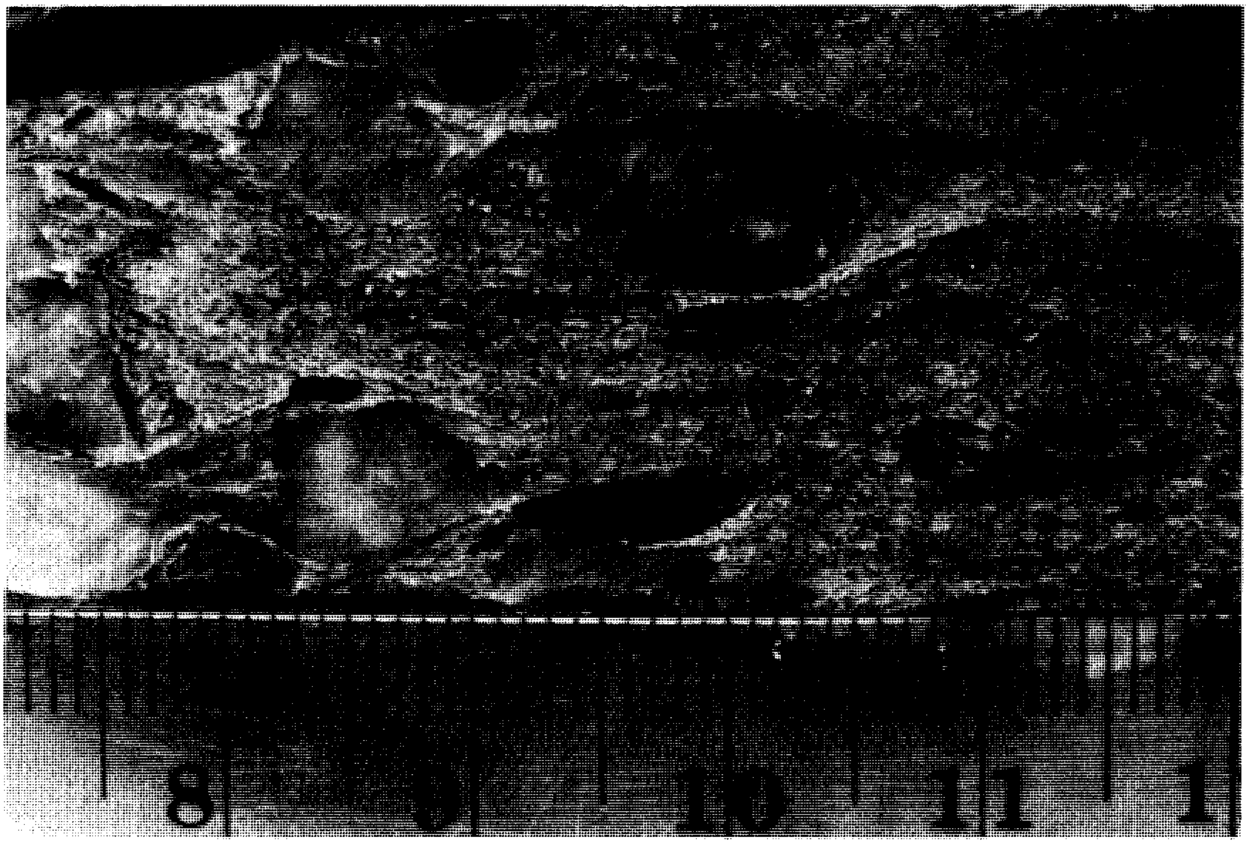


Figure 73. Sawn slab of Loudon conglomerate with "tails" of sand around pebbles indicating rotation of pebbles during deformation. Scale in cm.

Station 4. Interbedded sandstone and conglomerate in place. The conglomerate here has the potential for strain analysis. A sawn slab of conglomerate shows that many of the clasts have been rotated during deformation, as shown by curved "tails" of bedded sand around the pebbles with a consistent sense of rotation (Figure 73).

		LEAVE STOP 9. PROCEED STRAIGHT AHEAD.
0.4	34.8	**STOP SIGN. TURN RIGHT** onto Arendtsville-Shippensburg Road (SR 4009).
1.7	36.5	**STOP SIGN. TURN RIGHT** onto PA Route 233 N. The Conococheague Creek-Mountain Creek drainage divide is 0.5 mile to the left of this intersection.
6.0	42.5	**TURN RIGHT** onto Bendersville Road.
0.1	42.6	**STOP SIGN. TURN RIGHT AND THEN IMMEDIATELY LEFT** onto Quarry Road.
0.2	42.8	**PARK** in lunch area. Circle buses first. **STOP 10 AND LUNCH.**

STOP 10. GUIDE TO THE PINE GROVE FURNACE AREA

Discussant: John H. Way

Only a few traces of the once vigorous iron plantation at Pine Grove have survived. The stone furnace stack, the iron-master's mansion and some other buildings, ore pits, a limestone quarry, and a pit that provided rock as liner for the furnace provide some idea of this vanished iron industry.

1. Pine Grove Furnace. The restored stone furnace stack is all that remains of a cluster of buildings that formerly surrounded the stack (Figure 74). Built in 1764 (Bining. 1979, p. 50), Pine Grove Furnace was the second of nine furnaces constructed in Cumberland County. The first furnace on the site was primitive, constructed of local stone and contained only one tuyere (the nozzle through which the cold-air blast was delivered to the furnace). The property passed through several owners, each making modifications and improvements. Prior to its abandonment around 1891, it was modified to a hot-blast operation. But the day of the charcoal-iron furnaces was at an end. Steel was the metal of the future, and the coal- and coke-fired mills in the west were about to take over.

2. The Ironmaster's Mansion. Overlooking the furnace and much of the Pine Grove operation is the prominent L-shaped house known as the "big house" or "mansion house." This stately and spacious stone building was the home of the ironmaster and his family and served as a business office as well. Its rooms were said to have been handsomely furnished, its cupboards full of delicate china and glassware, and on cold winter days, its massive chimneys spewed smoke from its many fireplaces, a sign of warm hospitality to official visitors as well as members of the community. The ladies of the iron master's family *"dispensed medicine in times of illness and gave advice and smoothed out domestic difficulties in many cases.... [and] were, as we are told, 'great ladies' to*

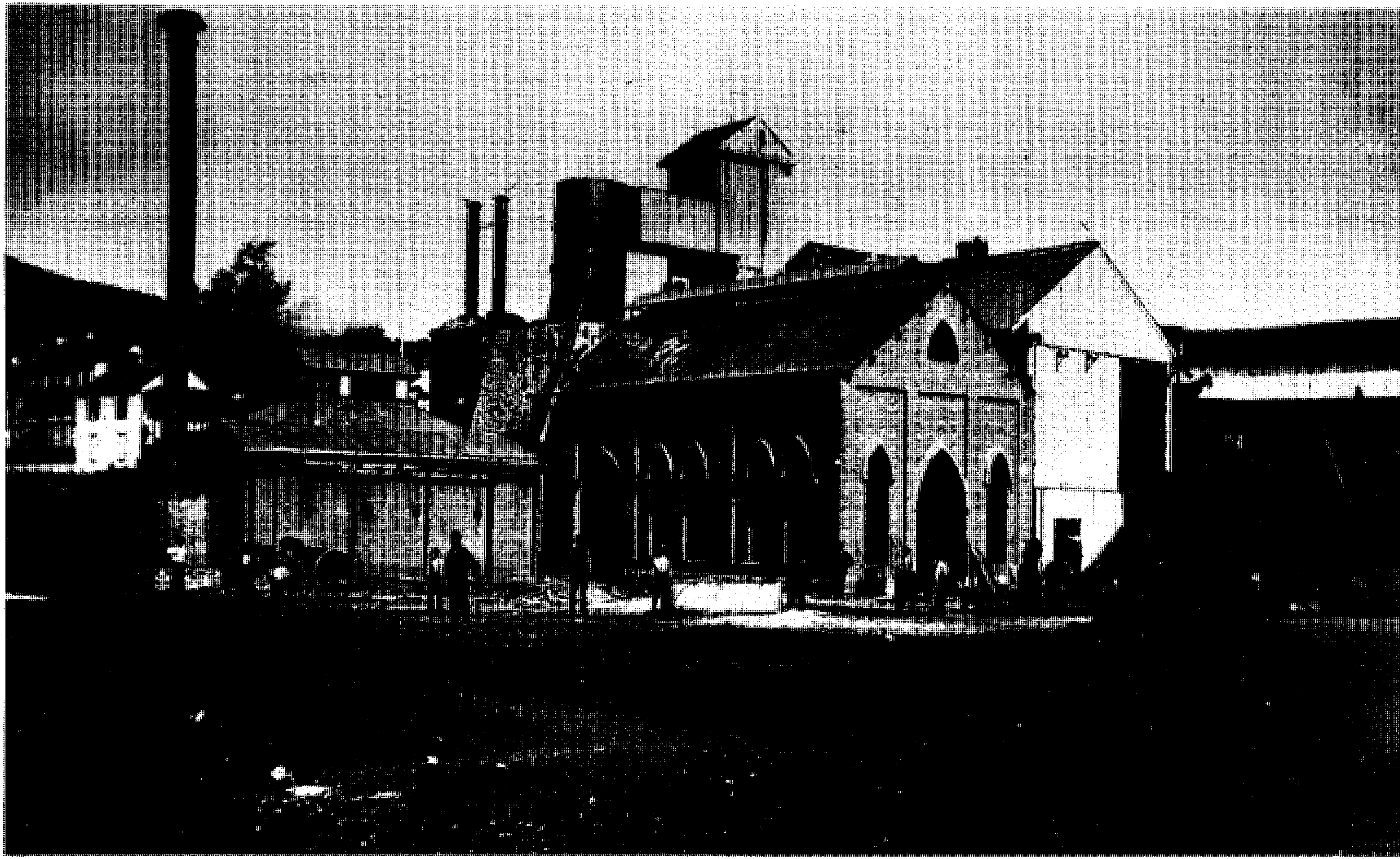


Figure 74. A northwest view of Pine Grove Furnace circa 1875. The ironmaster's mansion is in the background on the left and the stone furnace stack is near the center surrounded by buildings. Photograph courtesy of Cumberland County Historical Society (Potter, 1981, Figure 11, p. 23; Way, 1986, Figure 7-4, p. 14).

the workers" (Flower, 1975, p. 16).

3. Iron-ore Pits. The South Mountain landscape is covered with shallow, crater-like depressions, the remnants of iron-ore mining. The largest open-pit operation in the Pine Grove area is now filled with water and is Fuller Lake. Fragments of limonite can be found surrounding many of the old iron-ore pits. These assume many shapes and vary considerably in quality of iron.

Impurities, especially manganese and phosphorus, occur in some areas, and these presented serious problems to the iron-master. Two common manganese-oxide minerals, pyrolusite and cryptomelane, occur locally with the limonite. Both tend to be gray-black to black in color and frequently appear as nodules or in grape-like clusters. Pyrolusite is softer with a dull, earthy luster and cryptomelane is harder and more dense. A variety of other minerals occur in these residual iron-manganese deposits including wavellite, beraunite, cacoxenite, strengite, and turquoise (Way, 1986, p. 27).

