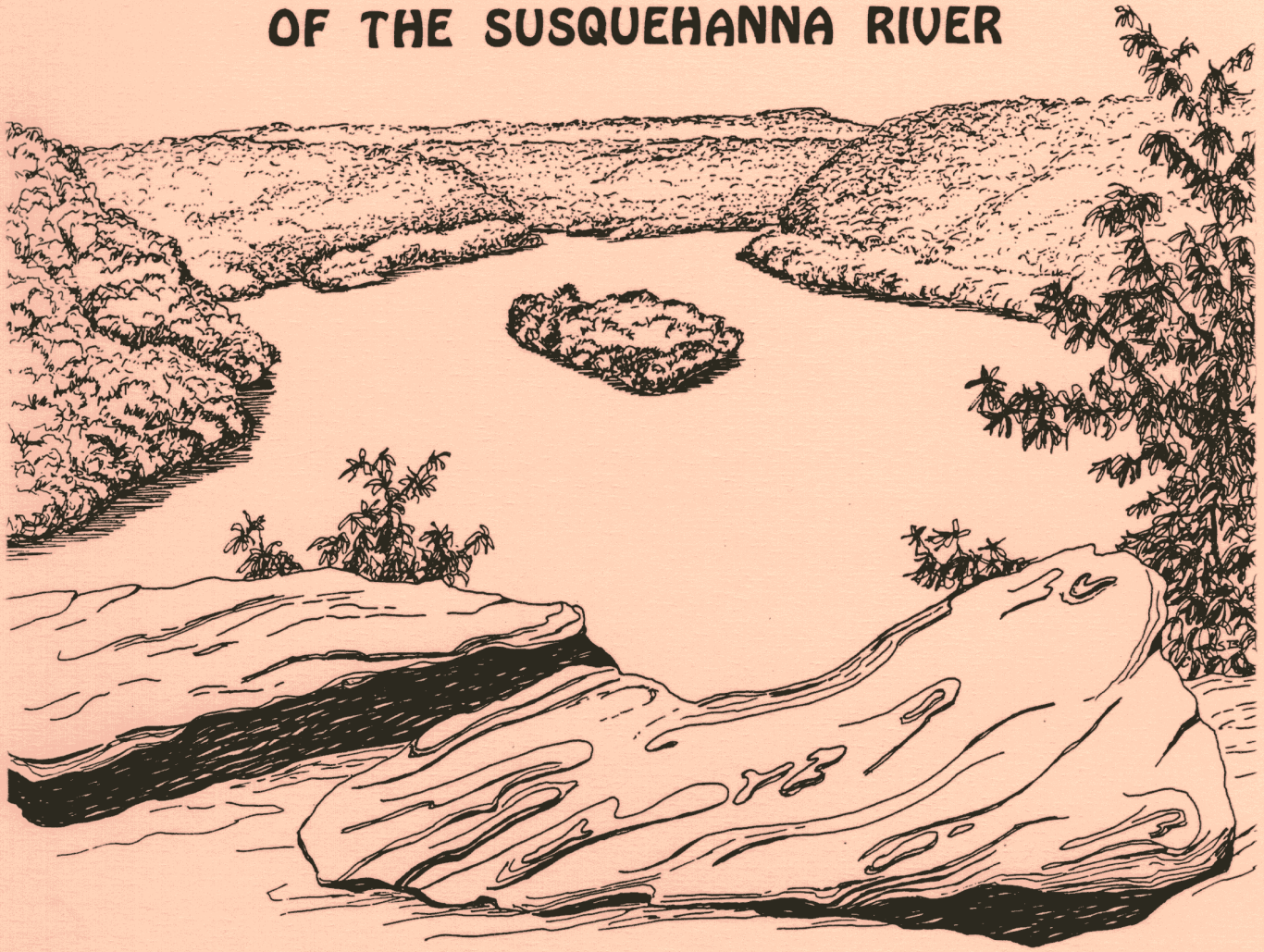
GUIDEBOOK

55th Annual Field Conference Of Pennsylvania Geologists

CARBONATES, SCHISTS, AND GEOMORPHOLOGY IN THE VICINITY OF THE LOWER REACHES OF THE SUSQUEHANNA RIVER



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55th Annual Field Conference of Pennsylvania Geologists
October 4-6, 1990
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Carbonates, Schists, and Geomorphology in the Vicinity of the Lower Reaches of the Susquehanna River

GUIDEBOOK CORRIGENDA

PAGE	LINE	READS	SHOULD READ
iii	4	Pre-WWI	Pre-WWII
9	27	(Faill, 1987)	(Faill and Geyer, 1987)
11	30	(Armbruster and Seeber, 1964)	(Armbruster and Seeber, 1987)
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86	32	Petters Creek Schists	Peters Creek Schists
112	FIG. VI-1	**Add to caption:** After Smeltzer, 1963	
156	20	framers	farmers
160	6	agricultural aggregate	construction aggregate
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223	4	mesotectonic	mesotectonite
225	48	on the river, crossing at	on the river-crossing at

Figure caption and figure caption citation corrections - Chapter III

55 Figure III-15A: S2 schistosity and F2 folds in chloritic quartzite from Peters Creek Formation south of Drumore. The view is looking east-northeast at a vertical surface.

57 Figure III-11E: Dextrally sheared quartz vein in Peters Creek lithology with in the zone of penetrative S2 schistosity; view is looking down on a subhorizontally orieted exposure surface.

57 Figure III-11F: Type I S-C mylonitic fabric in the Cardiff conglomeratic quartizite indicating dextral shear; view is looking down on a surface cut perpendicular to the S2 schistosity and parallel to the L2 lineation; field of view is 2.5 millimeters.

60 Paragraph 3, line 4: Figure III-1 citation should also include Figure III-15A.

61 Paragraph 2, line 9: Figure III-11E should be Figure III-11F

61 Paragraph 3, line 3: Figure III-11F should be Figure III-11E

61 Paragraph 3, line 5: Figure III-15A should be Figure III-11F

Guidebook for the

55th ANNUAL FIELD CONFERENCE OF PENNSYLVANIA GEOLOGISTS

CARBONATES, SCHISTS, AND GEOMORPHOLOGY IN THE VICINITY OF THE LOWER REACHES OF THE SUSQUEHANNA RIVER

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Cover: The gorge of the lower Susquehanna River, looking upstream from "The Pinnacle," near the crest of the Tucquan Antiform. Pen and ink drawing by Sheridan Bromer.

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Frontispiece. Aerial view of the quarry of Delta Carbonate, Inc., York, Pa. (STOP11); North is at the top.

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If anyone has been omitted from this list who should be there, it is solely the fault of the editor and he apologizes.

Additional acknowledgements of individual authors will be found in their respective chapters.

I. INTRODUCTION TO THE FIELD CONFERENCE AND OVERVIEW OF THE GEOLOGY OF THE LOWER SUSQUEHANNA REGION

Charles K. Scharnberger
Millersville University

INTRODUCTION

Themes of the Field Conference

Welcome to the 55th annual Field Conference of Pennsylvania Geologists, headquartered in historic Lancaster, Pennsylvania. This field conference will concentrate on three major aspects of geology in the vicinity of the lower reaches of the Susquehanna River. One of these, to be considered primarily on the first day of the conference, concerns the structural and metamorphic history of the schists (and other rock types) occurring south of the "Martic Line," the problematical boundary between the Conestoga Valley and the Piedmont Uplands. A second theme, also considered during the first day's excursion, is the geomorphic character of the uplands (including the origin and character of saprolite), and the Pleistocene/Holocene history of the Lower Susquehanna Gorge itself. A sub-theme here is the role that the river has played in the history of transportation and commerce in the region. The third theme, developed primarily on the second day, is the stratigraphy, paleontology, paleoenvironment, structural history, and economic importance of the Cambrian carbonate rocks in the Conestoga Valley. A consideration of the Conestoga Formation, a rock that varies from mildly deformed and unrecrystallized limestone to complexly deformed phyllitic marble, serves as a link between the first and third of our themes.

Local History and Culture

Laid out in 1730, Lancaster bills itself as "The oldest inland city in the United States." It was the home of George Ross, signer of the Declaration of Independence, Edward Hand, adjutant to Gen. George Washington, James Buchanan, the only President from Pennsylvania, and the birthplace of Robert Fulton. Today Lancaster County is a major agricultural, manufacturing, transportation and education center, and a popular tourist destination. The basis of the tourist industry is the presence of various sects of "plain people," particularly the Old Order Amish.

Although less well known to tourists, York County is no less historic or scenic than Lancaster. The City of York was founded in 1741, the first European settlement in Pennsylvania west of the Susquehanna River. York served as the nation's capital for nine months while the British occupied Philadelphia. It was in York that the Continental Congress adopted the Articles of Confederation, issued the first national currency, and commissioned von Steuben and Lafayette. York is an important industrial center and one of the leading producers of non-metallic mineral products in the United States, as discussed in Chapter X.

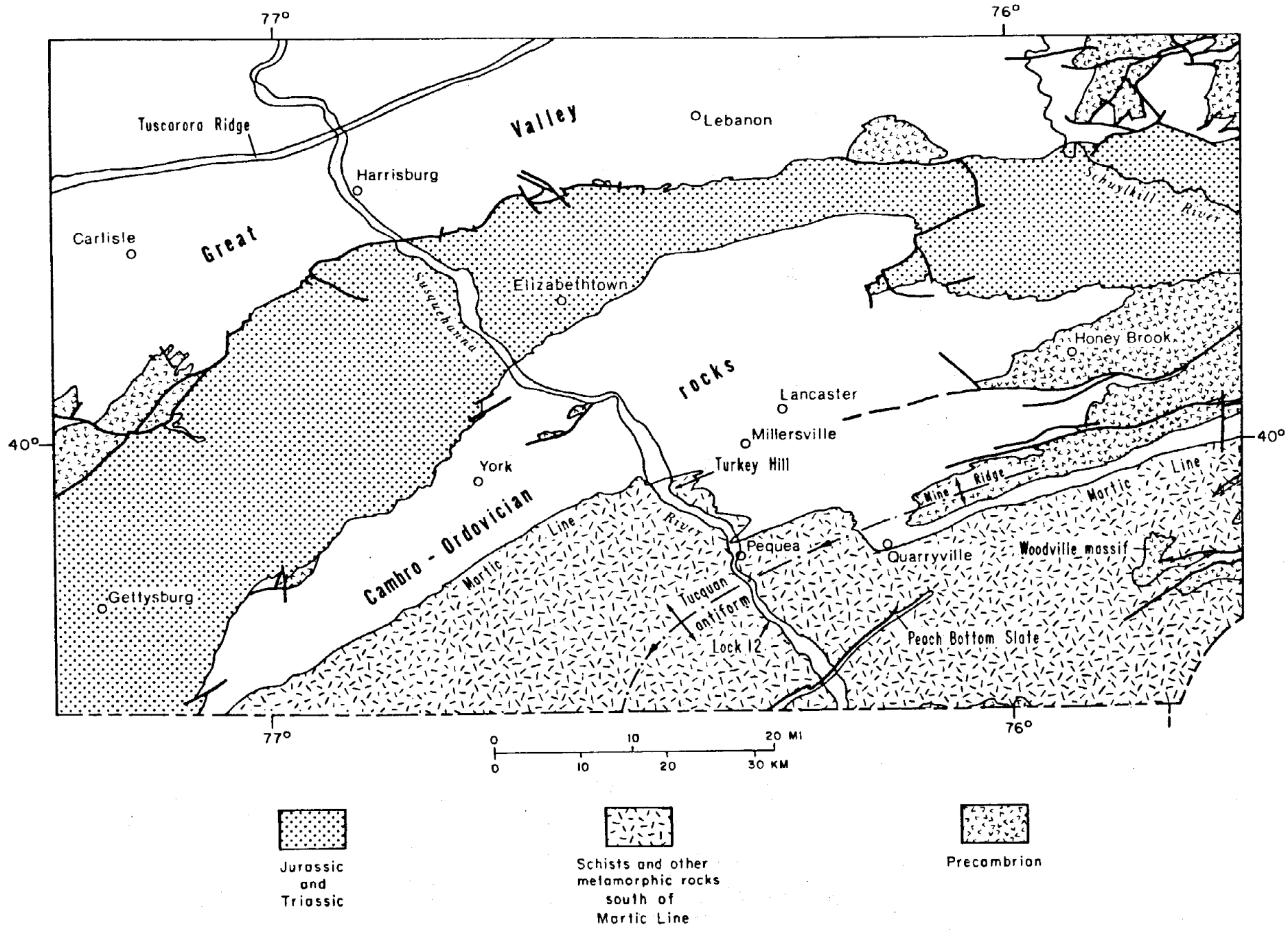

Figure I-1. Regional setting for the 55th Annual Field Conference of Pennsylvania Geologists

During the 19th century, the lower Susquehanna River was an important avenue of transportation via the Susquehanna and Tidewater Canal which ran along the west shore of the river. An account of the history of the canal appears in Chapter VI.

PHYSIOGRAPHY AND DRAINAGE

Physiography

Lancaster and York Counties are located primarily in the Piedmont and "Triassic Lowland" physiographic provinces. The latter term is put in quotation marks because: 1) the rocks in this province are partly Jurassic, and 2) locally, this province forms a highland relative to surrounding carbonate valleys. Even the designation "Piedmont" may be misleading insofar as that term implies the presence everywhere of highly deformed and recrystallized rocks.