4. Limestone Quarries. Immediately south (0.1 and 0.2 km) of Fuller Lake are two quarries where Cambrian Tomstown limestone was removed as flux for the furnace operation. Rodgers (1858, p. 205) reported that limestone from these quarries was carried south over the mountain into Adams County where it was used as agricultural lime. He also noted the presence of fluorite here. Limited exposures are visible in the southern-most quarry.

5. A Talcose Schist Pit. Traces of a very fine-grained sericite schist or phyllite with a 'soapy' feel occur in two overgrown trenches approximately 0.7 km northwest of Pine Grove Chapel on a paved road labelled "Appalachian Trail " on the Dickinson 7.5-minute quadrangle.

Rodgers (1858, p. 204-205) reported *"In the ridge N. of Pine Grove the rock is a more, talcose sandstone... Large veins of white quartz are here abundant. A whitish talcose slate rests conformably upon the talcose sandstone, dipping with it to the S. E. This latter rock forms an admirable material for the in-walls of a furnace and is used in that at Pine Grove."*

		LEAVE STOP 10. RETURN WAY CAME.
0.1	42.9	**STOP SIGN. TURN RIGHT. BEAR RIGHT** past the camp store.
0.1	43.0	**STOP SIGN. BEAR RIGHT** onto PA Route 233 N.
0.1	43.1	**BEAR LEFT** following PA Route 233 N.
2.0	45.1	Mountain crest.
2.3	47.4	**ROAD FORK. BEAR RIGHT** onto Point Road. This fork and road are not obvious until one is there.
0.4	47.8	**STOP SIGN. PROCEED STRAIGHT AHEAD.** Note the abundant karst on the left. The whole surface is covered with gravel.
0.8	48.6	**STOP SIGN. TURN RIGHT** onto Pine Road (SR 3006) at Huntsdale.
0.5	49.1	Poor view to left of a borrow pit in the distal part of the alluvial fan gravels derived from South Mountain. This site is described in Sevon 1985 and 1989. For the next 5 miles Pine Road crosses the distal edges of gravel fans.

0.9	50.0	Road on right leads to Kings Gap Environmental Education Center, a place well worth visiting.
3.1	52.1	Cross railroad. View to right of Stop 11.
0.9	54.0	**TURN RIGHT** onto Mountain View Road at Pennsy Supply sign. Traverse up a gravel surface.
0.8	54.8	**SHARP TURN. BEAR LEFT.**
0.1	54.9	**TURN RIGHT** following Mountain View Road.
0.2	55.1	**GATE.** Entrance to Pennsy Supply quarry.

STOP 11. GEOLOGY OF THE PENNSY SUPPLY QUARRY, MT. HOLLY, PENNSYLVANIA

Discussants: Marcus M. Key, Jr. and Samuel J. Sims.

Directions. Go through the gate and straight past the metal building on the right and up the hill. The road swings around to the right and heads up hill. At the fork, take the outside (right) lane which leads to Bench 1. Come down the same way.

GEOLOGY

The Pennsy Supply Quarry (formerly the R. A. Bender and Son Quarry) is 2.5 km west-south-west of Mt. Holly Springs in Dickinson Township, Cumberland County, Pennsylvania (Figure 75). The

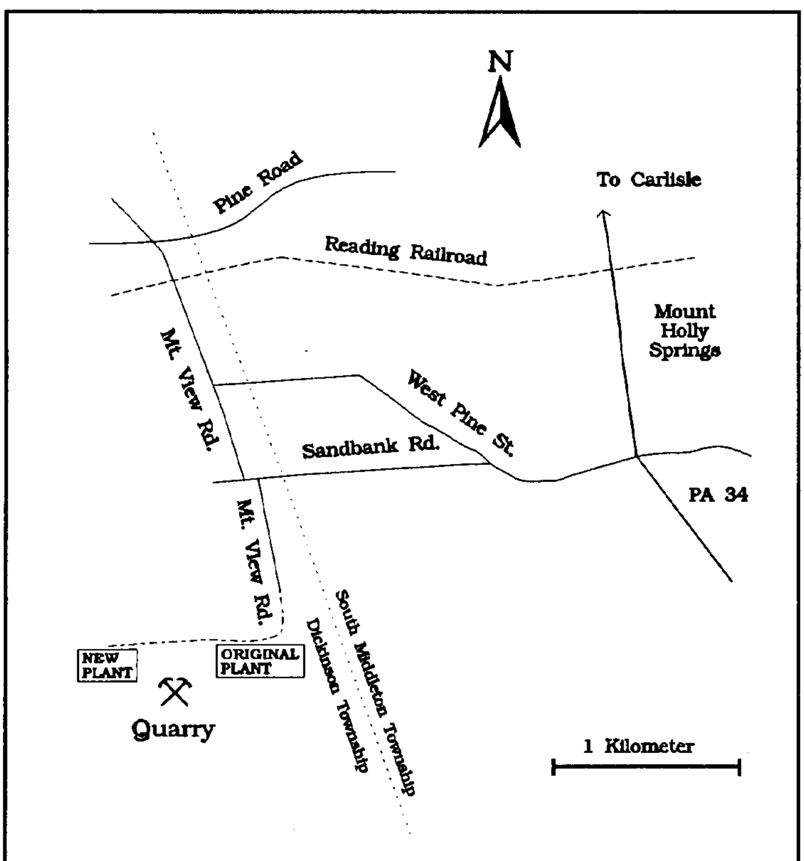


Figure 75. Map showing location of quarry.

quarry is on the northwest slope of Mt. Holly at the northeast end of the Blue Ridge physiographic province between the Great Valley to the north and west and the Piedmont to the south and east. The quarry exploits the Antietam Quartzite, which was deposited during the Lower Cambrian (See Key, this guidebook). The Antietam Quartzite was later folded during the Paleozoic Appalachian orogenies that formed the South Mountain Anticlinorium. The quarry straddles the axis of the Bender's Quarry Anticline (Freedman, 1967; 1968) which plunges gently to the northeast. The present quarry is in the northwestern limb of this anticline (Sims, 1985).

The Antietam Quartzite lies on the upper phyllitic member of the Harpers Formation and is overlain by the Tomstown Dolomite. The Harpers Formation crops out upslope (to the south and east) from the quarry while the more-soluble and easily-eroded Tomstown Dolomite forms the southern margin of the Cumberland Valley to the north and west. The Harpers Formation and Tomstown Dolomite do not crop out in the quarry. Based on the distribution of float and topographic slope changes, the Antietam Quartzite is estimated to be at least 440 feet thick in Mt. Holly (Freedman, 1967).

The lithology of the Antietam Quartzite that is exposed in the quarry varies greatly. Textures range from fine- to coarse-grained, subangular to subrounded, well- to poorly-sorted, thin-bedded to massive, with local crossbedding, laminations, and numerous *Skolithos linearis* worm tubes. Pre-deformation lithologies varied from thin clays, to quartz arenites, to granular quartz sandstones. Post-depositional deformation has altered these rocks to phyllites and quartzites. In some areas, slight schistose development is evident. The most common lithology is a relatively clean, massive, well-sorted, silica-cemented quartzite with up to 99 percent rounded quartz grains (Freedman, 1967; Fauth, 1968).

The *Skolithos linearis* worm tubes are ubiquitous in the Antietam Quartzite and are well illustrated in the slightly weathered, clean, indurated quartzites visible in the large blocks that have been set aside along the southwest edge of Bench 1 (Figure 76). Originally, these worm tubes had circular transverse, cross-sectional shapes. Because of subsequent stresses during the Appalachian orogenies, they have been deformed to a more-elliptical shape. Measurements of the resulting strain on the cross-sectional shapes of four worm tubes reveals a ratio of the short to long axis of 0.61.

The lithology and structure of the Antietam Quartzite are difficult to see in the quarry because of the massive nature of some units and well-developed jointing and weathering. It is especially difficult to determine the orientation of bedding. Look at the cliff between Benches 1 and 2 (Figures 76 and 77) for bedding planes. There are two reliable indicators of bedding plane surfaces: (1) the presence of clay layers interbedded in the quartzites and (2) *Skolithos linearis* worm tubes perpendicular to bedding. The strike of bedding ranges from N60^{o}E to N90^{o}E, and dips range from 45^{o}NW to vertical to 30^{o}SE; the dominant jointing strikes N47^{o}E and dips 38^{o}SE (Figure 77). These joints have been

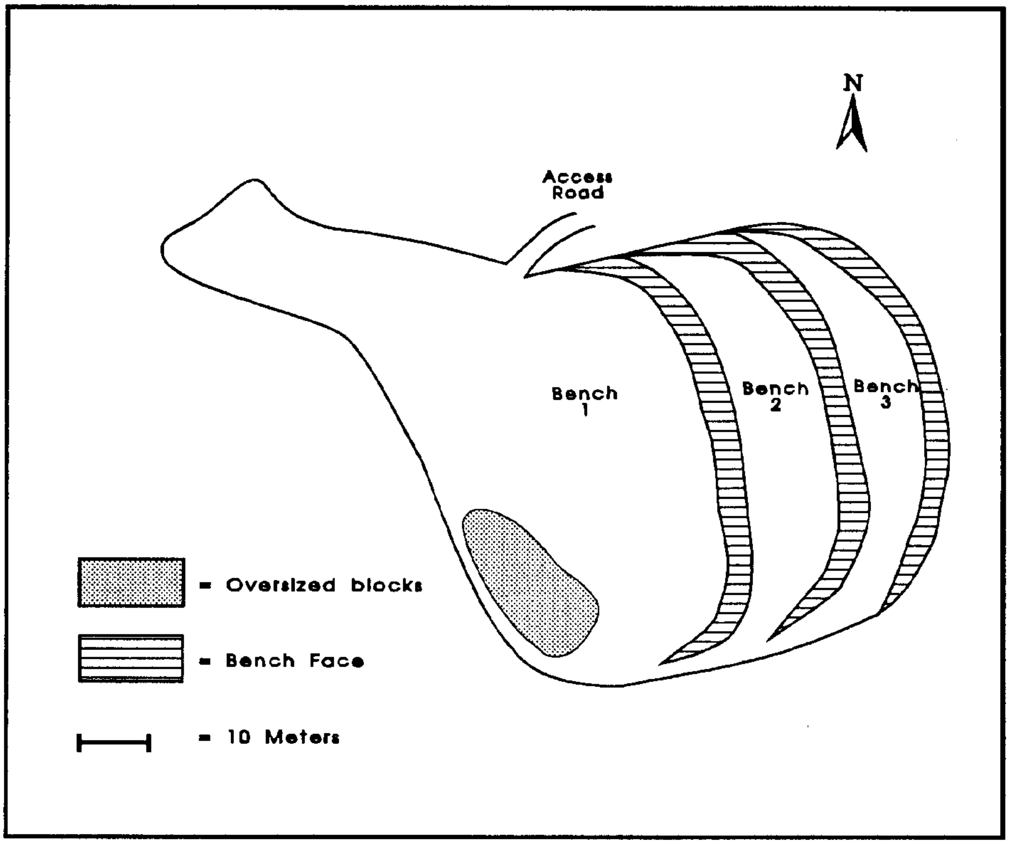


Figure 76. Diagrammatic plan view of active portion of quarry as it appeared in June, 1991.

interpreted as extension joints parallel to the regional trend of the Bender's Quarry Anticline (Freedman, 1967). The divergent attitude of the joints and bedding results in blocky exposures.