In general, the region may be described as comprising three broad belts trending ENE (Figure I-1): a northern belt of hills underlain by Triassic-Jurassic clastic sediments and diabase intrusions (the Furnace Hills, not visited on this field conference), a central valley underlain mostly by carbonates but interrupted by hills and ridges of quartzite (labelled "Cambro-Ordovician Rocks" on Figure I-1), and a southern upland developed on bedrock that is largely schist, but includes many varieties of metamorphic rock (the Piedmont Upland, known locally as the Martic Hills, sometimes the River Hills, and farther east as the South Valley Hills).

There is some confusion and difference of opinion about what name should be applied to the valley between the Triassic and Piedmont uplands. Stose and Stose (1944) refer to the part of the valley west of the Susquehanna as the "Hanover-York Valley." Knopf and Jonas (1929) use the name "Lancaster Valley" for the part east of the river. Gohn (1976) uses the term "Conestoga Valley" as a general name for both the eastern and western portions, a usage that is followed in Chapters VIII and IX of this guidebook. On the other hand, some believe that the name "Conestoga Valley" should be restricted to areas that are drained by the Conestoga River; see Chapter VII for that point of view. The author of this chapter has no strong view on the subject, but, for convenience, will use "Conestoga Valley" in the broad sense of Gohn (1976), while acknowledging that not everyone would agree with this usage.

In addition to the three belts described above, significant massifs of Precambrian basement gneisses occur in the eastern part of the region (the Honey Brook Upland and Mine Ridge, see Figure I-1).

At its widest point, east of the city of Lancaster, the Conestoga Valley is nearly 25 miles (40 km) wide. Because areas of

quartzite bedrock are relatively small and scattered, the Conestoga Valley is, for the most part, truly a valley in Lancaster County. West of the Susquehanna, the valley narrows, both because the Martic Line steps to the north at the river, and because the diagonal trend of the Triassic belt "squeezes" the valley against the Piedmont Upland. At the city of York, for example, the width of the valley is reduced to approximately 8 miles (13 km), and only half of this is a topographic valley because of hills underlain by relatively resistant clastic rocks along the north and south margins. Immediately west of the river, near Wrightsville, most of the width of the "valley" is occupied by quartzite highlands, so that the carbonate portion is only about 1.5 miles (2.4 km) wide. The valley eventually pinches out in Adams County, near Hanover. Eastward, the Conestoga Valley ends against the Honey Brook-Mine Ridge Massifs, although a narrow arm continues eastward as the Chester Valley.

Drainage, Relief and Soil

The area is drained by the Susquehanna River (the boundary between Lancaster and York Counties) and numerous tributaries. Principal tributaries that drain the area south of the Furnace Hills include, in Lancaster County: Conoy Creek, Chickies (or Chiques) Creek, the Conestoga River, Pequea Creek, and Octoraro Creek; on the York side are Codorus Creek, Fishing Creek, Otter Creek, and Muddy Creek. The drainage pattern generally is dendritic, though there is a suggestion of a rectangular pattern in many places. Entrenched meanders are common, and the Susquehanna has cut a spectacular gorge through the Martic Hills. A peculiar feature of the tributaries that enter the Susquehanna from the Martic Hills is their "inverted," i.e., convex upward, longitudinal profiles. This point and the general geomorphology of the area around the gorge is discussed in Chapters IV and V.

The landscape has low to moderate relief, and generally has an appearance that usually is described as "rolling." Elevations in the Martic Hills reach a maximum of about 900 feet (275 meters) above sea level. Maximum local relief in the vicinity of the Susquehanna Gorge is slightly over 500 feet (153 meters). A large variety of soils, greatly varying in thickness, have developed on the bedrock. Some of the residual soils on carbonate rocks are rich in clay; on schist, deep saprolite has formed in many places, as described in Chapter IV.

BEDROCK GEOLOGY

History of Geologic Study

The geology of Lancaster County was investigated by Frazer (1880) for the Second Pennsylvania Geological Survey. The "classic" mapping was done in the 1920s and '30s by Anna Stose (nee Jonas), Eleanora Knopf (nee Bliss), and George Stose (Jonas and Stose, 1926, 1930; Knopf and Jonas, 1929; Stose and Jonas, 1933; Stose and Stose, 1939b, 1944). This work formed the basis

for all subsequent studies and engendered the famous "Martic Line controversy" discussed in Chapter II. Part of the area was remapped by Ernst Cloos (Cloos and Hietanen, 1941) and during the 1950s and '60s, the faculty and students at Franklin and Marshall College undertook a number of studies of the geologic structure in Lancaster County (e.g., Wise and Kauffman, 1960; Freedman and others, 1964; Wise, 1970). Recently, the Pennsylvania Geological Survey has begun a major mapping project in the Piedmont Province, including southern Lancaster and York Counties. Some of the results of this work are discussed in Chapters III and VII, and at various field stops of the conference.

While much of the early work concentrated on the structural and stratigraphic relationships of the rocks in the vicinity of the Martic Line, a second area of investigation developed around the lower Paleozoic strata (quartzite, phyllite, limestone and dolomite) of the Conestoga Valley (Stose and Jonas, 1922; Jonas and Stose, 1926, 1930; Stose and Stose 1939b, 1944). Rodgers (1968) presented a regional synthesis that interpreted this part of the section as representing a Cambro-Ordovician carbonate bank, perhaps similar to the Great Bahama Bank, along the coast of paleo-North America. A key feature of Rodgers's interpretation was the recognition that the Conestoga Formation is a deeper-water facies, equivalent in age to many of the other formations. Subsequent work has borne out this interpretation, as will be demonstrated during the field conference (see Chapters VIII and IX). The carbonate rocks have been studied also by Meisler (1968), Meisler and Becher (1971), and by Gohn (1976). Poth (1977) published a study of the ground-water resources of Lancaster County (including a geologic map of the entire county on a non-topographic base), while Lloyd and Growitz (1977) conducted a similar study in York County. Their publication includes a reproduction of the geologic map of Stose and Stose (1939b).

Stratigraphy

As a starting point for discussion, the following generalized stratigraphic column can be given for the rocks north of the Martic Line (based on Meisler and Becher, 1971):

ORDOVICIAN:	Cocalico Shale (probably allochthonous)	
	Myerstown Limestone	
	Annville Limestone	
	Beekmantown Group	F C
CAMBRIAN:	Conococheague Group	O O
	Zooks Corner Fm.	R N
	Ledger Dolomite	M E
	Kinzers Formation	A S
	upper member	T T
	middle member	I O
	lower member	O G
	Vintage Dolomite	N A
	Antietam Fm. (quartzite & schist)	
	Harpers Phyllite	
	Chickies Quartzite	
	Hellam Conglomerate Mbr.	
PRECAMBRIAN:	Gneisses or metavolcanics	

Rock units above the Ledger Dolomite are not visited on this field conference and will not be considered further here.

The Conestoga Formation

The arrangement of the name "Conestoga Formation" in this column is intended to emphasize the time-transgressive nature of this formation, and its relationship to the other carbonate units as a deeper-water facies. The age of the Conestoga Formation has long been problematical. Stose and Stose (1944) gave it as "Ordovician (?)", but state very well the meager basis for that assignment (p. 37):

> At Henderson Station in the eastern part of Chester Valley south of Norristown, limestones tentatively correlated with the Conestoga have yielded the only determinable fossils that may belong to this formation. E. O. Ulrich and A. F. Foerste have assigned Beekmantown age to these cephalopods and gastropods. The limestones in Chester Valley are therefore of Lower Ordovician age. The age of the Conestoga limestone in the type locality in Lancaster Valley and westward in the Hanover-York district can be ascertained only if fossils are obtained in that region. Consequently, at present the formation is classed as of probably Ordovician age.

Fossils now have been obtained from the Conestoga Formation west of the city of York (Chapter IX; STOP 10), and the age is determined to be Middle Cambrian. That is not to say that the formation could not cross the Cambrian-Ordovician boundary and be Lower Ordovician in part, but in Lancaster and York Counties, at least, "Cambrian" seems a better general designation than "Ordovician."

Gohn (1976) informally divided the Conestoga Formation in York County into three members: an upper Wrightsville Member, underlain by both a West York Member (in the northern part of the York Valley), and a Kreutz Creek Member (in the southern part of

the valley). These rocks are described, and a paleoenvironmental interpretation of them given, in Chapter IX. The basic model is that of a deep-water basin to the southeast, shoaling through toe-of-slope, slope, shelf margin and shelf environments toward the northwest. Similar facies have been recognized in Lancaster County (Stose and Stose, 1944; Rodgers, 1968), but the greater degree of deformation and metamorphic overprint in the Conestoga Formation east of the Susquehanna River makes stratigraphic subdivision and paleogeographic interpretation more difficult there.

The Ledger Formation

The Ledger Formation is described and interpreted in Chapters VIII and IX. This formation includes reef facies and seems to represent, primarily, a shallow platform environment (Taylor, Chapter IX, this guidebook). The traditional designation for the lithology of the Ledger is "dolomite." Recently, however, Ganis and Hopkins (Chapter VIII, this guidebook) have discovered a middle limestone member in the West York Block, and have proposed a 3-fold subdivision of this formation into an Upper Dolomite Member, a middle Willis Run Member, and a Lower Dolomite Member.

The Kinzers Formation

The 3-fold division of the Kinzers Formation, though clearly recognizable in both York and Lancaster Counties, is not without ambiguity because of significant changes in thickness, facies and age of these members as one traces them across the Conestoga Valley. Rodgers (1968) considered the Kinzers (especially the upper and lower members) to be a tongue of Conestoga facies that encroached on the shallower-water carbonates during temporary subsidence of the shelf. Gohn (1976) proposed the names "Longs Park Member," "Thomasville Member," and "Emigsville Member" for the upper, middle and lower members, respectively, of the Kinzers Formation. The stratigraphy and paleontology of this formation is discussed in detail in Chapters VIII and IX, where new nomenclature for the York area (West York Block) is proposed.

The Vintage Dolomite

This formation represents the transition between the basal clastics and the carbonate platform and periplatform facies that will be examined on the second day of the field conference. The upper part of the Vintage is an off-platform, largely turbidite facies (Taylor and Durika, Chapter IX, this guidebook).

Basal Cambrian Clastics

The clastic rocks at the bottom of the Cambrian sequence (the Chillhowee Group) will not be visited on this field conference, except that the Hellam Conglomerate will be seen at STOP 8. These formations probably represent an Early Cambrian marine transgression. The trace fossil *Scolithus* (or *Skolithus*), found

in the sandy facies of the Chickies Formation, seems to have been a burrower of the shallow marine shelf.

Lateral facies changes are common in these rocks. For example, the Hellam Conglomerate occurs as discontinuous lenses near the base of the Chickies Formation in York County, but is absent east of the Susquehanna. The rest of the Chickies Formation is predominantly quartzite in the northern part of the valley, but slate on the southern side. This suggests that the model of deeper water to the south, shoaling toward the north, as indicated by the Conestoga Formation, may apply also in the time of deposition of the basal clastic rocks.

Precambrian Rocks

An interesting point, not elaborated here, is that in Lancaster County the basal clastic sequence rests on a basement of granitic gneiss, probably of Grenvillian age. In York County, however, the few exposures of rocks below the Chickies Formation are of metabasalt and metarhyolite (Stose and Stose, 1944). Whether these are the same as the Catoctin Volcanics of the Blue Ridge Province is not clear. To the best of this writer's knowledge, they have not been studied in detail, and it would not be surprising if the protoliths turned out to be more complex than simply basalt and rhyolite.

Rocks of the Piedmont Uplands

South of the Martic Line, stratigraphic relationships are obscure, both among the rocks there and between those rocks and the Cambrian strata to the north. This is, in part, what the "Martic Controversy" is about, as discussed in some detail in Chapter II. Rock-unit names which have been applied to the rocks south of the Martic Line include: Wissahickon Schist (also Wissahickon Gneiss), Marburg Schist (used only west of the Susquehanna), Peters Creek Schist, Octoraro Schist (also Octoraro Phyllite), Cardiff Conglomerate (a stretched-pebble conglomerate), and Peach Bottom Slate. These rocks probably are metasediments for the most part, but Stose and Stose (1939b, 1944) mapped many units within the schists of York County that they interpreted as metavolcanics.

Until recently, the Peach Bottom Slate was thought to be the youngest of these formations, located in the center of a syncline. Work by Higgins (1972) and Valentino (Chapter III, this guidebook), however, casts doubt on the syncline interpretation. The Peach Bottom Formation, nevertheless, could still be relatively young, perhaps even Ordovician as suggested by Stose and Stose (1944), largely on the basis of physical correlation of Peach Bottom rocks with slates of known Ordovician age in Virginia (the Quantico and Arvonia Slates). Stose and Stose (1944) cite Lesley (1879a) as reporting the presence of an Ordovician alga (*Buthotrephis flexuosa)* in the Peach Bottom Slate, but go on to say (p. 52), "...the specimens cannot be located and no other fossils have been obtained from the area."

Because Stose and Stose (1944) regarded the Wissahickon and Peters Creek Schists as probably Precambrian, they took an Ordovician age for the Peach Bottom Slate (and immediately underlying Cardiff Conglomerate) as indicative of a significant unconformity, with the entire Cambrian System absent. If, however, the Peach Bottom Formation is simply an argillaceous facies within what we might call the "Peters Creek Group," then an Ordovician age for the Peach Bottom would imply that many, if not all, the rocks in the Martic Hills should be assigned to the Lower Paleozoic rather than the Precambrian. It should be emphasized, however, that the case for an Ordovician age for the Peach Bottom Slate is a very weak one.

In the literature, the rocks south of the Martic Line generally are referred to as part of the "Glenarm series," but that term has fallen into disuse, partly because "series" has a chronostratigraphic meaning that is not properly applied here, and partly because the term implies correlation with rocks to the south and east (perhaps as far away as New York--see Knopf and Jonas, 1929, and Chapter II) that may not be correct. Clearly, some new stratigraphic nomenclature is needed for these rocks, but correlation with established time-rock units is likely to remain an elusive goal for the time being.

Structure

Folds and Associated Fabrics

In the northern part of the Lancaster Valley, the carbonate beds are folded into recumbent structures that are spectacular where exposed (Faill, 1987). This leads to the interpretation of this part of the Conestoga Valley as an early Paleozoic (Taconian) nappe (Rodgers, 1970). This early (D1) deformation also affected the Conestoga Formation and the schists in the southern part of the valley, producing an "S1 fabric" (see Chapters III, VII and discussion for STOP 7), but large recumbent folds are not obviously present there. The part of the valley underlain largely by the Conestoga Formation has been described as a "syncline" (Knopf and Jonas, 1929) or a "synclinorium" (Rodgers, 1970).

The nearly upright folds that are obvious in the southern part of the Lancaster Valley appear to have been formed by a second (early Alleghanian?) deformation, and so are designated F2 (Freedman and others, 1964; Valentino, Chapter III, this guidebook; Faill and MacLachlan, STOP 7 description). Recently, Valentino and MacLachlan (1990; also described in Chapter III and STOP 7 description) have recognized a "Lancaster Valley Tectonite Zone" characterized by prominent S2 cleavage. Valentino (1990 and Chapter III, this guidebook) proposes a significant amount of strike-slip displacement (dextral shear) in this zone, as well as both dextral and sinistral displacements on other faults in the lower Susquehanna region. In the extreme southern part of the

valley, folds (probably F2) generally are open and verge toward the south (Cloos and Heitanen, 1941; personal observation).

West of the Susquehanna River, folding becomes less intense as one travels westward and northward across two major faults: the Stoner and Gnatstown "Overthrusts," discussed below. In the West York Block (described in Chapter VIII), folding is gentle and metamorphic fabrics virtually absent. The reason for this lack of D1 fabrics and only mild D2(?) deformation, in marked contrast to the rocks just to the east, is not known at this time, but may be the subject of some lively speculation during the field conference.

A major fold, the Tucquan Antiform (Chapter III, and the descriptions for STOPS 1, 2 and 3) affects the rocks south of the Martic Line on both sides of the river. Eastward this fold appears to pass into the Mine Ridge Uplift, though there may be some discordance between the two structures (see Chapter III). Folds also are associated with the Honey Brook Uplift.

Faults

The early workers in the lower Susquehanna region put a great emphasis on overthrust faults (e.g., Knopf and Jonas, 1929; Stose and Jonas, 1935; Stose and Stose, 1939b, 1944). Besides the "Martic Overthrust," coincident with the Martic Line, they mapped a number of other overthrust faults, including the Mine Ridge, Stoner, Gnatstown, Highmount and Chickies Overthrusts (among others). The story of the overthrust controversy is well told by Wyckoff in Chapter II and will not be discussed in detail here. Suffice it to say that some of these faults, at least, certainly exist, but whether they have very low angles of dip (as the term "overthrust" implies) is difficult to determine. High-angle faulting might do as well as an explanation of the observed map patterns. It is this writer's preference to use the term "Gnatstown Fault", etc., rather than Gnatstown Overthrust when discussing these structures.

Besides the overthrusts, many (probably) high-angle faults striking ENE, parallel to the principal structural grain can be recognized. In many places, blocks of basal clastic rocks, especially the Antietam Formation, apparently have been raised relative to the carbonates to form quartzite ridges within the carbonate terrain. Alternatively, these exposures could be klippen derived from low-angle overthrusting.

Recently, Valentino (as discussed and referenced above) has introduced the idea of strike-slip faulting in the region, apparent today as shear zones, the deeper portions of transcurrent fault zones now exposed by uplift and erosion. Valentino suggests dextral slip on the Peach Bottom Structure (in the southern Lancaster and York Counties) as well as in the southern part of the Lancaster Valley Tectonite Zone. He further suggests sinistral movement on the Brandywine Manor Fault that cuts through the

Honey Brook Upland Massif (see Chapter III, especially Figure III-16). An especially intriguing idea is that the Brandywine Manor Fault may be continuous with the Stoner Fault, leading to further reconsideration of what the true nature of the "overthrusts" in York County may be.

On the other hand, Valentino does find evidence of thrusting (again, see Chapter III) suggesting that the Mine Ridge Overthrust, at least, is just that. MacLachlan (Chapter VII, this guidebook) argues for the reality of the Martic Overthrust. Finally, it may be said that the relatively small thrust faults mapped by Cloos (Cloos and Heitanen, 1941) just north of the Martic Line (see Chapter II, Figure II-5) probably provide the simplest interpretation of the structure there.

Post-Paleozoic Structures

Structures indicative of brittle deformation cross-cut all older structures in the region of the lower Susquehanna River. Most of these probably date from the time of crustal extension that accompanied Late Triassic-Jurassic rifting of Pangaea. Many prominent joints strike slightly east of north. Knopf and Jonas (1929) describe high-angle, probably normal faults with the same trend, many marked by brecciated and sheared rock. A number of diabase dikes (Late Triassic or Early Jurassic) cut across the Paleozoic grain with about the same strike as the faults and joints.

At least one of these faults (possibly more than one) is active today. Lancaster and York Counties have experienced approximately 20 locally-generated earthquakes since the late 18th century (the "Lancaster Seismic Zone" of Armbruster and Seeber, 1987; see also Scharnberger, 1989). Fault-plane solutions based on first-motion studies (Armbruster and Seeber, 1964) indicate that the fault responsible for the 1984 "Martic" earthquake (m_b=4.1) strikes N10°E, dips 60° east and experienced reverse slip (with some dextral strike-slip) at the time of the earthquake. This is consistent with a model of a Mesozoic normal fault that is being reactivated by modern east-west compressive stress.

Sevon (Chapter IV, this guidebook) discusses critically the possibility of relatively recent folding (warping) in the region, e.g., the Westminster Anticline.

II. A HISTORY OF THE MARTIC LINE CONTROVERSY
The Pre-WWII Years

A Series of Lectures Delivered by
Dorothy Wyckoff (1900-1982)
Professor of Geology (1930-1966)
Bryn Mawr College

Prepared for Publication by
W. A. Crawford
Department of Geology
Bryn Mawr College

PREFACE

Dorothy Wyckoff (Figure II-1) was born on the 22nd of July, 1900, in Topsfield, Massachusetts. Her father was a Congregational minister. All three of her degrees were earned at Bryn Mawr College: A.B. 1921, Greek and Latin; M.A. 1928, geology; and Ph.D. 1932, petrology. Dorothy joined the faculty of Bryn Mawr College in 1930 as a demonstrator and retired with the rank of professor in 1966. Her principal interests lay in metamorphic petrology, crystallography, and medieval science.

The influence of her great teacher and fellow New Englander, Florence Bascom, determined the direction of her geological work, beginning with the unraveling of the highly altered rocks of the Mt. Gausta region in Telemark, Norway, which was the subject of her dissertation, and culminating in the fifties in her exacting delineation of the metamorphic facies of the Wissahickon Schist of southeastern Pennsylvania. Her background and continued interest in classics resulted in the translations of Albertus Magnus on ore deposits.

During World War II she joined the Military Geology Unit of the U.S. Geological Survey to work with matters of strategic planning intelligence. From 1943 to 1945 she produced "terrain diagrams" which were used for planning assault operations and became famous for their accuracy and clarity.

Quiet, precise, with a sharply penetrating intellect and a very special order of integrity, without the slightest trace of flamboyance, she deeply affected her advanced students and colleagues. She steadfastly refused to compromise with careless or inferior work and a long line of students were rigorously trained in the intricacies of crystal optics, the universal stage, and phase equilibria.

Though a person of many talents and much loved by her students, Dorothy was also a very private person. She shunned all public recognition and in a letter to President McBride of Bryn Mawr College turning down the Lindback award upon her retirement stated, "I do not know, of course, what the donors of the Award have in mind; but if it is to encourage and reward good teaching,

Figure II-1. Dorothy Wyckoff (1900-1982)

this seems a sad way to do it--by a grant on retirement, when the teacher's days of teaching are over. Any money intended for such a purpose could so much better be spent on the young--as a small counterbalance, perhaps, to the many pressures toward research nowadays exerted even on those whose gifts are of another sort. And any recognition of good teaching would mean so much more if it came early in life--if it made possible, for instance, travel not tied to a research grant, or clerical assistance, or the enlargement of a personal library."

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[The biographic material for Dorothy Wyckoff was gleaned from an account by E. H. Watson, written on the occasion of her retirement, from her obituary prepared by the Office of Public Information, Bryn Mawr College, and from a 1958 Vita from the files of the Office of the President, Bryn Mawr College.]

INTRODUCTION

In the early 1960s Dorothy Wyckoff presented a series of lectures on the history of the Martic Line controversy to college teachers on NSF sponsored field trips explaining the geology of the Piedmont Province. I came across these hand written lectures in department files in 1987 while preparing a presentation to honor another of Florence Bascom's Ph.D. students, Isabel Fothergill Smith. These notes, written by Dorothy in pencil, contain no evidence of erasures. I was astounded at their clarity. The information just seemed to flow from her mind to the page. Now turn to the first page of her lectures and begin your enlightenment on the history of the Martic Line controversy.

William A. Crawford
Professor of Geology
Bryn Mawr College

THE LECTURES

Everybody who knows anything about the geology of the Piedmont knows that we have had a great controversy over something called the "Martic Overthrust"--but very few know just how the controversy began, or why it is so difficult to settle. I think the best approach here is along historical lines, so I will begin with Florence Bascom (1862-1945), who was a remarkable woman, (Knopf, 1946; Ogilvie, 1945; Smith, 1981).

She was the daughter of John Bascom, Professor of Philosophy at Williams College and later President of the University of Wisconsin He, too, was a remarkable person, not least in that he encouraged his daughter to complete a college education and to train for a profession--not very usual in the 1870s and '80s, when she was growing up. She took her first degrees at Wisconsin, and then applied to the Johns Hopkins University as a graduate student in geology.

If I had more time I ought to tell you more about the Johns Hopkins University--which was at that time an unusual and exciting place, unique among American universities, in that it laid great stress on graduate work, which was organized on the German plan, with a regular course of study leading to the Ph.D. degree--a system that has since spread to most other American universities. Many of the faculty were young men who had themselves recently obtained Ph.D.s from German universities.

In geology, the "latest things" were the new uses of the petrographic microscope--the study, naming, and classification of igneous rocks as taught by Zirkel (1893) and Rosenbush (1896)--and geometrical crystallography, as taught by Goldschmidt (1886-1893, 1897). (X-ray methods had not yet been discovered.)

Miss Bascom applied to the Johns Hopkins Department of Geology, and was accepted as a "special" student. She was told that she could attend all lectures and laboratory work, but that they did not grant degrees to women. After she had been there a couple of years, however, working especially with the petrographer George Huntington Williams, they changed their minds and accepted her as a candidate for the Ph.D., and in 1893 she became the first woman to receive that degree from the University.

Now we come to Bryn Mawr College ("Jane Hopkins")--which was to be the center of the Martic controversy in later years. The president of Bryn Mawr was M. Carey Thomas (1st dean and 2nd president)--who was a great feminist. She was on the lookout for bright young women for her new faculty (the college began only in 1885) and she offered Miss Bascom a rather junior position, teaching an elective course in geology. (Nobody then thought geology a suitable career for women, and such a course was quite a novelty for a women's college.)

So, in 1895, Miss Bascom came to Bryn Mawr--she was given two small rooms and a cubbyhole on the top floor of Dalton Hall--

and there she built up a department, offering full undergraduate and graduate work; and she built up a library and collections of rocks and minerals, filling every cranny in the attics, until the floors sagged. She was head of the department until she retired in 1928. But she did two other things that were perhaps even more important--she made the first adequate geologic maps for a large area here in the Piedmont, and she trained another generation of women geologists, who really broke down the prejudice against them in the profession.

First, the maps: when Miss Bascom came in 1895, there were no adequate geologic maps of the Philadelphia area. (She told me once how discouraged she was when she first began to look for places to take her students--nothing but complex metamorphic rocks everywhere!) The lack of maps had to be remedied at once, so she started right in on geologic mapping of the area. And you must try to imagine what it was like here in those days--geologic field work was done on foot, or with a horse and buggy. (Miss Bascom was passionately fond of horses, so she enjoyed this part of it very much.) Perhaps outcrops were better then, too--most of the roads were dirt or gravel, with ledges of rock sticking up on steep slopes or in the ditches; most of the country was farmland with soil and float undisturbed by anything more powerful than a plough; and there were many small quarries, opened up to build a house or two or a dam for a small pond, or to [provide stone to] burn for lime. Her maps are still remarkably accurate so far as the actual areal distribution of rock types is concerned, though the interpretation of the rocks themselves may have changed in the course of time.