Weathering causes the Antietam Quartzite to become extremely friable with a yellowish-tan to pinkish-red color. In some places such as in the fault zone along the crest of the anticlinal ridge, weathering may extend down as deep as 40 to 50 feet (Freedman, 1968).

What caused this deep weathering? Freedman (1967, 1968) argued that it was caused by leaching of the silica cement. He hypothesized that the silica dissolution was due to percolating groundwater which became alkaline as it seeped down through the overlying Tomstown Dolomite (Freedman, 1967). This does not seem plausible due to the difficulty of leaching silica in non-tropical climates (Friedman and Sanders, 1978).

Another alternative is that the deeply weathered zones reflect a greater abundance of clay. To test this hypothesis, the type and relative abundance of clays in three Antietam Quartzite samples from the quarry were determined using X-ray diffraction. Results indicate that illite and kaolinite clays are present. The amount of illite is greatest in the clay layers interbedded between the quartzites, intermediate in the indurated quartzite, and least in the friable quartzite. The amount of kaolinite is greatest in the clay layers interbedded between the quartzites, intermediate in the friable quartzite, and least in the indurated

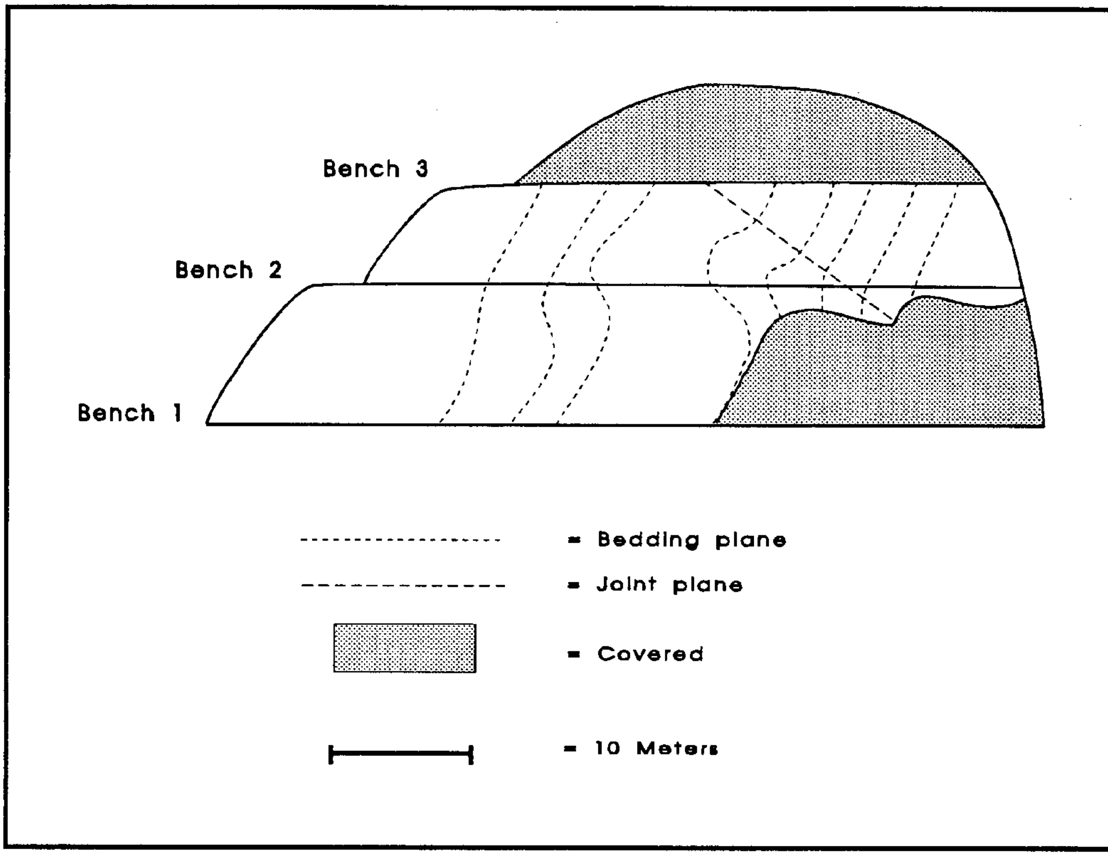


Figure 77. Diagrammatic cross section of Antietam Quartzite as it appeared in June, 1991 viewed from Bench 1 looking looking down strike (N77^{o}E).

quartzite. The interbedded clay layers are more commonly associated with the friable quartzite than the indurated quartzite. The deep weathering in some zones may simply indicate a greater kaolinite clay content. These deeply weathered zones are well known to the quarry operators because the clay-rich layers present a considerable problem of generation and disposal of slimes and are thus less desirable from an economic standpoint.

Spectacular liesegang rings are visible in the cliff between Benches 1 and 2 (Figures 76 and 77) as well as in the oversize blocks that have been set aside along the southwest edge of Bench 1 (Figure 76). Liesegang rings are secondary concentric rings formed by repeated precipitation within a fluid-saturated rock. Water in the once-saturated rock presumably transported iron in solution and precipitated it as hydrated iron oxide which we see today as rings of limonite. The liesegang rings found in the quarry are commonly formed around fractures in the quartzite. In the quarry, limonite can occasionally be found as pseudomorphs after another mineral, possibly pyrite. More commonly it occurs in botryoidal encrustations along fracture surfaces.

A problem is where did the iron come from because it is very rare in the Chilhowee Group. It may have come from dissolution of the overlying Tomstown Dolomite (Freedman, 1967). Dissolution of the Tomstown Dolomite has been associated with the economic iron deposits on the south slope of South Mountain (Stose, 1932). X-ray diffraction analysis of three Antietam Quartzite samples from the quarry revealed the presence of some dolomite or high magnesium calcite which may indicate dissolution of the overlying Tomstown Dolomite and reprecipitation in the Antietam Quartzite. This hypothesis implies that the leisegang rings formed when the Tomstown Dolomite overlay the Antietam Quartzite. How long ago did the Tomstown Dolomite rest on the Antietam Quartzite at South Mountain? The fact that Triassic rocks rest on Cambrian rocks in the Gettysburg Basin, suggests that the erosion of the lower Paleozoic carbonates in that area occurred before or while the Triassic rocks were deposited. If this is also true for the South Mountain area, then the introduction of the iron from the Tomstown Dolomite into the Antietam Quartzite had to occur before or while the Triassic rocks were deposited. This would mean that the liesegang rings are at least Triassic in age.

Another line of reasoning suggests that the introduction of the iron may have been associated with the deep weathering of the Antietam Quartzite during the Middle Tertiary of later (See Clark, this guidebook). Could there have been Tomstown Dolomite overlying the Antietam Quartzite in the Middle Tertiary? Erosion rates for this area shed some light on the question. Sevon (1989) calculated the present erosion rate for the Juniata River drainage basin as 27 m/Ma based on suspended and dissolved loads in the river. From previous studies, Sevon (1989) compiled a mean erosion rate for limestone in the eastern United States of 32 m/Ma. Using a value of 30 m/Ma and assuming erosion rates have not changed since the Middle Tertiary (15 Ma), the net amount of denudation since this time is roughly 450 m. If we project the Tomstown up 450 m, it would overlie the Antietam where the quarry is currently located. In fact the current elevation difference between the Tomstown in the Cumberland Valley and the Antietam at the top of the quarry is only 230 m. The problem with this kind of exercise is that we know that erosion rates changed radically in the past 15 Ma as the South Mountain area was exposed to glacial and interglacial climates (Braun, 1989). This casts doubt on our ability to predict the past position of the Tomstown Dolomite relative to the Antietam Quartzite, but leaves open the possibility that its removal is a relatively recent event.

Are there other sources for the iron other than the Tomstown Dolomite? The Antietam Quartzite in the Mt. Holly area contains iron-rich heavy minerals such as magnetite (Freedman, 1967). The weathering of these may have been the source of iron in the liesegang rings. The amount of iron-rich heavy minerals in the Antietam Quartzite at Mt. Holly is unknown. This may be relatively unimportant because very little iron is required to stain a rock with liesegang rings (Pettijohn, 1975). The liesegang rings may have formed in association with the deep weathering in the Antietam Quartzite itself. This may have occurred long after the Tomstown eroded away.

QUARRY OPERATIONS

The quarry is owned and operated by Pennsy Supply, Inc. of Harrisburg. The quarry produces sand, washed-sand, and fine-aggregate products for commercial use. The primary product is a concrete sand; other products are masonry sand, aggregate for concrete and asphalt, and general purpose sand and aggregate. There are two plants at the site for producing the range of products. The original plant that is nearest to the entrance to the quarry (Figure 75) has been upgraded and produces mainly masonry sand and sand approved by the Department of Environmental Resources (DER) for sand mounds for septic systems. This plant is a dry plant and consists of a primary jaw crusher, a secondary gyratory crusher, two sets of triple deck screens at 1 1/2 in., 3/8 in., and 5 mesh (4mm), a BARMAC impact crusher, and storage bins for loading directly into trucks or onto stackers for storage in stockpiles. The screens can be heated so that operation can continue during winter months.

The new and larger plant is both a dry and a wet plant sited to the west of the original plant (Figure 75). The dry part of the new plant consists of a primary jaw crusher, a secondary gyratory crusher, a BARMAC impact crusher, and triple deck screens set at 1 1/2, 1/2, and 3/8 inches. Material at minus 3/8 in. from the dry plant screens is fed to the wet plant that contains a triple deck wet screen set at 1/2, 1/4, and 3/16 in., an out-of-specification sand classifying tank, two screw classifiers, a cyclone, and a JADAIR clarifier. The main product from the wet plant is clean concrete sand, washed masonry sand, and clean fine Type A aggregate for asphalt that has a SRL rating of "E". The wet plant can be bypassed during cold weather, if necessary, to produce DER septic system sand, general purpose sand, and Type A aggregate. Products are conveyed to stockpiles by stackers around the plant and are loaded into trucks with front-end loaders.

The wet plant uses about 1,600 gallons of water per minute, which is supplied from a well that produces only about 100 gallons per hour. A well-designed recycling and storage program for plant water conserves almost all of the wash water. The JADAIR clarifier uses a flocculent to settle out the fine material in suspension that is fed from the classifier, and the clear water is recycled back through the plant. The flocculent is a polymer that neutralizes the naturally repelling cationic charges of the fine particles, which are then attracted to each other and flocculate and settle out of suspension (J. Kriz, personal communication, 1991). The slimes thus created are pumped from the clarifier as a slurry to settling ponds.

The quarry produces about 400,000 tons of product per year, about two-thirds of which is concrete sand. About 15 to 20 percent of the material quarried is lost as slimes and clay.

		LEAVE STOP 11. Return on Mountain View Road the same same way came.
0.7	55.8	**STOP SIGN. TURN LEFT THEN IMMEDIATELY TURN RIGHT AT THE STOP SIGN** following Mountain View Road.

0.8	56.6	**STOP SIGN. TURN RIGHT** onto Pine Road.
1.4	58.0	**STOP SIGN. TURN LEFT** onto PA Route 34 N.
3.8	61.8	**TURN RIGHT** onto Interstate 81 N.
5.0	66.8	**EXIT** at Exit 17 to US Route 11 N, New Kingston.
0.3	67.1	**YIELD SIGN. BEAR RIGHT** onto US Route 11 N.
0.3	67.4	**TURN RIGHT** to The Embers Inn.