By 1904, she had done enough to read a paper at the G.S.A. meetings, and this was published (Bascom, 1905). The paper includes some discussion of all the rocks in the region, but I will mention only the parts that relate to our subject--the Paleozoic series.

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**************************************************
O     Wissahickon Mica Schist and Mica Gneiss.
e-O   Shenandoah Limestone
e     Chickies Quartzite
~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~
pe    Baltimore Gneiss
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Notice that the two terms for the Wissahickon record that there are apparently different rocks in different parts of the area. The age relation was determined along the South Valley Hills, where the rock apparently overlying the limestone is a muscovite-chlorite schist, locally with albite porphyroblasts. But south of Buck Ridge and especially at the type locality along Wissahickon Creek in Fairmount Park, it is a much coarser crystalline rock, always containing feldspar (oligoclase-andesine), locally rich in biotite as well as muscovite, and with such accessories as garnet, staurolite, kyanite or sillimanite. Obviously this "gneiss" is more highly metamorphosed than the

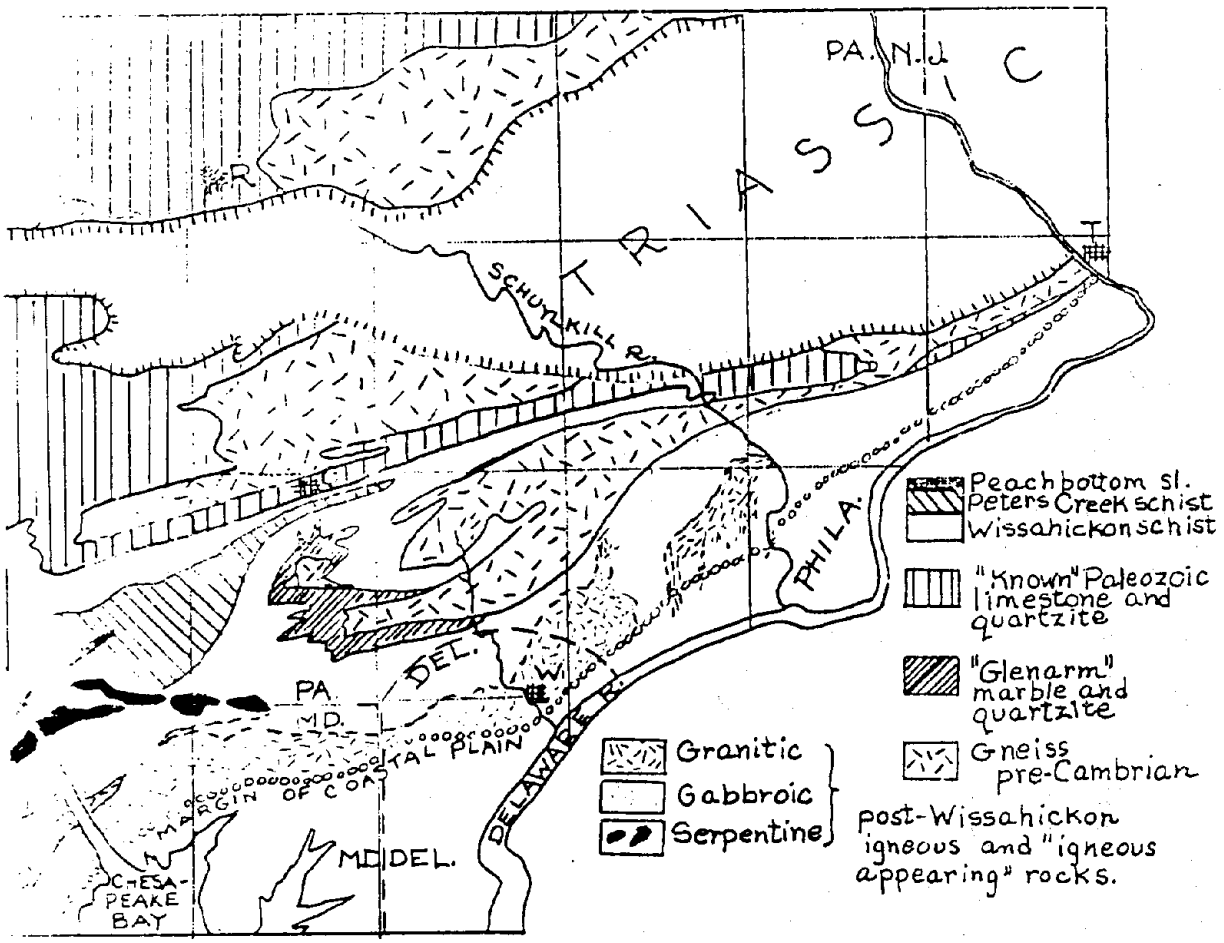
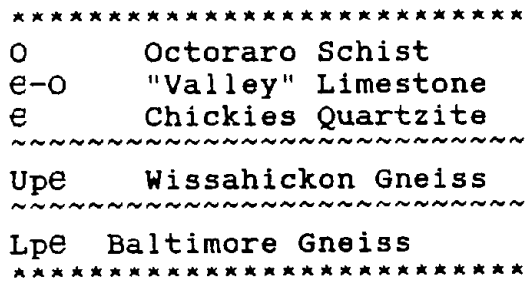


Figure II-2. Geologic map of SE Pennsylvania and nearby areas, ca. 1905

"schist," and this was attributed to the intrusions of igneous rocks in that part of the Wissahickon (Figure II-2).

But even before 1905 there were differences of opinion. Arthur Keith, George Otis Smith, and E. B. Matthews all took an interest in this work and they came up here and had a field conference with Miss Bascom (and their papers, too, were published (in Bascom, 1905; Matthews, 1905). They agreed that the Wissahickon Schist, which lies above the Chester Valley Limestone, was probably Ordovician, but they questioned the correlation of this with the Wissahickon Gneiss farther south and east. These counsels evidently prevailed, for by 1909, when the Philadelphia Folio (162, Bascom and others, 1909a) and the Trenton Folio (167, Bascom and others, 1909b) came out, she had separated the two. The Wissahickon Schist she renamed Octoraro Schist (for a type locality on Octoraro Creek) and retained in the Paleozoic series. The Wissahickon Gneiss she relegated to the Precambrian.



The folios gave two reasons supporting this interpretation. [First was] the more intense metamorphism of the Wissahickon, as compared with the Octoraro. And here, too, you must try to see all this in historical perspective. In 1909 modern studies of metamorphism were barely beginning. Van Hise's great treatise appeared in 1904 (Van Hise, 1904), Leith's first thesis bulletin on rock cleavage in 1905 (Leith, 1905)--"zones," "grades," and "facies" are concepts still in the making; e.g. Barrow's work in the Scottish Highlands did not come until 1912 (Barrow, 1912; Grubenmann, 1904), etc. The generally accepted assumption was that the greater the metamorphism, the older the rock must be, and therefore the Wissahickon "ought to be" older than the Octoraro.

The second reason was the "igneous unconformity"--the fact that the Wissahickon is intruded by numerous igneous rocks which do not intrude the "known" Paleozoics or the Octoraro. This seemed to mean that all the igneous activity and accompanying intense metamorphism were over and done with before the Paleozoic series was laid down.

Having got these two folios off her hands, Miss Bascom went right ahead with field work for two more, Wilmington-Elkton (211, Bascom and Miller, 1920) and Coatesville-West Chester (223, Bascom and Stose, 1932). But besides continuing with field work, she was teaching more and more courses at Bryn Mawr and had a number of excellent students. Among these were Anna I. Jonas (1881-1974, Dietrich, 1977) and Eleanora Bliss (1883-1974, Rodgers, 1977). They were great friends and worked together in the field, and for a dissertation subject, Miss Bascom suggested to them the area around Woodville and Avondale, in the middle of the Coatesville Quadrangle, part of the folio she was working on herself (Folio 223, Figure II-4). This joint dissertation was published (Bliss and Jonas, 1916).

As mapping proceeded westward--even within the Norristown quadrangle--difficulties began to develop. For instance, north of the Buck Ridge anticline, there seems to be Wissahickon Gneiss lying next to Octoraro Schist. Miss Bascom admitted that the two rocks are difficult to separate in the field; but if both are present, the contact between them must be a fault (Cream Valley Fault extending westward? See Figure II-3). But farther west, the Wissahickon and the Octoraro seemed actually to grade into each other. And in the Avondale District the structure is evidently very complex. Everyone agreed that at the S.W. end there

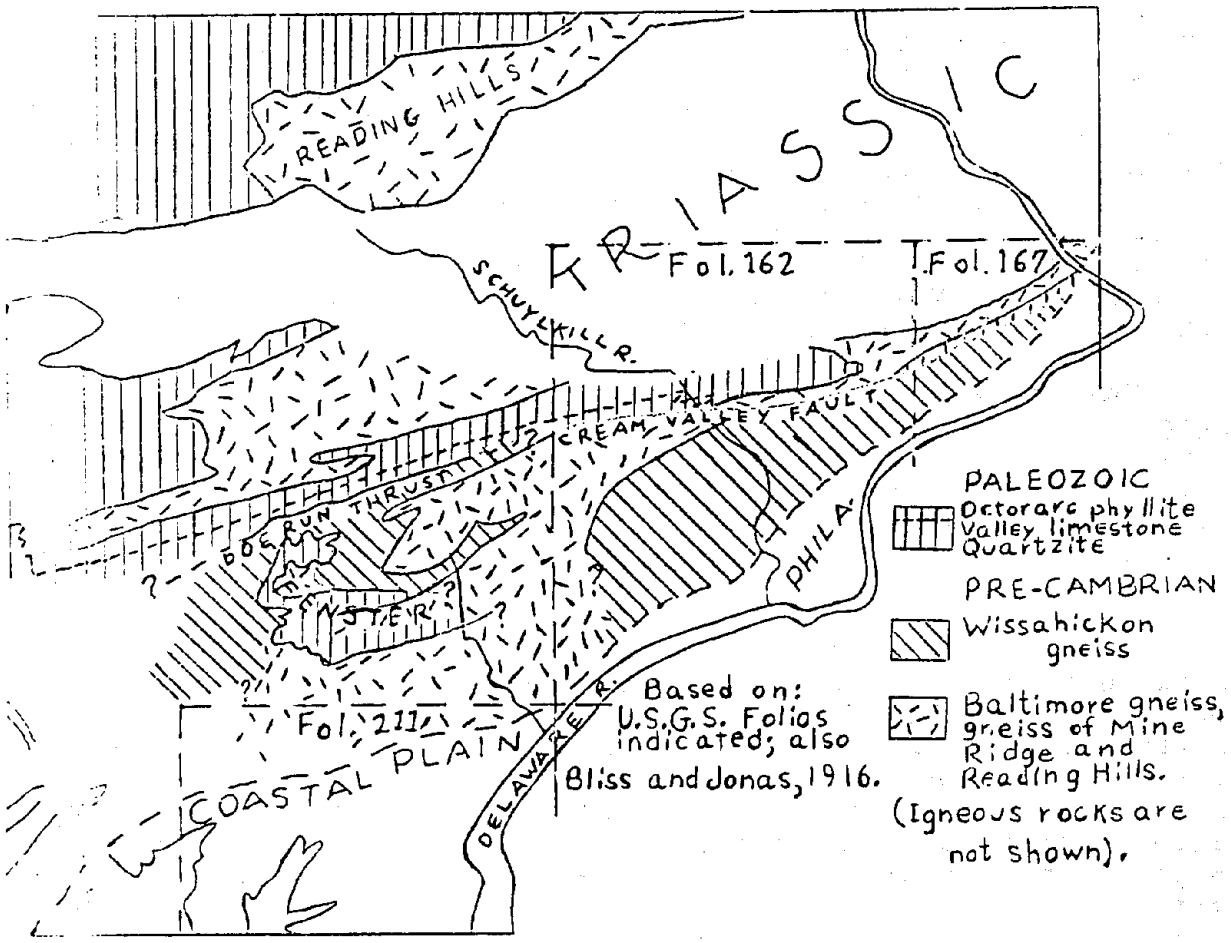


Figure II-3. Geology of SE Pennsylvania as mapped by Florence Bascom and her students

are two anticlinal domes or blocks, with old gneiss (correlated with the Baltimore Gneiss) in the cores; and on the gneiss rests a quartzite, succeeded by limestone (marble), succeeded by Wissahickon Gneiss.

Bliss and Jonas (following Miss Bascom's interpretation at the time), correlated the quartzite with the Chickies (Ꞓ) a few miles to the north, and the marble with the Chester Valley Limestone--but it had already been decided that the Wissahickon Gneiss was Precambrian and here it was, lying above Ꞓ-O marble! What they proposed, then, was an overthrust which had moved a great sheet of pꞒ rocks up and over the Paleozoics--the pꞒ Wissahickon and (as you will note on Figure II-2), the southern part of the area--which we now call the Wilmington complex--was also mapped as pꞒ gneiss. The thrust sheet had presumably been bowed up or domed in places, and especially around Woodville and Avondale the tops of these arches had been removed by erosion, forming "windows" or fensters exposing the younger Paleozoic rocks beneath the thrust plane.

This fault was called the Doe Run Overthrust and it is important to be clear that this is not the same as the Martic Overthrust proposed later, though it may, in a sense, be regarded as a sort of "ancestor" of the Martic Thrust. Notice that, at Avondale, the thrust plane is between the Wissahickon and the Octoraro (still regarded as Paleozoic), and so does not extend as far as the edge of Chester Valley--either the thrust thins out, or it is cut off by the westward extension of the Cream Valley Fault. The horizontal displacement was estimated to be at least 15-20 miles (to N or NW) and the authors cited some other examples of comparable magnitude in the southern Appalachians.

This paper (Bliss and Jonas, 1916) was, in a way, the beginning of a lot of trouble, ending in permanent estrangement between the authors and Miss Bascom (Arnold, 1983). And as we look back, we may think that at that time, the whole question of the age of the Wissahickon Gneiss might well have been reconsidered: for if the Wissahickon were Paleozoic and not Precambrian, there would have been no need for any overthrust at all.

Miss Bliss and Miss Jonas soon repudiated the views expressed in their 1916 paper. They sedulously avoided any reference to it in their later work, but in private they said they had never believed in the Doe Run Overthrust and had been forced into publishing it by Miss Bascom. This was both unkind and unfair to Miss Bascom--and there is no evidence that in 1916 they had themselves worked out any better alternative. It was not until 5 or 6 years later that they proposed a completely different interpretation. On the other hand, one of the psychological oddities of this whole story is that Miss Bascom never abandoned this interpretation. She was bitterly opposed to the Martic Overthrust when that was later advocated; but many of the arguments against the Martic Thrust would apply equally well to the Doe Run Thrust--and this she would never admit.

During the next few years, Miss Bliss and Miss Jonas started on their careers as professional geologists, preparing geologic maps and reports that were published by the Maryland and Pennsylvania Surveys, and by the U.S.Geological Survey and various national journals (e.g. Jonas, 1929, 1937; Knopf and Jonas, 1929; Stose and Jonas, 1939b and 1944; Stose, 1924a, 1924b; and Stose and Stose, 1946). Eventually they both married geologists: Miss Bliss became Mrs. Adolph Knopf (1920), and went to live at New Haven, where her husband was a professor at Yale. Miss Jonas later married her colleague, George Stose (1938), and continued to work with him for the U. S. Geological Survey in Washington. (These changes of name introduce some confusion into the bibliographies on this subject).

But Mrs. Knopf and Mrs. Stose continued to work together, and in 1921 they began to discuss--and by 1923-24 they published --a complete revision of the stratigraphy of the rocks of the

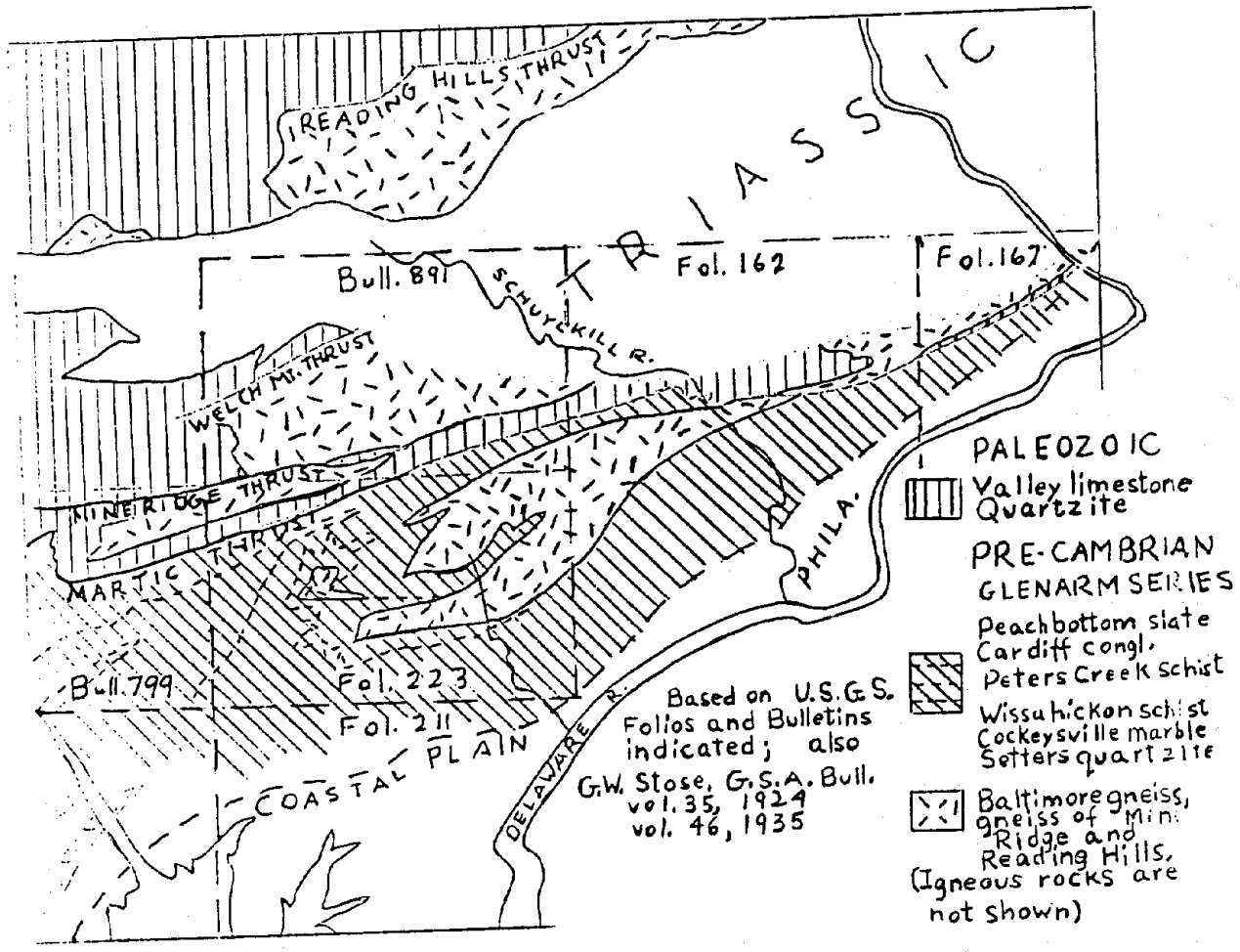


Figure II-4. Thrust faults as mapped by the Stoses.

Piedmont, setting up the Glenarm series as part of the Precambrian, named for a locality near Baltimore (Jonas and Knopf, 1923; Jonas, 1924; Hawkins, 1924). The column (which you already have) is shown on Figure II-2.

There was, of course, a good deal of resistance to this, and in judging the justification for it, it is important to remember that Mr. and Mrs. Stose had done, and for years continued to do, a great amount of field work on the early Paleozoic rocks of Pennsylvania, Maryland and Virginia. The "Valley" Limestone and the underlying arenaceous rocks were mapped in detail and subdivided into numerous distinct formations, etc. One must therefore take seriously their contention that these formations are not the same as the Cockeysville and the Setters--in lithology and thickness. But this opinion was challenged by other well-qualified geologists, whom we shall come to later.

One unfortunate result of the controversy was that George Stose (1869-1960, Miser, 1960), who had become chief editor of geologic maps at the [U. S. Geological] Survey in Washington

(1900-1943), was able to hold up for years the publication of the folios which Miss Bascom had completed before she retired in 1928. The Coatesville-West Chester Folio (233, Bascom and Stose, 1932 did not come out until 1932, and the Honeybrook-Phoenixville maps were put into a Bulletin (891, Bascom and Stose, 1938) which did not come out until 1938. And even then, although Miss Bascom's interpretations were briefly mentioned in the text, the maps had all been re-done according to the Stose-Jonas interpretation, showing the Martic Overthrust.

This was a great grief to Miss Bascom, and many other people felt that it was unfair, and began to "take sides" as the controversy widened. It was bound to widen, because of course the insertion of the Glenarm Series into the stratigraphic sequence made some thorough-going structural interpretations necessary.

In 1929 the Martic Overthrust was proposed, on the basis of structure in the McCalls Ferry-Quarryville District (Knopf and Jonas, 1929). The type locality is Martic Forge, near the Susquehanna (Figure II-5). But it is perhaps more interesting for us to consider it in the Doe Run-Avondale District where all the trouble had really begun.

At first sight the structural relations seem to be much simplified [by a thrust fault]. If the rocks overlying the Baltimore Gneiss are all Glenarm, they are in their proper order: Setters at the bottom (on the gneiss), then Cockeysville, then Wissahickon, then Peter's Creek, surrounding the anticlinal cores of the Avondale and Woodville Domes. But the difficulty is now shifted to the south edge of Chester Valley. If what used to be called the Octoraro is now taken to be part of the Wissahickon, of Precambrian age, the apparent conformable contact with the Paleozoic limestone cannot be what it seems--it must be a thrust plane, and here is where the whole sheet of Precambrian rocks must have been overthrust onto the younger Paleozoics. This, then, is the Martic Overthrust, and it has been discussed in detail in numerous reports and papers published by the Stoses and by Mrs. Knopf after 1929 (see Reference List). This interpretation was, of course that of the U.S. Geological Survey, and the Martic Overthrust is shown on the 1944 Tectonic Map of the U.S. (Longwell, 1944). In fact, the "Martic Line" came to be generally understood to be the contact between Paleozoic rocks on the N or NW and Precambrian on the S or SE.

Perhaps here, too, we can understand how the controversy developed further, if we know a little more about the personalities involved. Mrs. Stose was a bluff, brusque, active sort of person. In her prime she had enormous physical energy and a passion for field work. She did a vast amount of areal mapping and had a great deal of field experience to draw on.

Mrs. Knopf was more the intellectual--or if you like, the academic-type. She, too, had done much field work, but she gradually became more interested in theory--especially the new techniques for studying the fabric of metamorphic rocks being devel-

oped in Europe by Sander (1930), Schmidt (1925), and others. She is one of the pioneers of fabric analysis or petrofabrics in this country: she is co-author (with Ingerson) of the G.S.A. Memoir on petrofabrics (Knopf and Ingerson, 1939); she worked with Turner and others on the experimental investigation of the Yule marble, etc. (Knopf, 1949a & b; Turner, 1949). In the 1930s she was just beginning to turn her thoughts in this direction, but it is, I think, fair to say that Mr. and Mrs. Stose's field observations were largely interpreted by Mrs. Knopf's ideas of geologic structure and metamorphism.

They made a strong team. As Ernst Cloos used to say, "Those ladies convinced everybody by sheer rhetoric that the Martic Overthrust existed." They both published many papers in the 1930s and '40s (see Reference List). Mrs. Stose was working for the U. S. Geological Survey, and Mr. Stose was editor of geologic maps. Mrs. Knopf was in New Haven, where the American Journal of Science had its home, and her husband was one of the editors of the Journal.

But nevertheless, they did not really convince "everybody" and other voices began to be heard, and a sort of anti-Martic opposition began to gather. In the early 1930s, when Watson [Edward H. Watson, 1902-1975] and Dryden [Lincoln Dryden, 1903-1977] and I were beginning our teaching at Bryn Mawr, lots of people wanted to come and look at the Martic Overthrust on the spot. We had a number of field conferences or excursions in those years. Miss Bascom came back from Washington, the Stoses came up from Washington, Marland Billings came from New England, B. L. Miller from Lehigh, Balk from New York, Cloos from Baltimore, etc. All these people have contributed something, directly or indirectly, to the discussion, and I will say more about them later.

But before I go further, I want to emphasize that the so-called "Martic controversy" really embraces two different questions, though the way the thing had developed rather confused the issue. There are really two distinct problems, and it is not true that settling one of them would necessarily settle the other.

First, there is the question of the status of the Glenarm series--are there really two series of sediments, one Precambrian and one lower Paleozoic, or are the Glenarm rocks merely the metamorphic equivalents of the Cambrian-Ordovician rocks?

The other question is that of the Martic Thrust itself. Is this, so to speak, a purely mental construct, designed to account for the position of supposedly Precambrian rocks on top of known Paleozoic rocks? Or is there independent evidence that large scale thrusting has actually occurred?

As I say, answering one question does not automatically answer the other. For instance, it is conceivable that if there are two series, the Glenarm may not be overthrust onto the Paleozoics--possibly other faults, the Cream Valley Fault or branches

of it, separate the older from the younger rocks in some other structural relation. Or, it is conceivable that there is a great thrust above the Paleozoic limestone, but this would not prove that the rocks above the thrust are Precambrian--they might just be the higher grade facies of the Paleozoic rocks, once more deeply buried and later overthrust onto the lower grade rocks in Chester Valley.

In the early 1930s it was felt that the burden of proof rested on those who has proposed the thrust fault. The geologists who came here "to see the Martic Thrust with their own eyes"--what did they hope to see? There are at least two sorts of evidence that would have been convincing:

1. The existence of breccia, mylonites, mullion structures, etc., along the fault plane itself.

2. The cutting out of formations or structures at the fault line.

Let's see what evidence of this sort was found. The contact between the Chester Valley limestones and the overlying schist is in general badly exposed. In a few places, where it is seen, the two formations look as if they are conformable (as Miss Bascom had noted when she assigned the "Octoraro" Schist to the Ordovician).