END OF DAY 2 AND THE 56th ANNUAL FIELD CONFERENCE!!!

DRIVE HOME SAFELY!!

REFERENCES CITED

Adams, G. F., ed., 1975, Planation surfaces. peneplains, pediplains, and etchplains: Harrisburg, Dowden, Hutchinson and Ross, Inc., 476 p.

Aleinikoff, J. N., Zartman, R. E., Rankin, D. W., Lyttle, P. T., Burton, W. C., and McDowell, R. C., 1991, New U-Pb zircon ages for rhyolite of the Catoctin and Mount Rogers Formations-More evidence for two pulses of Iapetan rifting in the central and southern Appalachians (abs.): Geological Society of America Abstracts with Programs, v. 23, no. 1, p. 2.

Allen, J. R. L., 1982, Sedimentary structures, their character and physical basis: New York, Elsevier Publishing Co., 1256 p.

Amsden, T. W. 1951, Paleontology of Washington County: Maryland Department of Geology and Mines, Water Resources Bulletin, No. 10, p. 98-123.

Anonymous, We. R., 1985, Gross Minerals, a force in the fine phyllite fillers field, Pennsylvania Geology, v. 16, no. 6, p. 2-5.

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.

Badger, R. L. and Sinha, A. K., 1991, Nature of late Precambrian mafic magmatism within the Appalachian orogen (abs.): Geological Society of America Abstracts with Programs, v. 23, no. 1, p. 5.

Barrell, J., 1925, The nature and environment of the Lower Cambrian sediment of the southern Appalachians: American Journal of Science, v. 9, p. 1-20.

Barron, E. J., 1989, Climate variations and the Appalachians from the Late Paleozoic to the Present: results from model simulations, in Gardner, T. W., and Sevon, W. D., eds., Appalachian Geomorphology (reprinted from Geomorphology, v. 2, no. 1-3): Amsterdam, Elsevier Science Publishers, p. 99-118.

Bascom, F., 1893, The structures, origin, and nomenclature of the acid volcanic rocks of South Mountain, Pennsylvania: Journal of Geology, v. 1, p. 813-832.

Bassler, R. S., 1919, Cambrian and Ordovician: Maryland Geological Survey, Johns Hopkins University Press, Baltimore, 408 p.

Baulig, H., 1952, Surfaces d'Aplanissement: Annales de Géographie, v. 61, p. 161-183; 245-262.

Becher, A. E. and Root, S. I., 1981, Groundwater and geology of the Cumberland Valley, Cumberland County, Pennsylvania: Pennsylvania Geological Survey, 4th ser., Water Resource Report 50, 95 p.

Becher, A. E. and Taylor, L. E., 1982, Groundwater resources in the Cumberland and contiguous valleys of Franklin County, Pennsylvania: Pennsylvania Geological Survey, 4th ser., Water Resource Report W-53, 67 p.

Berg, T. M., 1975, Geology and mineral resources of the Brodheadsville quadrangle, Monroe and Carbon Counties, Pennsylvania: Pennsylvania Geological Survey, 4th ser., Atlas 205a, 60 p.

Berg, T. M., and others, 1980, Geologic Map of Pennsylvania: Pennsylvania Geological Survey, 4th ser., Map 1, scale 1:250,000.

Berg, T. M., Sevon, W. D., and Abel, R., 1984, Rock types of Pennsylvania: Pennsylvania Geological Survey, 4th ser., Map 63, scale 1:500,000.

Berkheiser, S. W., Jr., 1985, High-purity silica occurrences in Pennsylvania: Pennsylvania Geological Survey, 4th ser., Mineral Resource Report 88, 67 p.

Berkheiser, S. W., Jr., Barnes, J. H., and Smith, R. C., II, 1985, Directory of the nonfuel-mineral producers in Pennsylvania: Pennsylvania Geological Survey, 4th ser., Information Circular 54, 4th ed., 165 p.

Berkheiser, S. W., Jr., Potter, N., Jr., and Sevon, W. D., 1982, Hempt Brothers clay pit at Toland: in Potter, N., Jr., ed., South Mountain: Guidebook, Harrisburg Area Geological Society, p. 10-14.

Berkheiser, S. W., Jr. and Smith, R. C., II, 1990, Overhead protection, Pennsylvania's mineral industry contribution: Pennsylvania Geology, v. 21, no. 3, p. 2-6.

Bining, A. C., 1979, Pennsylvania iron manufacture in the eighteenth century, 2nd ed.: Harrisburg, Pennsylvania Historical and Museum Commission, 215 p.

Bjornerud, M. G., 1991, Conveying principles of finite strain with standard graphics software: Journal of Geological Education, v. 39, p. 23-27.

Black, R. F., 1976a, Periglacial features indicative of permafrost: ice and soil wedges: Quaternary Research, v. 6, p. 3-26.

Black, R. F., 1976b, Features indicative of permafrost: Annual Review of Earth and Planetary Sciences, v. 4, p. 75-94.

Black, R. F., 1983, Pseudo-ice-wedge casts of Connecticut, northeastern United States: Quaternary Research, v. 20, p. 74-89.

Blatt, H., 1980, Origin of sedimentary rocks, 2nd ed.: Englewood Cliffs, NJ, Prentice Hall, 782 p.

Bloomer, R. O., 1950, Late Pre-Cambrian and Lower Cambrian formations in central Virginia: American Journal of Science, v. 248, p. 753-783.

Bloomer, R. O. and Bloomer, R. R., 1947, The Catoctin Formation in central Virginia: Journal of Geology, v. 55, p. 94-106.

Bloomer, R. O. and Werner, H. J., 1955, Geology of the Blue Ridge in central Virginia: Geological Society of America Bulletin, v. 65, p. 579-606.

Bond, G. C., Nickeson, P. A., and Kominz, M. A., 1984, Breakup of a supercontinent between 625 Ma and 555 Ma: New evidence and implications for continental histories: Earth and Planetary Science Letters, v. 70, p. 325-345.

Bowring, C. and Spencer, E., 1987, Catoctin pillow lavas (abs.): Geological Society of America Abstracts with Programs, v. 19, no. 2, p. 77.

Bradt, P. T. and Wimpfheimer, L., 1987, Survey of Pocono lakes to evaluate sensitivity to acidification: Environmental Studies Center, Lehigh University, Bethlehem, PA.

Braun, D. D., 1989a, A revised Pleistocene glaciation sequence in eastern Pennsylvania: Support for limited Early Wisconsin ice and a single Late Illinoian advance beyond the Late Wisconsin border (abs.): 28th International Geological Congress Abstracts, v. 1: p. 1-196-1-197.

Braun, D. D., 1989b, Glacial and periglacial erosion of the Appalachians, in Gardner, T. W. and Sevon, W. D., eds., Appalachian Geomorphology (reprinted from Geomorphology, v. 2, no. 1-3): Amsterdam, Elsevier Science Publishers, p. 233-256.

Braun, D. D., and Inners, J. D., 1990, Weathering of conglomerate ledges and tors within the glacial limit in northeastern Pennsylvania: Evidence for single stage tor development under Pleistocene periglacial conditions (abs.): Geological Society of America Abstracts with Programs, v. 22, no. 2, p. 6.

Braun, E. L., 1950, Deciduous Forests of Eastern North America: Philadelphia, Blackiston Company, 596 p.

Brezinski, D. K., 1991, Lithostratigraphy of the Lower Cambrian Tomstown Formation (abs.): Geological Society of America Abstracts with Programs, v. 23, no. 1, p. 10.

Brown, W. R., 1970, Investigations of the sedimentary record in the Piedmont and Blue Ridge of Virginia, in Fischer, G. W., Pettijohn, F. J., Reed, J. C., Jr., and Weaver, K. N., eds., Studies of Appalachian Geology: Central and Southern: New York, Interscience, p. 335-349.

Bryan, K., Cleaves, A. B., and Smith, H. T. U., 1932/33, The present status of the Appalachian problem: Zeitschrift für Geomorphologie, v. 7, p. 312-320.

Bryan, K., 1946, Cryopedology the study of frozen ground and intensive frost-action with suggestions on nomenclature: American Journal of Science, v. 244, p. 622-642.

Büdel, J., 1982, Climatic Geomorphology: Princeton, Princeton University Press, 443 p.

Camuto, C., 1991, Dropping acid in the Southern Appalachians: A wild trout resource at considerable risk: Trout, Winter, 1991, p. 16-39.

Carter, B. J., and Ciolkosz, E. J., 1980, Soil temperature regimes of the Central Appalachians: Soil Science Society of America Journal, v. 44, p. 1052-1058.

Carter, B. J., and Ciolkosz, E. J., 1986, Sorting and thickness of waste mantle material in a sandstone spur in central Pennsylvania: Catena, v. 13, p. 241-256.

Cecil, C. B., 1990, Paleoclimatic controls on stratigraphic repetition of chemical and siliclastic rocks: Geology, v. 19, p. 533-536.

Ciolkosz, E. J., Carter, B. J., Hoover, M. T., Cronce, R. C., Waltman, W. J., and Dobos, R. R., 1990, Genesis of soils and landscapes in the Ridge and Valley province of central Pennsylvania, in Knuepfer, P. L. K. and McFadden, L. D., eds., Soils and landscape evolution: Amsterdam, Elsevier, p. 245-261.

Ciolkosz, E. J., Cronce, R. C., and Sevon, W. D., 1986, Periglacial features in Pennsylvania: University Park, PA, The Pennsylvania State University, Agronomy Department, Agronomy Series Number 92, 15 p.

Clark, G. M., 1989a, Central and southern Appalachian accordant ridge crest elevations south of glacial border: Regional erosional remnants or incipient periglacial surfaces? (abs.): 28th International Geological Congress Abstracts, v. 1, p. 1-299-1-300.

Clark, G. M., 1989b, Central and southern Appalachian water and wind gap origins: Review and new data, in Gardner T. W. and Sevon, W. D., eds., Appalachian Geomorphology (reprinted from Geomorphology, v. 2, no. 1-3): Amsterdam, Elsevier Science Publishers, p. 209-232.

Clark, G. M., and Ciolkosz, E. J., 1988, Periglacial geomorphology of the Appalachian Highlands and Interior Highlands south of the glacial border: Geomorphology, v. 1, p. 191-220.

Clark, G. M., Ciolkosz, E. J., Kite, J. S., and Lietzke, D. A., 1989, Central and Southern Appalachian Geomorphology - Tennessee, Virginia, and West Virginia: 28th International Geological Congress Field Trip Guidebook T150, 105 p.

Clark, G. M., and Hedges, J., in press, Origin of certain high-elevation local broad uplands in the Central Appalachians south of the glacial border - a paleoperiglacial hypothesis: 21st Annual Binghamton Geomorphology Symposium, Chichester, Wiley, in press.

Clark, M. J., ed., 1988, Advances in periglacial geomorphology: New York, John Wiley and Sons, 481 p.

Clarke, M. S., 1968, Pioneer iron works: Philadelphia, Chilton Book Company, 80 p.

Cleaves, E. T., Edwards, J. E., Jr., and Glaser, J. D., 1968, Geologic map of Maryland: Maryland Geological Survey, scale 1:250,000.

Cloos, E., 1947, Oolite deformation in the South Mountain fold, Maryland: Geological Society of America Bulletin, v. 58, p. 843-918.