But Mrs. Stose made quite a point of a zone of limonite and quartz between the limestone and the schist, interpreting it as a crush zone, with mineral replacement due to solutions moving along the fault plane. These limonite-quartz zones do exist--in fact some of them were mined in Revolutionary times and supplied the ore smelted at the small iron works whose memory is preserved in place names like Valley Forge, Martic Forge, etc. But the limonite zones are not confined to the "Martic line"--they occur elsewhere in the Paleozoics (even in the quartzite), apparently anywhere rocks of different lithologic character are in contact and ground water has percolated in the zone of weakness between them.

Mrs. Knopf also entered the fray on this point, with her paper on retrogressive metamorphism and phyllonitization (Knopf, 1931). The Martic Overthrust is not specifically mentioned in this paper, but anyone following the controversy could easily "read between the lines.". She made the point that in large scale overthrusting, such as occurs in the Alpine nappes, an incompetent material like a schist would not be crushed into a recognizable breccia or mylonite zone, it would merely undergo diaphthoresis, or retrogressive metamorphism, probably at fairly low temperature, but under intense shearing stress. Thus a high grade schist or gneiss would be transformed into a phyllite (or phyllonite, short for "phyllite-mylonite") of the greenschist facies.

The application of this to the Martic Thrust is clear--the whole zone called by Miss Bascom the "Octoraro" Schist, and renamed by the Stoses the "chlorite-albite facies" of the Wissahickon Schist is really a phyllonite, produced from the higher grade Wissahickon Schist by intense crushing and low grade recrystalization along the sole of the great Martic Thrust. This also explains why, in the Coatesville quadrangle, and farther west the higher grade schists seem to grade into the lower grade schists--and in the Peach Bottom Syncline, the south limb "goes under" as a schist and "comes up" on the north as a phyllite (phyllonite). Moreover, this would explain, too, the apparently conformable structures seen where the contact between schist (or phyllite) and limestone can be examined. The limestone (a notoriously plastic type of rock) has also been dragged along and sheared out on the sole of the fault. In fact, this whole argument makes the absence of breccias or obvious structural disconformities an evidence for the Martic Thrust rather than against it.

The papers in your bibliography by Woodward (1935), Mackin (1935), Fraser (1938), Miller (1935), Miller and Fraser (1935), and Stose (1935) represent this phase of the controversy (about 1934-36). In a sense the leader of the "anti-Martic" faction was B. L. Miller (1874-1944) (Ashley, 1945). As a young man he had taught here at Bryn Mawr College, and had worked with Miss Bascom on the very first folios, doing the sections on the Paleozoic rocks of Chester Valley. Then he went to Lehigh University at Bethlehem in the Great Valley, and besides teaching there he built up quite a reputation as consulting geologist in the cement industry. He certainly knew the Paleozoic limestones--we used to say he could tell the MgO content by smelling them!

He was a charming and kindly man and I have always thought that he was drawn into the controversy at least partly out of generous regard for Miss Bascom. He thought she was getting a "raw deal" at the Survey, and had had no real opportunity to publish her own views. But he also thought that the Martic Overthrust did not exist, at least along the south side of Chester Valley; and that it was being "put over" on the geologic public without really adequate evidence. And finally, he thought there had been altogether too much talk of overthrusts anyway, because it was just about this time that the Stoses proposed the Reading Overthrust (Stose and Jonas, 1935, 1939a, and 1940) which was in Miller's own bailiwick, the Lehigh Valley (Whitcomb, 1983).

I don't want to digress too much from our main topic, but I must say a little about this other controversy, since it had at least some psychological effect on most people who were concerned in the Martic controversy. The accepted interpretation of the structure of the "Reading Prong" was that it was a faulted area, with uplifted horsts of Precambrian gneiss, and down-dropped grabens which still preserve the Paleozoic limestone that once covered the horsts as well. Stose (Stose and Jonas, 1935) now interpreted the gneiss areas as klippe--"eroded remnants of a

great thrust sheet" of Precambrian rocks that had over-ridden the Paleozoic rocks of the Great Valley.

For some years this controversy bubbled along--field conferences were organized, and groups of geologists minutely inspected critical contacts. Miller collected information about well borings, and tried to raise money to have test bores put down at critical spots--and then had trouble in getting Stose to agree as to what locality they would both accept as "critical." I may say--to anticipate by many years, that recent work in that area has shown that the structures are not at all so simple as Miller supposed, and that the notion of large scale thrusts or nappe-like structures is probably not so fantastic as it seemed then. But at the time, distrust of the Reading Hills Overthrust rather reinforced distrust of the Martic Overthrust. And these misgivings were not allayed when Stose (1937) also proposed to make the Honeybrook Upland into another overthrust (Welch Mountain and Mine Ridge Thrusts, Figure II-4). And now some people began to dig in their heels and get stubborn about the whole thing. And they came to our field conferences, saying they wanted to "walk out" the whole contact at the Martic line, and to "get their noses right down on the fault plane."

This was easier said than done. In 1935 Miller published a very good summary of the state of the Martic controversy at that time, and I will now briefly review some of his main points.

One is the topographic expression of the rocks. From the Schuylkill River to Quarryville is 48 miles. Throughout this length the South Valley Hills make a very straight line, being held up by the schist, with the Valley itself floored by limestone. If the Hills are the topographic expression of an overthrust sheet, it is rather curious that erosion should have provided such a straight "front."

Second, Miller noted that everywhere along this line the schist is in contact with the upper Conestoga Limestone--which he said is only about 500' thick (E. H. Watson says more). This, too, would seem to be an unlikely coincidence--why should erosion have removed the edge of the thrust sheet just to this same line all the way along?

But if the schist is simply the next formation above the Conestoga Limestone, both the topographic expression and the stratigraphic sequence are just what would be expected, and similar to relations found in other parts of the Appalachians.

As to the other question--the age of the Glenarm Series, Miller also made some points [about fossil evidence] that we may keep in mind for future reference. Miller resurrected and put on record several old reports.

1. In the Survey of Pennsylvania published in 1858 H. D. Rogers reported Scolithus from the quartzite at Avondale (now called Setters). Scolithus is of course quite common in the

Chickies Quartzite of the North Valley Hills, only a few miles away. Roger's specimens have disappeared and many geologists have searched in vain at Avondale, without finding any more Scolithus there.

2. Then there is the question of fossils in the Peach Bottom slate (in the center of the Peach Bottom syncline, the youngest formation of the Glenarm series). Fossils were reported in 1879 by Lesley (1879b)and in 1884 by Frazer, and some specimens were sent to James Hall for identification. He thought some were graptolites, and some algae (or seaweed), Buthotrephis; and he correlated the Peach Bottom slate with the Hudson River slates of New York State, of Ordovician age. Miller wished to accept this correlation, but there has been a lot of dispute about it, and I will come back to this point again later: for the Peach Bottom slate was carefully re-studied in 1950 by Agron.

Miller, as I have said, was really going back to the earlier view, that the Wissahickon Formation (including the Octoraro) is all Ordovician. He completely rejected the Glenarm as a separate series, considering the Setters to be the metamorphic equivalent of the Cambrian Chickies Quartzite, and the Cockeysville that of the Conestoga Limestone (which is the only member of the "Valley" Limestone Series in the narrowest part of Chester Valley, nearest to the Avondale District). But unless the formations overlying the Conestoga are included in the lower Paleozoic, the sequence stops suddenly with the Conestoga, and there is no equivalent of the Martinsburg Slate which is an important member of the lower Paleozoic in the Lehigh Valley.

So Miller wished to correlate thus:

Lehigh Valley		(Pa. Piedmont)
		{Peachbottom Sl.
O	Martinsburg Slate	{Peters Creek Sch.
		{Wissahickon Sch.
Є-O	"Valley" Limestones	"Valley" Limestones
Є	Hardyston Quartzite	Chickies Quartzite
pЄ	gneiss	pЄ gneiss

Miller also called attention to the fact that in the Lehigh Valley the "cement rock" is between the limestone and the Martinsburg, and in Chester Valley, too, there is a band of "cement rock" between the limestone and the schist of the S. Valley Hills (this has been used for making cement in a plant near W. Conshohocken. Finally, Miller also raised still another question: if the Glenarm series is Precambrian, coming between the old gneiss and the known Paleozoic, how can we explain the fact that this whole thick series is entirely missing on the Honeybrook-Mine Ridge Upland, where Cambrian rests directly on the old gneiss? And yet it is present around and west of Avondale, only a few miles away. (Of course, the proponents of the Martic

Thrust could say that the two areas were not originally so close together--the Glenarm was presumably laid down much farther to the south or southeast, and only reached its present position so near to Mine Ridge because of tectonic transport.)

Miller's correlation, however, seemed good to a number of other geologists at the time (Mackin, 1935; Woodward, 1935). But today I think no one believes that the upper part of the Glenarm series is equivalent to the Martinsburg. The work that has led to this change of opinion I will review very soon.

But first, there is one more question--a very obvious one where stratigraphic relations are being discussed: isn't there any place where "known" Paleozoics are found resting upon indubitable Glenarm rocks? Such a relation would certainly settle the matter--and don't imagine that such field relations haven't been looked for--especially by the Stoses and Mrs. Knopf. The fact that even they reported very few such localities is in itself remarkable; and other geologists who have visited these localities have not been unanimously convinced.

The "critical" exposures are poor: at best, rocks supposed to be of Glenarm and of Paleozoic ages crop out near together, and although contacts are covered, the dips permit structural inferences to be made--but made by different people in different ways, since structures are complex and over-turned folds or small local thrusts are not impossible.

And in some cases even the Stoses have made different inferences at different times. One case was discussed by Cloos [and Hietanen] (1941): when Carroll County was mapped by Knopf and Jonas in 1928, they showed basal Cambrian resting on Glenarm schist and marble. But 10 years later, the new map (Jonas and Stose, 1938) of the same area showed all the rocks as Precambrian, so the importance of the area for establishing the Precambrian age of the Glenarm quite disappeared, since now there seems to be no Cambrian present.

Of course, such re-interpretation is always going on--and rightly so--as more detailed studies are pursued in regions of complex structure. But the fact that such uncertainties exist seems to show that Cambrian-Glenarm contact is pretty hard to find, and some people began to wonder if the difficulty wasn't that they were looking for something that isn't there!

There remains also the indirect correlation based on the fact that at South Mountain the basal Cambrian (Weaverton) rests on older volcanic rocks--and volcanics supposed to be of the same age are interbedded with Glenarm schists farther to the east. This would seem to indicate that the Glenarm also is Precambrian. But here too there has been argument--we do not really know that volcanic tuffs etc. were produced only in Precambrian time--volcanic activity is a common feature of the early stages of development of a geosyncline, and might have occurred in early Paleozoic just as well as in "Glenarm" time.

I have now brought the history of the controversy up to the late '30s, when it began to seem that the whole argument was getting "bogged down" for lack of anything new to say. What was needed were new ideas, new methods and new workers--and this now came about, with the contributions of Ernst Cloos (1898-) [now deceased] and Robert Balk (1899-1955). Not only were they younger than the original protagonists, and trained in newer methods of investigation, but coming to the problem fresh from Germany, they had the advantage of being uncommitted to either side of the dispute.

Ernst Cloos is the younger brother of Hans Cloos, who had already inaugurated the methods of structural petrology--then known as granit tectonik because he first applied them to granite massifs--the detailed measurement and analysis of structural elements--foliation, lineation, cleavage and jointings--as a means of unravelling the tectonic history of a rock body (Cloos, 1925). Ernst Cloos was interested in the same methods and lines of investigation. Robert Balk was about the same age, and had also been trained by Hans Cloos in Germany. Both of them began to apply these methods in America. Cloos went to teach at the Johns Hopkins University, and began to study the Piedmont rocks around Baltimore. Balk went to Hunter College, and began to work on the geology of New York State--first a study of the central Adirondacks, and then work in Dutchess County.

Since I am trying to keep to a more or less chronological plan, I am going to digress here to discuss some of Balk's work, because it had some effect--though only indirectly--on the thinking of those concerned with the "Martic controversy" in this region. In 1932, Balk published a preliminary statement and in 1936, collaborating with Tom F. W. Barth, a long paper, [in two parts]: "Structural and petrologic studies in Dutchess Co., N.Y." This was important to us because the rocks of southeastern New York and western Connecticut are very much like those of the Pennsylvania-Maryland Piedmont; and although the two areas are completely separated by the Triassic basin and the overlap of the Coastal Plain in New Jersey, it had generally been taken for granted that the Manhattan Schist should be correlated with the Wissahickon schist. Not only are the rocks similar, but the history of the geological investigation of them has also been rather similar, so I must give you a little more historical background here.

As in Pennsylvania, the earliest workers considered the rocks around New York City to be highly metamorphic equivalents of the lower Paleozoics recognized north of the Hudson Highlands.

****

O	Hudson River Slate
e-O	ls, dol. ("Wappinger"-now subdivided)
e	Poughquag Quartzite
pe	gneiss

****

But the idea that the more highly metamorphosed rocks were probably Precambrian came at just about the same time as in Pennsylvania (1907-1919) and the man chiefly responsible for this was C. P. Berkey, professor of geology at Columbia. He was consulting geologist for the Catskill aqueduct, and during this work he examined the whole section from Newburg southward through the Hudson Highlands and then across Westchester Co. to New York City (Berkey, 1907 and 1922). Nowhere along this line are "high grade" and "low grade" rocks in contact--the "known" Paleozoics are north of the Highlands, or found in valleys between hills of old gneiss within the Highlands. The rocks south of the Highlands were known as the Manhattan series, and Berkey in 1907 expressed doubt whether these could be Paleozoic rocks, and in 1922 stated that he would prefer to correlate them with the Precambrian, though he admitted that the question could not be settled from this section alone, and thought that further studies ought to be made in surrounding areas, to see whether or not there was a transition between the two groups.

But the idea that the Manhattan series was Precambrian gained ground, and when Miss Jonas and Mrs. Knopf set up the Glenarm Series (Jonas and Knopf, 1923; Jonas, 1924), the correlation was made thus:

(Pa. Piedmont)	(New York City)
Wissahickon Schist	Manhattan Schist
Cockeysville Marb.	Inwood Marb.
Setters Quartzite	Lowerre Quartzite
Baltimore Gneiss	Fordham Gneiss

And the reasons for separating the New York rocks from the "known" Paleozoics farther north were exactly the same as those stated in the Philadelphia folio for separating the Wissahickon and Octoraro.

1. The much higher grade of metamorphism.

2. The "igneous unconformity"--i. e., the Manhattan Schist is invaded by igneous rocks that do not cut the "known" Paleozoics.

And this interpretation was not seriously challenged until after 1925, when Robert Balk began his work in Dutchess County. No doubt Balk would have preferred to start right on the Manhattan Schist, which is exposed in Central Park in the middle of New York City--but New York City is so densely built up for so many miles that he had to go farther out to find enough outcrops to make any sort of geologic map. He had, of course, Berkey's report on the aqueduct section, but there, as I have said, the known Paleozoics and the Manhattan Schist are everywhere separated by blocks of the Highlands Gneiss--though at one place, near Peekskill, the two series are less than 2 miles apart.

So Balk (1936b) chose Dutchess Co., which adjoins the Connecticut line, where the Housatonic Highlands swing west into the Hudson Highlands. And here it is possible to pass right across the Highlands, keeping to the valleys always on the pelitic and calcareous rocks, from western Connecticut where the pelitic rocks are sillimanite schist, apparently identical with, and supposed to be continuous with the Manhattan Schist, across successively lower grades of metamorphism, out into the valley at Poughkeepsie, where one is on the Hudson River slate of Ordovician age.

Balk mapped a large area in a general way, and mapped in detail the Clove quadrangle and part of the Carmel quadrangle. He claimed there is a complete transition. Along the Hudson, say between Beacon and Wappinger Falls, the Hudson River rocks are chlorite slate, though in places much deformed and filled with innumerable quartz veins.

About 10-12 miles to the east, biotite appears, and Balk drew a biotite isograd, followed 2 or 3 miles farther east by a garnet isograd. (And note that it would be presumably about here that one would have to put the contact between the Hudson River slate and the biotite-garnet schist, if one were trying to separate them as two formations of different ages). From here eastward both schist and marble become more highly metamorphosed. Staurolite and kyanite appear locally (in rocks of suitable chemical composition), and Balk drew the sillimanite isograd a mile or two west of the Connecticut line. But even beyond this, metamorphism still increases, the rocks becoming coarser in grain. Within a few miles the slaty cleavage is entirely obliterated by a coarse foliation in the schist, and the marble becomes large-grained and highly crystalline.

Balk worked on this study for 8 or 9 years, and word of his conclusions "leaked out" long before he published them. And this roused up Mrs. Knopf, who was then living in New Haven, and had begun to interest herself in the geology of Connecticut and Massachusetts. So a controversy began, which for some years was bubbling away "below the surface," and that makes it rather difficult to follow the arguments if one reads the various papers by Mrs. Knopf (1927, 1931, and 1935), Prindle (Prindle and Knopf, 1932), and Agar (1932) in chronological order, between 1927 and 1935. Many of these are really directed at Balk, and are meant to anticipate points which he made in print only in 1932, many of them not until 1936.

Mrs. Knopf, of course was not prepared to accept the correlation of the Manhattan series with the Cambro-Ordovician rocks of the Hudson Valley, nor even the correlation of the Manhattan [series] with the schist of western Connecticut and Massachusetts, since that schist, the "Berkshire" Schist, had also been presumed to be Ordovician, correlated thus:

Hudson Valley		W. Conn. and Mass.
O	Hudson River Sl.	Berkshire Schist
e-O	"Wappinger Ls. (sub.div.)	"Stockbridge" Ls. (sub.div.)
e	Poughquag Quartzite Highlands Gneiss	Cheshire Quartzite gneiss

The reason she got involved was really because the geology of that area is in some way comparable to that of the Pa.-Md. Piedmont, where the Martic controversy was going on. There had for years been a school of thought that proposed a large overthrust--the Taconic Overthrust--that had carried more highly metamorphosed schists and gneisses, presumably of Precambrian age, westward from the front of the Taconics in Connecticut, the Berkshires in Massachusetts, and the Green Mountains in New York, out over the eastern margin of Paleozoic rocks in the Hudson-Champlain Valley. And Mrs. Knopf had begun doing field work in the Taconics. Like other workers before her, and like her contemporaries who were working farther north in the Berkshires, she found high to medium grade schists on the east side of the range grading into low grade phyllites on the west--and she claimed that the phyllites were diaphthonites--indications of retrogressive metamorphism resulting from intense stress during movement on the Taconic or other westward thrusts. And she declared that the transition that Balk had found between high and low grade rocks in Dutchess Co., was of a similar character and origin--retrogressive rather than progressive regional metamorphism.

In 1933, the International Geological Congress met in the U.S., and the Guidebook for the Hudson Valley excursion was prepared under the direction of the "Yale school", Longwell being chief editor of this section. Therefore the accompanying maps show the "Knopf" interpretation, with many overthrusts, even though Balk's preliminary conclusions had already been published (1932). Balk was not invited to contribute anything to this excursion, although he did lead the excursion to the Adirondacks.

By 1935 Mrs. Knopf had got extremely annoyed with Balk, and her paper called "Recognition of overthrusts in metamorphic terranes" (Knopf, 1935) [was] so patently aimed at him that he felt it necessary to write a brief reply (Balk, 1936a), saying that his Dutchess County study was in process of publication, and that further argument might well be postponed until others could study his results. The full paper came out in 1936; Tom Barth collaborated with him on the petrographic and petrologic work, and it is a valuable contribution to the study of regional metamorphism, even if one does not accept its conclusions on the age of the Manhattan Schist. Balk soon went to teach at Chicago, but he continued to work in New York State, advancing slowly northward along the Taconic front, making careful structural studies and using petrofabric methods on the rocks alleged to be part of the Taconic overthrust. But he died in 1955. The Taconic contro-

versy is not yet settled either, though a good deal more work has been done on it, which I will not attempt to review.

I have said enough, I hope, to show why what was going on up there did have some effect, indirectly at least, on the thinking of those who were working on the Martic controversy down here. And before I go back to Pennsylvania, I must mention very briefly another line of work that also affected interpretations here.

This was the mapping being pushed westward in New Hampshire. The leader here was another young man, Marland K. Billings (and he too had taught here at Bryn Mawr for two years, in the interval between Miss Bascom's departure in 1928 and the arrival of Watson and Dryden in 1930--Mrs. Billings was a Bryn Mawr geologist, Katharine Fowler).

But Billings always had his heart in the Highland of New England, and when he went back to Harvard in 1930 he began to put his students (and any young and willing colleagues) to work making a new geological survey of New Hampshire--advancing quadrangle by quadrangle through the Ossipee Mountains, the Franconias, the Presidential range of the White Mountains to the Connecticut River (by now [1964] they are getting over into Vermont).

This is a large area of crystalline rocks--slates and schists, marbles and quartzites, intruded by a great variety of abyssal and hypabyssal igneous rocks. In a long series of five studies Billings and his co-workers have discussed the various magma series and the differentiation of the igneous rocks, and have also studied in great detail the regional metamorphism of the sedimentary rocks.

I won't pretend to review any of this--but I must note that, so far as thinking about the Martic controversy was concerned, there were three points that had some effect down here:

1. The metamorphic rocks, some of them high grade schists and gneisses, are of Paleozoic age--in New Hampshire mostly upper Paleozoic (in Vermont mostly lower Paleozoic). This was established partly by fossil evidence--the most famous find being a brachiopod (Spirifer) in sillimanite schist. This emphasized what had gradually been realized anyway--that "grade" of metamorphism is not any real guide to the age of rocks--the Glenarm, though highly metamorphosed, was not necessary Precambrian for that reason alone.

2. The structures are complex and include a number of large overthrusts--both eastward and westward thrusts have occurred at different times. But these thrusts are readily recognized by the usual criteria--cuttings out or repetition of stratigraphic units, zones of mylonite or retrogressive metamorphism plainly localized on the thrust soles, etc. But there is no regional phyllonitization, such as had been postulated for the Martic Thrust.

3. The period of metamorphism was mostly Taconic (post Ord.) or Acadian (Dev.), and the Appalachian revolution had apparently little effect. This, too, had been known for a good while--the general idea being that the Appalachian was the important revolution in Pennsylvania (and the earlier ones negligible), but that the Appalachian folding simply dies out to the north.

In any case, it is a long way from the White Mountains to Philadelphia (or even to New York City), and conditions of sedimentation and orogeny probably were not the same throughout the whole region. Nevertheless, workers in Maryland and Pennsylvania had these points to think about as they struggled with their own problems.

So now I come back to Ernst Cloos. He, too, had begun work on the igneous rocks around Baltimore, but soon "branched out" to study the metamorphic sediments as well. Anyone who wants to understand the problems of the Piedmont thoroughly should read the Maryland Survey volumes and the other publications which Cloos has been producing--in conjunction with his students and colleagues--for some 30 years now.

Today I have time to mention only a small part of this work and in particular the G.S.A. Special Paper 35, Geology of the "Martic Overthrust" and the Glenarm Series in Pennsylvania and Maryland (Cloos and Hietanen, 1941). Anna Hietanen came to Bryn Mawr from Finland in 1938, worked here for a year and then went to Baltimore to work with Cloos. She had studied under Eskola at Helsinki, and had been trained in the "Sander (1930) methods" of petrofabrics, which were just coming into use in this country. Both she and Cloos made a large number of fabric diagrams in connection with this study of the Martic Overthrust. This report contains a great amount of detail, and is divided into several sections, both geographically and because different parts of the work were done by different people--so it is not always easy to follow the main arguments--so I will summarize now only the chief points in the parts relating to the Martic Thrust in Pennsylvania.

Cloos remapped, in great detail, the "key" area lying between the western end of Mine Ridge and the Susquehanna (Figure II-5). This includes the "type section" for the Martic Overthrust (Martic Hills, Martic Forge). As you all know, Cloos decided that there is no Martic Thrust, and I will come back to his evidence on that later. But if "getting rid of" the Martic Thrust might seem to make the general structural relations simpler, Cloos's mapping of the Paleozoic sediments shows that they are really very much more complicated than Stose and Jonas had thought. The column here is:

Wissahickon Schist
Conestoga (limestone & shale)
Ledger Dolomite
Kinzers (shale and limestone)
Vintage Dolomite
Antietam Quartzite
Harpers Schist
Chickies Quartzite
Hellam Conglomerate
~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~
Precambrian

So as we go west or northwest, we should expect to get these in order. The sequence Antietam-(and perhaps Harpers)-Vintage-Conestoga is repeated 4 or 5 times between the Mine Ridge and the Susquehanna River.

Jonas and Knopf had interpreted this area as one of repeated small folds, the Antietam being exposed in the crests of anticlines; but in that case the Antietam should be surrounded by Vintage, and then Conestoga. Cloos's mapping shows that the Vintage appears only on the west and south and is missing on the northeast side of these areas of Antietam.

This might be explained in two ways:

1. Perhaps this sequence ss, dol, ls was repeated several times during deposition. This does not seem very likely, especially since such repetitions do not occur anywhere else in the region.

So Cloos accepts the second possibility:

2. Repeated faults toward north or northeast; moreover, small thrusts in this direction can be seen in some field exposures. But these thrust sheets have obviously been folded (note curved shape of outcrops). So Cloos supposed that this thrusting took place very early in the history of the rocks--before the main period of folding (and regional metamorphism) which affects all the "known" Paleozoics (and also the Wissahickon Schist).