Cloos, E., 1951, Physical features of Washington County: Maryland Department of Geology and Mines, Water Resources Bulletin 10, p. 17-94.

Cloos, E., 1958, Structural geology of South Mountain and Appalachians in Maryland: 23rd Annual Field Conference of Pennsylvania Geologists, Hagerstown, MD, Guidebook, 85 p.

Cloos, E. and others, 1968, The geology of mineral deposits in south-central Pennsylvania: 33rd Annual Field Conference of Pennsylvania Geologists, Harrisburg, PA, Guidebook, 80 p.

Cloos, E., 1971, Microtectonics along the western edge of the Blue Ridge, Maryland and Virginia: Johns Hopkins University, Studies in Geology No. 20, 234 p.

Cosby, B. J., Hornberger, G. M., Wright, R. F., and Galloway, J. N., 1986, Modeling the effects of acid deposition: control of long-term sulfate dynamics by soil sulfate adsorption: Water Resources Research, v. 22, p. 1283-1291.

Crimes, T. P., 1970, The significance of trace fossils in sedimentology, stratigraphy and palaeoecology with examples from Lower Palaeozoic strata, in Crimes, T. P. and Harper, J. C., eds., Trace Fossils: Geological Journal Special Issue 3, p. 106-126.

Daly, R. A., 1905, The accordance of summit levels among Alpine mountains: the fact and its significance: Journal of Geology, v. 13, p. 105-125.

Davis, W. M., 1889, The rivers and valleys of Pennsylvania: National Geographic Magazine, v. 1, p. 183-253.

Delcourt, P. A. and Delcourt, H. R., 1981, Vegetation maps for eastern North America; 40,000 yr BP to the present, in Romans, R. C., ed., Geobotany II: New York, Plenum, p. 123-165.

Delcourt, P. A. and Delcourt, H. R., 1984, Late Quaternary paleoclimates and biotic responses in eastern North America and the western North Atlantic Ocean: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 48: p. 263-284.

Demek, J., 1969, Cryoplanation terraces, their geographical distribution, genesis and development: Rozpravy Československé Akademie Věd. Rada Matematických A Přírodnich Věd, v. 79, no. 4, 80 p., 16 photos.

Demek, J., 1972, Manual of Detailed Geomorphological Mapping: Prague, Academia, 344 p.

d'Invilliers, E. V., 1887, Report on the iron ore mines and limestone quarries of Cumberland-Lebanon Valley, in Annual report of the Geological Survey for 1886: Pennsylvania Geological Survey, 2nd ser., Annual Report 1886, Pt. IV, p. 1411-1567.

Dockter, G. N., 1990, The Catoctin metabasalts: evidence of a rifting environment: unpublished independent research paper, Dickinson College, Carlisle, PA, 27 p.

Duffy, D. F. and Wittecar, G. R., 1991, Geomorphic development of segmented alluvial fans in the Shenandoah Valley, Stuarts Draft, Virginia (abs.): Geological Society of America Abstracts with Programs, v. 23, no. 1, p. 24.

Edwards, J., Jr., compiler, 1978, Geologic map of Washington County: Maryland Geological Survey, scale 1:62,500.

Ehlen, J., 1990, Geomorphic, petrographic and structural classification of granite landforms using spatial patterns (abs.): Geological Society of America Abstracts With Programs, v. 22, no. 7, p. A21.

Espenshade, G. H., 1986, Geology of the Marshall quadrangle, Faquier County, Virginia: U. S. Geological Survey Bulletin 1560, 59 p.

Faill, R. T. and MacLachlan, D. B., 1989, Tectonic terranes of southeastern Pennsylvania (abs.): Geological Society of America Abstracts with Programs, v. 21, no. 1, p. 13.

Farlekas, G., 1961, The geology of part of South Mountain of the Blue Ridge province north of the Pennsylvania-Maryland border: unpublished M.S. thesis, University Park, The Pennsylvania State University, 64 p.

Fauth, J. L., 1968, Geology of the Caledonia Park quadrangle Area, South Mountain, Pennsylvania: Pennsylvania Geological Survey, 4th ser., Atlas 129a, 133 p.

Fauth, J. L., 1977, Geologic map of the Catoctin Furnace and Blue Ridge Summit Quadrangles, Maryland: Maryland Geological Survey, scale 1:24,000.

Fauth, J. L., 1978, Geology and mineral resources of the Iron Springs area, Adams and Franklin Counties, Pennsylvania: Pennsylvania Geological Survey, 4th ser., Atlas 129c, 72 p.

Fauth, J. L., 1981, Geologic map of the Myersville Quadrangle, Maryland: Maryland Geological Survey, scale 1:24,000.

Fedonkin, M. A., 1981, Palaeoichnology of the Precambrian-Cambrian transition, in Taylor, M. E., ed., Short papers from the Second International Symposium on the Cambrian System: U.S. Geological Survey Open-File Report 81-743, p. 89-90.

Fenneman, N. M., 1938, Physiography of eastern United States: New York, McGraw-Hill Book Company, 714 p.

Flemal, R. C., 1971, The attack on the Davisian system of geomorphology: a symposium: Journal of Geologic Education, v. 19, p. 3-13.

Flinn, D. 1962, On folding during three-dimensional progressive deformation: Geological Society of London Quarterly Journal, v. 118, p. 385-433.

Flippo, H. N., Jr., 1974, Springs of Pennsylvania: Pennsylvania Department of Environmental Resources, Water Resources Bulletin 10, 46 p.

Flower, L. E., 1975, History of Pine Grove Furnace: Carlisle, PA, Cumberland County Historical Society, 21 p.

Foose, R. M., 1945a, Iron-manganese ore deposits of White Rocks, Cumberland County: Pennsylvania Geological Survey, 4th ser., Mineral Resource Report 26, 35 p.

Foose, R. M., 1945b, Manganese minerals of Pennsylvania: Pennsylvania Geological Survey, 4th ser., Mineral Resource Report 27, 130 p.

Foose, R. M. and others, 1948, Excursion 1, South Mountain, in Cathcart, S. H., ed., 14th Annual Field

Conference of Pennsylvania Geologists, Harrisburg, PA, Guidebook, p. 1-6.
Frakes, L. A., 1979, Climates throughout geologic time: Amsterdam, Elsevier Science Publishers, 310 p.
Freedman, J., 1967, Geology of a portion of the Mount Holly Springs Quadrangle, Adams and Cumberland Counties, Pennsylvania: Pennsylvania Geological Survey, 4th ser., Progress Report 169, 66 p.
Freedman, J., 1968, Bender's Quarry - Mt. Holly Springs, Pennsylvania, in The geology of mineral deposits in south-central Pennsylvania: 33rd Annual Field Conference of Pennsylvania Geologists, Harrisburg, PA, Guidebook, p. 28-37.
French, H. M., 1976, The periglacial environment: London, Longman, 309 p.
French, H. M. and Karte, J., 1988, A periglacial overview, in Clark, M. J., ed., Advances in Periglacial Geomorphology: New York, John Wiley and Sons, p. 463-473.
Friedman, G. M. and Sanders, J. E., 1978, Principles of Sedimentology: New York, John Wiley and Sons, 792 p.
Fritz, W. H., 1972, Lower Cambrian trilobites from the Sekawi Formation type section, Mackenzie Mountains, north-western Canada: Geological Survey of Canada Bulletin, v. 212, p. 1-58.
Fritz, W. H. and Crimes, T. P., 1985, Lithology, trace fossils, and correlation of the Precambrian-Cambrian boundary beds, Cassiar Mountains, north-central British Columbia: Geological Survey of Canada Paper 83-13, p. 1-24.
Fulton, R. J., 1989, ed., Quaternary Geology of Canada and Greenland: Geological Society of America, Volume K-1, Geological Survey of Canada, no. 1, 839 p.
Gagen, C. J. and Sharpe, W. E., 1987, Net sodium loss and mortality of three salmonid species exposed to a stream acidified by atmospheric deposition: Environmental Contamination and Toxicology Bulletin, v. 39, p. 7-14.
Gannett Flemming Corddry and Carpenter, Inc., 1945, Report on water treatment project for the Borough of Chambersburg, PA: Harrisburg, PA.
Gardiner, V. and Dackombe, R., 1983, Geomorphological Field Manual: London, George Allen and Unwin, 254 p.
Gardner, T. W. and Sevon, W. D., eds., 1989, Appalachian Geomorphology: Amsterdam, Elsevier, 318 p.
Geiser, P. A. and Engelder, T., 1983, The distribution of layer-parallel shortening fabrics in the Appalachian foreland of New York and Pennsylvania: Evidence for two non-coaxial phases of the Alleghanian orogeny, in Hatcher, R. D., Jr., Williams, H., and Zeitz, I., eds., Contributions to the tectonics and geophysics of mountain chains: Geological Society of America Memoir 158, p. 161-175.
Gerhart, J. M. and Lazorchick, G. J., 1988, Evaluation of the ground-water resources of the lower Susquehanna River Basin, Pennsylvania and Maryland: U.S. Geological Survey Water-Supply Paper 2284, 128 p.
Geyer, A. R., Smith, R. C., II, and Barnes, J. H., 1976, Mineral collecting in Pennsylvania: Pennsylvania Geological Survey, 4th ser., General Geology Report 33, 260 p.
Godfrey, A. E., 1975, Chemical and physical erosion in the South Mountain anticlinorium, Maryland: Maryland Geological Survey, Information Circular 19, 35 p.
Godfrey, A. E. and Cleaves, E. T., 1991, Landscape analysis: Theoretical considerations and practical needs: Environmental Geology and Water Sciences, v. 17, no. 2, p. 141-155.
Goldthwait, R. P., 1976, Frost sorted patterned ground: A review: Quaternary Research, v. 6, p. 27-35.
Gordon, S. E., 1922, The mineralogy of Pennsylvania: Philadelphioa, Academy of Natural Sciences Special Publication 1, 158 p.
Grimvall, A., Cole, C. A., Allard, B., and Sanden, P., 1986, Quality trends of public water supplies in Sweden: Water Quality Bulletin, v. 11, no. 4, p. 6-11.
Hack, J. T., 1960, Interpretation of erosional topography in humid temperate regions: American Journal of Science Bradley Volume, v. 258-A, p. 80-97.
Hack, J. T., 1965, Geomorphology of the Shenandoah Valley Virginia and West Virginia and origin of the residual ore deposits: U.S. Geological Survey Professional Paper 484, 84 p.
Hack, J. T., 1975, Dynamic equilibrium and landscape evolution, in Melhorn, W. N. and Flemal, R. C., eds., Theories of landform development: London, George Allen and Unwin, p. 87-102.
Haines, T. A., 1981, Acidic precipitation and its consequences for aquatic ecosystems: a review: Transactions of the American Fisheries Society, v. 110, p. 669-707.
Hatcher, R. D., Jr., 1990, Structural geology: principles, concepts, and problems: Columbus, OH, Merrill Publishing Co., 531 p.
Hatcher, R. D., Thomas, W. A., and Viele, G. W., 1989, The Appalachian-Ouachita Orogen in the United States: The Geology of North America, v. F-2: Geological Society of America, 767 p.
Hedges, J., 1969, Opferkessel: Zeitschrift für Geomorphologie, v. 13, p. 22-55.
Hedges, J., 1975, Multiple cycles of cryoplanation on Sugarloaf Mountain, Maryland: Biuletyn Peryglacjalny, v. 24, p. 233-243.
Heron, S. D., Jr., Judd, J. B., and Johnson, H. S., Jr., 1971, Clastic dikes associated with soil horizons in the North and South Carolina Coastal Plain: Geological Society of America Bulletin, v. 82, p. 1801-1810.
Hofmann, A. W., 1988, Chemical differentiation of the Earth: the relationship between mantle, continental crust, and oceanic crust: Earth and Planetary Science Letters, v. 90, p. 297-314.
Hosterman, J. W., 1968, White clay deposits near Mt. Holly Springs, Cumberland County, Pennsylvania, in The geology of mineral deposits in south-central Pennsylvania: 33rd Annual Field Conference of Pennsylvania Geologists, Harrisburg, PA, Guidebook, p. 38-51.

Hosterman, J. W., 1969, White clay deposits near Mt. Holly Springs, Cumberland County, Pennsylvania: U.S. Geological Survey Professional Paper 650B, p. 66-72.
Hupp, C. R., 1983, Geo-botanical evidence of Late Quaternary mass wasting in block field areas of Virginia: Earth Surface Processes and Landforms, v. 8, p. 439-450.
Jacobson Jr., G. L., Webb III, T., and Grimm, E. C., 1987, Patterns and rates of vegetation change during the deglaciation of eastern North America, in Ruddiman W. F. and Wright, H. E., Jr., eds., North America and adjacent oceans during the last deglaciation: Geological Society of America, p. 277-288.
Johnson, F., 1986, Acid precipitation: Pennsylvania Fish Commission, PFC 1003 7/86 10 M.
Karrasch, H., 1974, Hangglättung und Kryoplanation an Beispielen aus den Alpen und kanadischen Rocky Mountains: Akademie Der Wissenschaften, Göttingen. Mathematische-Physikale Abhandlungen, v. 29, p. 287-300.
Karte, J., 1982, Development and present state of German periglacial research in arctic and alpine environments: Biuletyn Peryglacjalny, v. 29, p. 183-201.
Kauffman, M. E. and Frey, E. P., 1979, Antietam sandstone ridges - exhumed barrier islands or fault-bounded blocks? (abs.): Geological Society of America Abstracts with Programs, v. 11, no. 1, p. 18.
Kaufmann, P. R., Herlihy, A. T., Elwood, J. W., Mitch, M. E., Overton, W. S., Sale, M. J., Messer, J. J., Cougan, K. A., Peck, D. V., Reckhow, K. H., Kinney, A. J., Christie, S. J., Brown, D. D., Hagley, C. A., and Jager, H. I., 1988, Chemical characteristics of streams in the mid-Atlantic and southeastern United States (Results of the National Stream Survey Phase I), in Vol. 1: Population descriptions and physico-chemical relationships: U.S. Environmental Protection agency, Washington, D.C., EPA/600/3-88/021a.
Kay, M., 1951, North American Geosynclines: Geological Society of America Memoir 48, 143 p.
Kemper, J., III, 1968?, American charcoal making: National Park Service, U.S. Department of the Interior, Eastern National Park and Monument Association, 25 p.
Kimmel, W. G., Murphy, D. J., Sharpe, W. E., and DeWalle, D. R., 1985, Macroinvertebrate community structure and detritus processing rates in two southwestern Pennsylvania streams acidified by atmospheric deposition: Hydrobiologia, v. 124, p. 97-102.
King, P. B., 1949, The base of the Cambrian in the southern Appalachians: American Journal of Science, v. 247, p. 513-530; 622-645.
King, P. B., 1950, Geology of the Elktin area, Virginia: U.S. Geological Survey Professional Paper 230, 82 p.
King, P. B. and Fergusen, H. W., 1960, Geology of northeasternmost Tennessee: U.S. Geological Survey Professional Paper 311, 136 p.
Kline, S. W., Conley, J. F., and Evans, N., 1987, The Catoctin Formation in the eastern Blue Ridge of Virginia: Evidence for submarine volcanism (abs.): Geological Society of America Abstracts with Programs, v. 19, no. 2, p. 93.
Kochel, R. C., 1990, Humid fans of the Appalachian Mountains, in Rachocki, A. H. and Church, M., eds., Alluvial fans: New York, John Wiley and Sons, p. 109-129.
Kulander, B. R. and Dean, S. L., 1986, Structure and tectonics of the central and southern Appalachian Valley and Ridge Plateau Provinces, West Virginia, and Virginia: American Association of Petroleum Geologists Bulletin, v. 70, p. 1674-1684.
Kurjack, D. C., 1958, Hopewell Village national historic site: Washington, D.C., National Park Service Historical Handbook Series No. 8, 44 p.
Lachenbruch, A. H., 1962, Mechanics of thermal contraction cracks and ice-wedge polygons in permafrost: Geological Society of America Special Paper 70, 69 p.
Laurence, R. A. and Palmer, A. R., 1963, Age of the Murray Shale and Hesse Quartzite on Chilhowee Mountain, Blount County, Tennessee: U.S. Geological Survey Professional Paper 475C, p. 53-54.
Lauriol, B., 1990, Canadian landform examples 18: Cryoplanation terraces, northern Yukon: Le Geographe canadien, v. 34, no. 4, p. 347-351.
Leffingwell, E. de K., 1915, Ground-ice wedges. The dominant form of ground-ice on the north coast of Alaska: Journal of Geology, v. 23, p. 635-654.
Leffler, R. J., 1981, Estimating average temperatures on Appalachian summits: Journal of Applied Meteorology, v. 20, p. 637-642.
Lesley, J. P., 1892, A summary description of the geology of Pennsylvania: Pennsylvania Geological Survey, 2nd ser., v. 1, 719 p.
Liestøl, O., 1961, Talus taluses in Arctic regions: Norsk Polarinstitutt Årbok, p. 102-105.
Lisle, R. J., 1985, Geological strain analysis, a manual for the Rf/f technique: New York, Pergamon Press, 99 p.
Lozinski, M. W., 1909, Über die mechanische Verwitterung der Sandsteine im gemässigten Klima: Bulletin International Akademie Sciences Crakovie, Classe Scientific Mathematische Naturelles, v. 1, p. 1-25.
Lynch, J. A., Corbett, E. S., and Kostelnik, K. M., 1986, Atmospheric deposition: spatial and temporal variation in Pennsylvania, 1985: Institute for Research on Land and Water Resources, The Pennsylvania State University, University Park, PA.
Lynch, J. A., Corbett, E. S., and Kostelnik, K. M., 1987, Atmospheric deposition: spatial and temporal variation in Pennsylvania, 1986: Institute for Research on Land and Water Resources, The Pennsylvania

State University, University Park, PA.
M'Cauley, I. H., 1878, Historical sketch of Franklin County, Pennsylvania: Chambersburg, PA, D. F. Pursel, 322 p.
Mackay, J. R., 1974, Ice-wedge cracks, Garry Island, Northwest Territories: Canadian Journal of Earth Science, v. 11, p. 1366-1383.
Mackay, J. R., 1990, Some observations on the growth and deformation of epigenetic, syngenetic, and anti-syngenetic ice wedges: Permafrost and Periglacial Processes, v. 1, p. 15-29.
MacLachlan, D. B. and Root, S. I., 1966, Comparative tectonics of the Cumberland and Lebanon Valleys: 31st Annual Field Conference of Pennsylvania Geologists, Harrisburg, PA, Guidebook, 90 p.
McDonald, M. E., 1985, Acid deposition and drinking water: Environmental Science and Technology, v. 19, p. 772-776.
McDonald, M. G. and Harbaugh, A. W., 1988, A modular three-dimensional finite-difference ground-water flow model: U.S. Geological Survey Techniques of Water Resources Investigations, book 6, chap. A1.
Middlekauff, B. D., 1987, Relict periglacial morphosequences in the northern Blue Ridge: unpublished Ph.D. thesis, East Lansing, Michigan State University, 172 p.
Mills, H. H., 1981, Boulder deposits and the retreat of mountain slopes, or, "gully gravure" revisited: Journal of Geology, v. 89, p. 649-660.
Mills, H. H., Brakenridge, G. R., Jacobson, R. B., Newell, W. L., Pavich, M. J., and Pomeroy, J. S., 1987, Appalachian Mountains and Plateaus: Geological Society of America, Centennial Volume 2, p. 5-50.
Mills, H. H. and Delcourt, P. A., in press, Appalachian highlands and interior low plateaus, in Morrison, R. B., ed., Quaternary non-glacial geology of the conterminous United States: Geological Society of America, Volume K-2, in press.
Monmonier, M. S., 1967, Upland accordance in the Ridge and Valley section of Pennsylvania: unpublished M.S. thesis, University Park, The Pennsylvania State University, 58 p.
Monmonier, M. S., 1968, Trends in upland accordance in Pennsylvania's Ridge and Valley section: Pennsylvania Academy of Science Proceedings, v. 42, p. 157-162.
Monmonier, M. S., 1971, Upland adjustment to regional drainage in central Pennsylvania: an application of trend surface analysis: Journal of Geography, v. 70, p. 360-370.
Mose, D. G., Diecchio, R. J., DiGuiseppi, W. H., and Nagel, M. S., 1985, Confirmation of a latest Precambrian (@600 m.y.) age for the Catoctin Formation in Virginia (abs.): Geological Society of America Abstracts with Programs, v. 17, no. 2, p. 126.
Moss, J. H., 1976, Periglacial origin of extensive lobate colluvial deposits on the south flank of Blue Mountain near Shartlesville and Strausstown, Berks County, Pennsylvania: Pennsylvania Academy of Science Proceedings, v. 50, p. 42-44.
Nagel, M. S. and Mose, D. G., 1984, A revised geochemical and geochronological picture of the Catoctin Formation (abs.): Geological Society of America Abstracts with Programs, v. 16, no. 3, p. 182.
National Atmospheric Deposition Program, 1990, NADP/NTN annual data summary, precipitaion chemistry in the United States, 1989.
Neal, J. T., Langer, A. M., and Kerr, P. F., 1968, Giant desiccation polygons of Great Basin playas: Geological Society of America Bulletin, v. 79, p. 69-90.
Nelson, A., 1968, GAF Corporation quarries--Charmian, Pennsylvania, in Cloos, E. and others, The geology of mineral deposits in south-central Pennsylvania, Annual Field Conference of Pennsylvania Geologists, 33rd, Harrisburg, PA, Guidebook, p. 6-10.
Nickelsen, R. P., 1956, Geology of the Blue Ridge near Harpers Ferry, West Virginia: Geological Society of America Bulletin, v. 67, p. 239-270.
Nickelsen, R. P., 1979, Sequence of structural stages of the Allegheny orogeny at the Bear Valley strip mine, Shamokin, Pennsylvania: American Journal of Science, v. 279, p. 225-271.
Nutter, L. J., 1973, Hydrogeology of the carbonate rocks, Frederick and Hagerstown Valleys Maryland: Maryland Geological Survey, Report of Investigations No. 19, 70 p.
Olson, C. G., 1989, Mountain soils of eastern Appalachians (abs.): 28th International Geological Congress Abstracts. v. 2, p. 2-548.
Palmer, A. R., 1981, Subdivision of the Sauk Sequence, in Taylor, M. E., ed., Short Papers from the Second International Symposium on the Cambrian System: U.S. Geological Survey Open-File Report 81-743, p. 160-161.
Palmer, A. R., 1983, The Decade of North American Geology, 1983 geologic time scale: Geology, v. 11, p. 503-504.
Patterson, J. G. and Simpson, E. L., 1991, Paleoenvironmental constraints on the Chilhowee Group of the eastern Blue Ridge: Thoroughfare Gap, Virginia (abs.): Geological Society of America Abstracts with Programs, v. 23, no. 1, p. 113.
Pavich, M. J., 1989, Regolith residence time and the concept of surface age of the Piedmont "peneplain", in Gardner, T. W. and Sevon, W. D., eds., Appalachian Geomorphology (reprinted from Geomorphology, v. 2, no. 1-3): Amsterdam, Elsevier Science Publishers, p. 181-196.
Pearce, J. A. and Cann, J. R., 1973, Tectonic setting of basic volcanic rocks determined using trace element analysis: Earth and Planetary Science Letters, v. 19, p. 290-300.

Pearse, J. B., 1970, A concise history of the iron manufacture of the American colonies up to the Revolution and of Pennsylvania up to the present time: New York, Burt Franklin, Research and Source Works Series 497, Selected essays in history, economics, and social sciences 142 (originally published in 1876, author "metallurgist, engineer, Commissioner of Geological Survey of Pennsylvania").

Peltier, L. C., 1949, Pleistocene terraces of the Susquehanna River, Pennsylvania: Pennsylvania Geological Survey, 4th ser., Bulletin G 23, 158 p.

Peltier, L. C., 1950, The geographic cycle in periglacial regions as it is related to climatic geomorphology: Association of American Geographers Annals, v. 40, p. 214-236.

Pettijohn, F. J., 1957, Sedimentary rocks: New York, Harper and Row, 628 p.

Pettyjohn, W. A. and Henning, R., 1979, Preliminary estimate of ground-water recharge rates, related streamflow and water quality in Ohio: The Ohio State University Department of Geology and Mineralogy Project Completion Report No. 522, 323 p.

Péwé, T. L., 1970, Altiplanation terraces of early Quaternary age near Fairbanks, Alaska: Acta Geographica Lodziensia, v. 24, p. 357-363.

Péwé, T. L., 1975, Quaternary geology of Alaska: U.S. Geological Survey Professional Paper 835, 145 p.

Péwé, T. L., 1983, The periglacial environment in North America during Wisconsin time, in Porter, S. C., ed., Late-Quaternary Environments of the United States. Volume 1. The Late Pleistocene: Minneapolis, University of Minnesota Press, p. 157-189.

Péwé, T. L., Church, R. E., Andresen, M. J., 1969, Origin and paleoclimatic significance of large-scale patterned ground in the Donnelly Dome area, Alaska: Geological Society of America Special Paper 103, 87 p.

Pierce, K. L., 1965, Geomorphic significance of a Cretaceous deposit in the Great Valley of southern Pennsylvania: U.S. Geological Survey Professional Paper 525-C, p. C152-C156.

Pierce, K. L., 1966, Bedrock and surficial geology of the McConnellsburg quadrangle, Pennsylvania: Pennsylvania Geological Survey, 4th ser., Atlas 109a, 111 p.

Poag, C. W., and Sevon, W. D., 1989, A record of Appalachian denudation in postrift Mesozoic and Cenozoic sedimentary deposits of the U.S. Middle Atlantic continental margin, in Gardner, T. W., and Sevon, W. D., eds., Appalachian Geomorphology (reprinted from Geomorphology, v. 2, no. 1-3): Amsterdam, Elsevier Science Publishers, p. 119-157.

Priesnitz, K., 1988, Cryoplanation, in Clark, M. J., ed., Advances in Periglacial Geomorphology: Chichester, John Wiley and Sons, p. 49-67.

Ramsay, J. G., 1967, Folding and fracturing of rocks: New York, McGraw-Hill, 568 p.

Ramsay, J. G. and Huber, M. I., 1983, The tectonics of modern structural geology, Vol. 1, Strain Analysis: New York, Academic Press, Inc., 307 p.

Rankin, D. W., 1967, Guide to the geology of the Mount Rogers area, Virginia, North Carolina, and Tennessee: Carolina Geological Society Field Trip, Guidebook, 48 p.

Rankin, D. W., 1975, The continental margin of eastern North America in the southern Appalachians: The opening and closing of the Proto-Atlantic Ocean: American Journal of Science, v. 275-A, p. 298-336.

Rankin, D. W., Stern, T. W., Reed, J. C., Jr., and Newell, M. F., 1969, Zircon ages of felsic rocks in the Upper Precambrian of the Blue Ridge, Appalachian Mountains: Science, v. 166, p. 741-744.

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

Reger, R. D. and Péwé, T. L., 1976, Cryoplanation terraces: indicators of a permafrost environment: Quaternary Research, v. 6, p. 99-109.

Resser, C. E. and Howell, B. J., 1938, Lower Cambrian Olenellus zone of the Appalachians: Geological Society of America Bulletin, v. 49, p. 195-248.

Rogers, H. D., 1858, The geology of Pennsylvania: Philadelphia, v. 1, 586 p.

Root, S. I., 1968, Geology and mineral resources of southeastern Franklin County, Pennsylvania: Pennsylvania Geological Survey, 4th ser., Atlas 119cd.

Root, S. I., 1970, Structure of the northern terminus of the Blue Ridge in Pennsylvania: Geological Society of America Bulletin, v. 81, p. 815-830.

Root, S. I., 1971, Geology and mineral resources of northeastern Franklin County, Pennsylvania: Pennsylvania Geological Survey, 4th ser., Atlas 119ab, 104 p.

Root, S. I., 1977, Geology and mineral resources of the Harrisburg West area of Cumberland and York Counties, Pennsylvania: Pennsylvania Geological Survey, 4th ser., Atlas 148ab, 106 p.

Root, S. I., 1978, Geology and mineral resources of the Carlisle and Mechanicsburg Quadrangles, Cumberland County, Pennsylvania: Pennsylvania Geological Survey, 4th ser., Atlas 138ab, 1 p.

Root, S. I., 1989, Basement control of structure in the Gettysburg rift basin, Pennsylvania and Maryland: Tectonophysics, v. 166, p. 281-292.

Root, S. I. and Hoskins, D. M., 1977, Lat. 40°N fault-zone, Pennsylvania: a new interpretation: Geology, v. 5, p. 719-723.

Root, S. I. and MacLachlan, D. B., 1978, Western limit of Taconic allochthons in Pennsylvania: Geological Society of America Bulletin, v. 89, p. 1515-1528.

Rorabaugh, M. I., 1964, Estimating changes in bank storage and ground-water contribution to streamflow:

International Association of Scientific Hydrology Publication no. 63, p. 432-441.
Rose, A. W., 1970, Metal mines and occurrences in Pennsylvania: Pennsylvania Geological Survey, 4th ser., Mineral Resources Report 50, Part 3, 14 p.
Russell, R. J., 1933, Alpine land forms in western United States: Geological Society of America Bulletin, v. 44, p. 927-950.
Saad, D. A. and Hippe, D. J., 1990, Large springs in the Valley and Ridge physiographic province of Pennsylvania: U.S. Geological Survey Open-File Report 90-164, 17 p.
Schmidlin, T. W., 1982, Leffler's method of estimating average temperatures of Appalachian summits: Evaluation in New York: Journal of Applied Meteorology, v. 21, p. 745-747.
Schofield, C. and Trojnar, J. R., 1980, Aluminum toxicity to brook trout (Salvelinus fontinalis) in acidified waters, in Toribara, T. Y., Miller, M. W., and Morrow, P. E., eds., Polluted Rain: New York, Plenum Press, p. 341-365.
Schwab, F. L., 1970, Origin of the Antietam Formation (Late Precambrian?-Lower Cambrian), central Virginia: Journal of Sedimentary Petrology, v. 40, p. 354-366.
Schwab, F. L., 1972, The Chilhowee Group and the Late Precambrian-Early Paleozoic sedimentary framework in the central and southern Appalachians, in Lessing, P. and others, eds., Appalachian structures - origin, evolution, and possible potential for new exploration frontiers: Morgantown, West Virginia University and West Virginia Geological and Economic Survey, p. 59-101.
Scotese, C. R. and others, 1979, Paleozoic base maps: Journal of Geology, v. 87, p. 217-277.
Sears, C. E., 1957, Late Cretaceous erosion surface in southwest Virginia (abs.): Geological Society of America Bulletin, v. 68, p. 1883.
Seilacher, A., 1967, Bathymetry of trace fossils: Marine Geology, v. 5, p. 413-428.
Sepkoski, J. J., Jr. and Knoll, A. H., 1983, Precambrian-Cambrian boundary: The spike is driven and the monolith crumbles: Paleobiology, v. 9, p. 199-206.
Sevon, W. D., 1984, A sandstone weathering rate from northeastern Pennsylvania (abs.): Geological Society of America Abstracts with Programs, v. 16, no. 1, p. 63.
Sevon, W. D., 1985, Pennsylvania's Polygenetic Landscape: Harrisburg Area Geological Society, 4th Annual Field Trip Guidebook, 55 p.
Sevon, W. D., 1987, The Hickory Run boulder field, a periglacial relict, Carbon County, Pennsylvania, in Roy, D. C., ed., Northeastern Section of the Geological Society of America, Centennial Field Guide, Vol. 5: Geological Society of America DNAG Volume, p. 75-76.
Sevon, W. D., 1989, Erosion in the Juniata drainage basin, Pennsylvania, in Gardner, T. W., and Sevon, W. D., eds., Appalachian Geomorphology (reprinted from Geomorphology, v. 2, no. 1-3): Amsterdam, Elsevier Science Publishers, p. 303-318.
Sevon, W. D., 1990, The Hickory Run boulder field, Hickory Run State Park, Carbon County, Pennsylvania: Northeastern Geology, v. 12, p. 42-45.
Sevon, W. D., Potter, Jr., N., and Crowl, G. H., 1983, Appalachian peneplains: an historical review: Earth Sciences History, v. 2, p. 156-164.
Sharpe, W. E., DeWalle, D. R., Liebfried, R. T., Dinicola, R. S., Kimmel, W. G., and Sherwin, L. S., 1984, Causes of acidification of four streams on Laurel Hill in southwestern Pennsylvania: Journal of Environmental Quality, v. 13, p. 619-631.
Shirk, W. R., 1980, Geology of southcentral Pennsylvania: Chambersburg, PA, Robson and Kaye, Inc., 136 p.
Siddens, L. K., Seim, W. K., and Curtis, L. R., 1986, Comparison of continuous and episodic exposure to acidic, aluminum-contaminated waters of brook trout (Salvelinus fontinalis): Canadian Journal of Fisheries and Aquatic Sciences, v. 43, p. 2036-2040.
Simpson, E. L., Linski, D., Mull, M. F., Keiser, J. P., Horsnall, S. L., and Hendricks, J. S., 1991, Depositional processes in outer-shelf sediments of the Lower Cambrian Harpers Formation of the Chilhowee Group, south-central Pennsylvania (abs.): Geological Society of America Abstracts with Programs, v. 23, no. 1, p. 127.
Simpson, E. L. and Sundberg, F. A., 1987, Early Cambrian age for synrift deposits of the Chilhowee Group of southwestern Virginia: Geology, v. 15, p. 123-126.
Simpson, K. W., Bode, R. W., and Colquhoun, J. R., 1985, The macroinvertebrate fauna of an acid-stressed headwater stream in the Adirondack Mountains, New York: Freshwater Biology, v. 15, p. 671-681.
Sims, S. J., 1985, Geological reconnaissance survey and reserve estimate - R. A. Bender & Son Sand Quarry, Mt. Holly Springs, Dickinson Township, Cumberland County, Pennsylvania: unpublished private report for Pennsy Supply, Inc., 18 p.
Sloss, L. L., 1963, Sequences in the cratonic interior of North America: Geological Society of America Bulletin, v. 74, p. 93-114.
Smith, H. T. U., and Smith, A. P., 1945, Periglacial rock streams in the Blue Ridge area (abs.): Geological Society of America Bulletin, v. 56, p. 1198.
Smith, R. C., II, 1978, The mineralogy of Pennsylvania, 1966-1975: Friends of Mineralogy, Pennsylvania Chapter, Special Publication No. 1, 304 p.
Smith, R. C., II and Berkheiser, S. W., Jr., 1985, Silica 99 and 44/100% pure, waiting to be be mined: Pennsylvania Geology, v. 16, no. 1, p. 2-4.

Spencer, E. W., 1988, Introduction to the structure of the Earth: New York, McGraw-Hill, 551 p.
Spencer, E. W., Bell, J. D., and Kozak, S. J., 1989, Valley and Ridge and Blue Ridge traverse, central Virginia: 28th International Geological Congress, Washington, D.C., Field Trip Guidebook, in Glacial geology and geomorphology of North America, vol. 2: American Geophysical Union, 69 p.
Stone, R. W., 1923, Roofing granules industry in southeastern Pennsylvania: Pennsylvania Geological Survey, 4th ser., Progress Report 82, 4 p.
Stone, R. W. and Hickok, W. O., 1933, Trip 4, South Mountain, in Ashley, G. W., ed., 3rd Annual Field Conference of Pennsylvania Geologists, Harrisburg, PA, Guidebook, p. 11-14.
Stose, G. W., 1906, The sedimentary rocks of South Mountain, Pennsylvania: Journal of Geology, v. 14, p. 201-220.
Stose, G. W., 1907, White clay of South Mountain, Pennsylvania: U.S. Geological Survey Bulletin, v. 315, p. 322-334.
Stose, G. W., 1909, Mercersburg-Chambersburg folio: U.S. Geological Survey Geologic Atlas, Folio No. 170, 19 p.
Stose, G. W., 1932, Geology and mineral resources of Adams County, Pennsylvania: Pennsylvania Geological Survey, 4th ser., Bulletin C1, 153 p.
Stose, G. W., 1953, Geology of the Carlisle Quadrangle, Pennsylvania: U.S. Geological Survey Geological Quadrangle Map GQ-28.
Stose, G. W. and Bascom, F., 1929, Fairfield-Gettysburg folio: U.S. Geological Survey Geologic Atlas, Folio No. 225, 22 p.
Stose, A. J. and Stose, G. W., 1957, Geology and mineral resources of the Gossan lead district and adjacent areas: Virginia Division of Mineral Resources Bulletin 72, 233 p.
Swain, L. A., Hollyday, E. F., Daniel, C. C., III, Zapecza, O. S., Mesko, T. O., and Wright, W. G., 1991, Plan of study for the regional aquifer-system analysis of the Appalachian Valley and Ridge, Piedmont, and Blue Ridge physiographic provinces of the eastern and southeastern United States, with a description of study-area geology and geohydrology: U.S. Geological Survey Water-Resources Investigations Report 91-4066, 54 p.
Swartz, F. M., 1948, Trenton and sub-Trenton of outcrop areas in New York, Pennsylvania, and Maryland: American Association of Petroleum Geologists Bulletin, v. 32, p. 1493-1595.
Taber, S., 1943, Perennially frozen ground in Alaska: its origin and history: Geological Society of America Bulletin, v. 54, p. 1433-1548.
Tarr, W. S., 1898, The peneplain: American Geologist, v. 21, p. 351-370.
Taylor, F. B., Dion, J. A., and Collins, J. J., 1986, Drinking water quality and acid rain in the eastern United States: Water Quality Bulletin, v. 11, p. 50-57.
Thomas, W. A., 1977, Evolution of the Appalachian-Ouachita salients and recesses from reentrants and promontories in the continental margin: American Journal of Science, v. 277, p. 1233-1278.
Thornbury, W. D., 1954, Principles of Geomorphology: New York, John Wiley and Sons, 618 p.
Thornbury, W. D., 1965, Regional geomorphology of the United States: New York, John Wiley and Sons, 609 p.
Thornbury, W. D., 1969, Principles of Geomorphology, 2nd ed.: New York, John Wiley and Sons, 594 p.
Tiffney, B. H., 1985, The Eocene North Atlantic land bridge: its importance in Tertiary and modern phytogeography of the northern hemisphere: Journal of the Arnold Arboretum, v. 66, p. 243-273.
Troll, C., 1948, Der subnivale oder periglaziale Zyklus der Denudation: Erdkunde, v. 2: p. 1-21.
Tschudy, R. H., 1965, An Upper Cretaceous deposit in the Appalachian Mountains: U.S. Geological Survey Professional Paper 525-B: p. B64-B68.
U.S. Geological Survey, 1991, Water resources data for Pennsylvania, water year 1990--volume 2; U.S. Geological Survey Water-Data Report PA-90-2, 274 p.
Walcott, C. D., 1891, Correlation papers, Cambrian: U.S. Geological Survey Bulletin 81, 447 p.
Walcott, C. D., 1892a, Notes on the Cambrian rocks of Virginia and the southern Appalachians: American Journal of Science, v. 44, p. 52-57.
Walcott, C. D., 1892b, Notes on the Cambrian rocks of Pennsylvania and Maryland from the Susquehanna to the Potomac: American Journal of Science, v. 44, p. 669-682.
Walcott, C. D., 1893, The geologist at Blue Mountain, Maryland: National Geographic Magazine, v. 5, p. 84-88.
Walcott, C. D., 1894a, Notes on the Cambrian rocks of Pennsylvania from the Susquehanna to the Delaware: American Journal of Science, v. 47, p. 309-311.
Walcott, C. D., 1894b, On the Occurrence of Olenellus in the Green Pond Mountain series of northern New Jersey, with a note on the conglomerates: American Journal of Science, v. 47, p. 309-311.
Walcott, C. D., 1894c, Paleozoic intraformational conglomerates: U.S. Geological Survey Bulletin 134, p. 1-43.
Walcott, C. D., 1896, The Cambrian rocks of Pennsylvania: U.S. Geological Survey Bulletin 134, 43 p.
Walker, J. E., 1966, Hopewell village, a social and economic history of an iron-making community: Philadelphia, University of Pennsylvania Press, 526 p.
Wallace, P. A. W., 1961, Indians in Pennsylvania: The Pennsylvania Historical and Museum Commission, 138 p.
Washburn, A. L., 1980, Geocryology: New York, John Wiley and Sons, 406 p.
Washburn, A. L., 1985, Periglacial problems, in Church, M. and Slaymaker, O., eds., Field and Theory.

Lectures in Geocryology: Vancouver, University of British Columbia Press, p. 166-202.
Watts, W. A., 1979, Late Quaternary vegetation of central Appalachia and the New Jersey Coastal Plain: Ecological Monographs, v. 49, p. 427-469.
Way, J. H., 1986, Your guide to the geology of the Kings Gap area, Cumberland County, Pennsylvania: Pennsylvania Geological Survey, 4th ser., Environmental Geology Report 8, 31 p.
Webb, J. R., Cosby, B. J., Galloway, J. N., Hornberger, G.M., and Ryan, P. F., 1989a, The Shenandoah watershed study: An overview: Department of Environmental Sciences, University of Virginia, Charlottesville, VA.
Webb, J. R., Cosby, B. J., Galloway, J. N., Hornberger, G.M., and Ryan, P. F., 1989b, Acidification of native brook trout streams in Virginia, a report to the Virginia Department of Game and Inland Fisheries: The Virginia Trout Stream Sensitivity Study, Project Number F-72.
Whisonant, J. C., 1970, Paleocurrents in basal Cambrian rocks of eastern Tennessee: Geological Society of America Bulletin, v. 81, p. 2781-2786.
Whitaker, J. C., 1955a, Direction of current flow in some Lower Cambrian clastics in Maryland: Geological Society of America Bulletin, v. 66, p. 763-766.
Whitaker, J.C., 1955b, Geology of Catoctin Mountain, Maryland and Virginia: Geological Society of America Bulletin, v. 66, p. 435-462.
White, S. E., 1976, Rock glaciers and block fields, review and new data: Quaternary Research, v. 6, p. 77-97.
Wilden, R. and Mabey, D. R., 1961, Giant desiccation fissures on the Black Rock and Smoke Creek Deserts, Nevada: Science, v. 133, p. 1359-1360.
Wilderman, C.C., 1990, Patterns of variation of pH and alkalinity in Pennsylvania streams, based on data obtained by the Alliance for Acid Rain Monitoring (ALLARM), a citizen's monitoring program, in Proceedings of the Conference on Atmospheric Deposition in Pennsylvania: A Critical Assessment, September 11-14, 1989: Environmental Resources Research Institute, The Pennsylvania State University, University Park, PA., p. 94.
Wilderman, C. C., Cole, C. A., and Nguyen, Q. T., in press, Quality trends of some public water supplies in Pennsylvania: Northeastern Geology.
Wilderman, C. C. and Reuss, C. L., 1991, Analysis of patterns and causes of variation in pH and alkalinity concentrations in Pennsylvania streams, using data collected by the Alliance for Acid Rain Monitoring (ALLARM): Pennsylvania Academy of Science Journal, v. 64, p. 210.
Williams, P. J. and Smith, M. W., 1989, The Frozen Earth: Fundamentals of Geocryology: Cambridge, Cambridge University Press, Studies in Polar Research, 306 p.
Wilshusen, J. P. and Sevon, W. D., 1981, Giant Ripples at Mount Cydonia: Pennsylvania Geology, v. 12, p. 2-8.
Wilshusen, J. P. and Sevon, W. D., 1982, Mount Cydonia Sand Company pit, in Potter, N., Jr., ed., South Mountain: Harrisburg Area Geological Society, Guidebook, p. 28-32.
Wolfe, J. A., 1985, Distribution of major vegetational types during the Tertiary, in Sundquist, E. T. and Broecker, W. S., eds., The carbon cycle and atmospheric CO_2: natural variation Archean to present: American Geophysical Union Monograph 32, p. 357-375.
Wood, G. D. and Clendening, J. A., 1982, Acritarchs from the Lower Cambrian Murray Shale, Chilhowee Group of Tennessee, U.S.A.: Palynology, v. 6, p. 255-265.
Woodward, N. B., 1985, Valley and Ridge thrust belt: balanced structural sections Pennsylvania to Alabama: University of Tennessee, Department of Geological Sciences, Studies in Geology 12, 64 p.
Worsley, P., 1984, Periglacial environment: Progress in Physical Geography, v. 8, p. 270-276.
Zietz, I., Haworth, R. T., Williams, H., and Daniels, D. L., 1980, Magnetic anomaly map of the Appalachians: Memorial University of Newfoundland, Map No. 2, scale 1:1,000,000.