Another important part of Cloos's work is the structural analysis, on both a megascopic and microscopic scale--and here he was able to demonstrate that the major folding in this region is overturned toward the south or southeast. This, as he points out, is suggested on a large scale by the outcrop pattern of the Mine Ridge-Honeybrook uplift (Figure II-4). The Cambrian quartzite dips fairly steeply into Chester Valley, forming a narrow band. It broadens out around the plunging west end, and is much more gently dipping (broad band) on the north limb. This general impression was confirmed by plotting many hundred field measurements of folds, fold axes, lineations and joints; and by the direction of rotations observed in the fabric diagrams of all the different kinds of rocks involved. (We may note in passing
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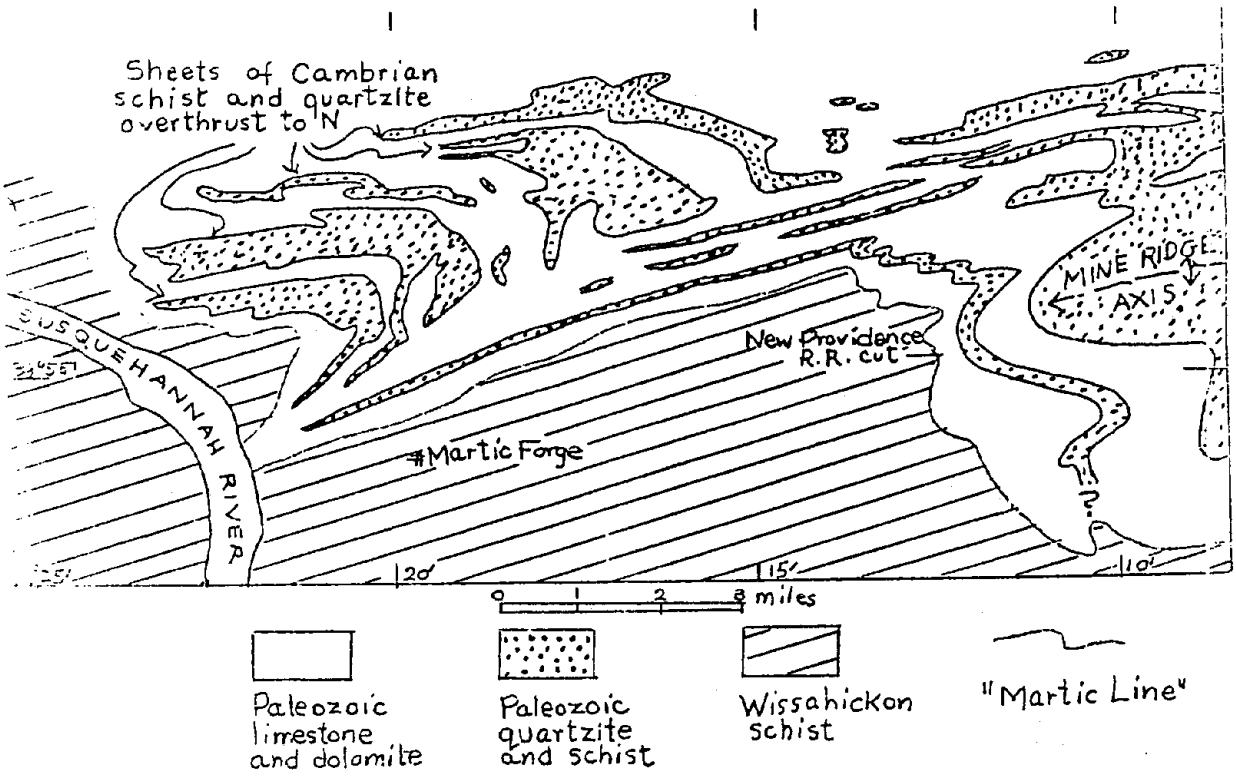

Figure II-5. Part of Ernst Cloos's geologic map of the "Martic Overthrust" area

that this is just the opposite direction from the overturning of folds in the Wissahickon farther east in the Philadelphia area.

As to the age of the Wissahickon Formation, Cloos was pretty well convinced that it is not Precambrian, and his main points were these:

1. All the structural elements have the same general character and the same symmetry in both Wissahickon and "known" Paleozoics. Mrs. Stose had made quite a point of the fact that the Wissahickon has "one more cleavage" than the Paleozoics, but Cloos's structure data do not confirm this. It is true that the cleavages may be better developed in some rocks than in others, depending on the physical properties of the rock--calcareous, quartzitic or micaceous. But Cloos does not believe that the Wissahickon has any "relict" structure inherited from a Precambrian past.

2. Therefore, he does not believe that the Wissahickon is a phyllonite--it shows no evidence of retrogressive metamorphism--or at least no more than is observed generally throughout the region--chloritization of biotite, epidotization of feldspars, etc. In fact, it is at the same grade of metamorphism as the adjacent Paleozoic rocks--in some cases almost indistinguishable in the field from the Antietam. (I will say more about the distribution of regional metamorphism in the whole area).

Cloos therefore concluded that the Martic Overthrust does not exist--the so-called "Martic line" is not a boundary between rocks having different structures or different grades of metamorphism, and there is no reason to suppose that the Wissahickon is older than the lower Paleozoic rocks. But he was rather vague about assigning a precise age to it. He rejected Miller's suggestion that it be correlated with the Martinsburg, and seemed to lean toward the notion that it might be Cambrian in age. And there is one curious, unresolved point on his map (southeast corner, Figure II-5). A long, narrow strip of Antietam follows the general course of the "Martic line", and then "looks as if" it might almost "join up" with the Wissahickon. (The actual distance at the point marked "?" is only a few hundred yards!) Cloos went no further with this suggestion in 1941, and indeed it is difficult to see just how the structure could be worked out if the Wissahickon is older than the Conestoga (especially if the Conestoga is really equivalent to the Cockeysville)--I will come back to this point again later, in reviewing more recent work done in Maryland on the Glenarm series.

Finally, there is one more general point that Cloos makes in his introduction to this paper. If the Glenarm rocks are really Paleozoic, they would represent the "core zone" or "root zone" of the whole Appalachian mountain system--a belt where the sediments have been most intensely folded, metamorphosed, and intruded by igneous rocks. And in all this it would be much more like other mountain ranges, such as the Variscan system in Europe; and the whole ratio of original length and breadth of the geosyncline, and the amount of shortening would be much more "normal," as judged by comparative studies of other geosynclines. If the whole wide belt of the Piedmont does not belong to the Paleozoic, then the Appalachian system has no igneous and metamorphic core, and the folded belt seems to have no easily explicable relation to a disproportionately large area of "basement rocks."

Cloos was undoubtedly influenced here by the theories of the German geologist Kober (1933) on the mechanism of mountain building, and it is probably not impossible to "explain" the relations of the folded Appalachians and the Piedmont in various ways. All the same, one may perhaps think again of New England, where the rocks in the central White Mountains, which have been deeply involved with regional metamorphism, igneous intrusions and granitization, are known to be of Paleozoic age. And if it is risky to try to correlate everything that happened in New England with what happened in Pennsylvania and Maryland, perhaps equally risky to assume that the whole mechanics of mountain building was completely different in the two areas--especially since it now appears that the main period here was not the "Appalachian Revolution" (at the end of the Paleozoic) but probably took place earlier--perhaps Taconic or Acadian (as in New England).

[Dorothy Wyckoff continued her lectures to cover the history of the Martic Line controversy from 1941 through 1964. Space does not permit their inclusion in this presentation. Those who

wish to read her sequel may do so by visiting the Science Library, Bryn Mawr College, or by examining the copy on file at the Pennsylvania Geological Survey in Harrisburg. A limited number of copies are available at cost of reproduction from W. A. Crawford.]

CONCLUSIONS

[The questions that follow are those of Dorothy Wyckoff as she finished her lectures in 1965.]

So we have still a number of unanswered questions: 1. Were there two major periods of deformation? And if so, why is the earlier one more obvious in the Baltimore region, and the later one along the Susquehanna? 2. The age of the Glenarm series: if we accept the earlier ages of some of the intrusives as late Precambrian or even Cambrian, the bottom of the Glenarm series, at least, must be Precambrian--though we might imagine that igneous activity accompanies the rise of the gneiss domes in the east while sedimentation was still going on in a trough farther west, so that Precambrian rocks merge westward into the base of the Paleozoic "proper." 3. If we have to accept a Precambrian age for at least the lower part of the Glenarm series, how do we explain, here in the Philadelphia region, its structural relations to the Paleozoic limestone in Chester Valley? Is there a Martic Thrust after all? Or is there a great lateral fault? In fact, are the structures north of Chester Valley really related to those south of it?

First Field Conference of Pennsylvania Geologists, 1931. Anna Jonas and Eleanora Bliss Knopf are seated first and second, respectively, to George Ashley's right.

III. POST-TACONIAN STRUCTURES OF THE WESTERN PIEDMONT PROVINCE OF PENNSYLVANIA: THE TUCQUAN ANTIFORM, THE LANCASTER VALLEY TECTONITE ZONE AND THE PEACH BOTTOM STRUCTURE

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INTRODUCTION

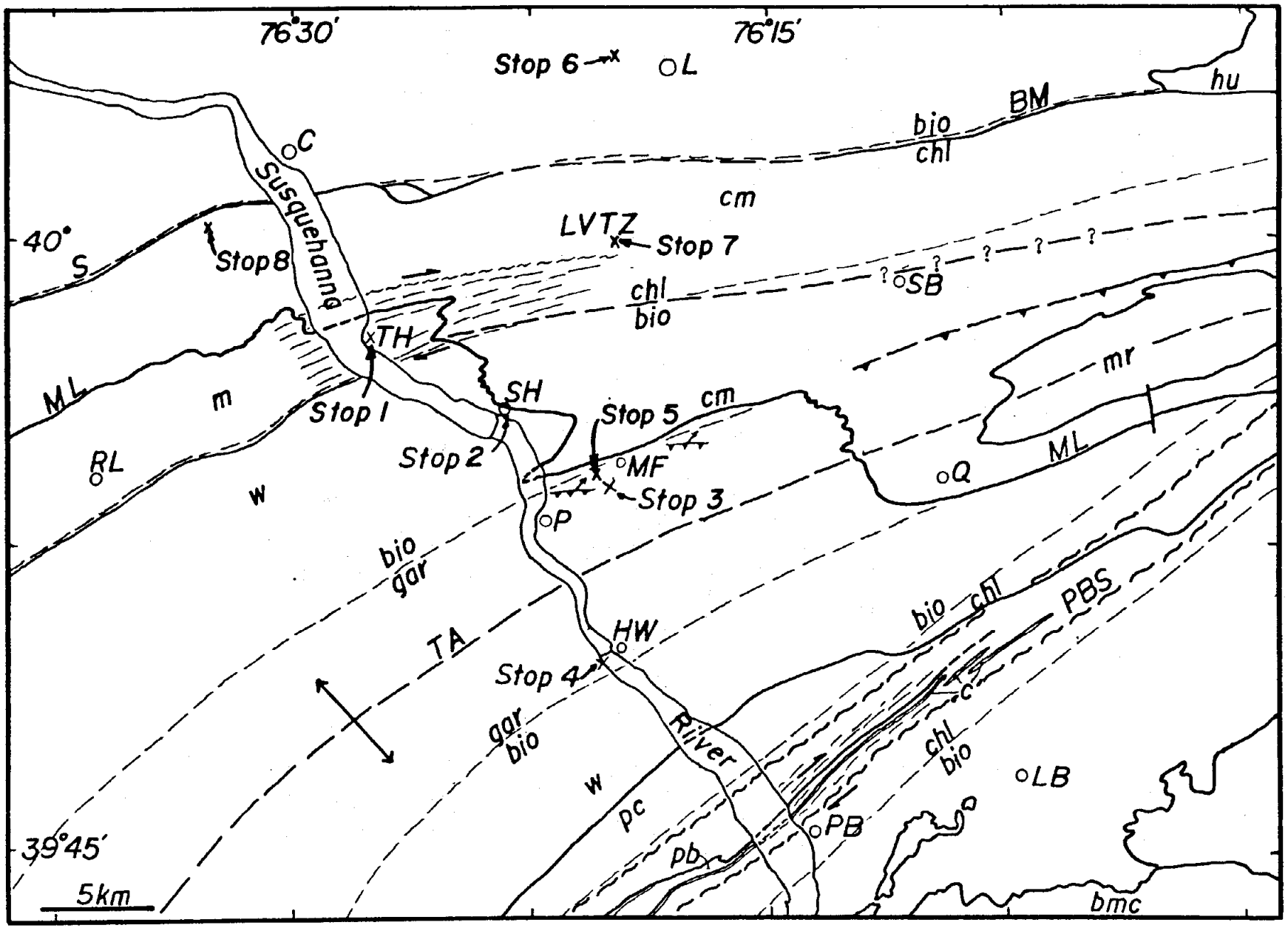
Thirty years ago Donald U. Wise, then at Franklin and Marshall College, presented his ideas on S-surface development and its relationship to various fold generations at the 25th Annual Field Conference of Pennsylvania Geologists held in Lancaster County. Since that time only a few papers have been published on the geology of the Lancaster area (Freedman and others, 1964; Lapham and Bassett, 1964; Wise, 1970). The Pennsylvania Geological Survey recently has begun investigations in the Piedmont province after more than 20 years of little or no geologic research in that area. Many new discoveries have resulted from these investigations. This paper focuses on the major post-Taconian structures of the western Piedmont including the Tucquan Antiform, the Lancaster Valley Tectonite Zone and the Peach Bottom Structure (Figure III-1).

THE TUCQUAN ANTIFORM

General Description

The rocks of the western Piedmont have a long history of study. Early work by Frazer (1880) with the second Pennsylvania Geological Survey entailed description of structure and lithology. Although the methods of study were technologically limited, his initial structure descriptions and documentation of the gross lithologic distribution were the foundation upon which future mapping in Lancaster County was based. Frazer (1880) gave the name "Tocquan Creek Anticlinal" (Figure III-2) to a structure he

Figure III-1 (facing page). Regional geological map showing major structures and metamorphic isograds. Towns: C=Columbia, L=Lancaster, RL=Red Lion, SB=Strasburg, MF=Martic Forge, SH=Safe Harbor, Q=Quarryville, HW=Holtwood, PB=Peach Bottom, LB=Little Britain; TH=Turkey Hill. Structure: S=Stoner fault, BM=Brandywine Manor fault, LVTZ=Lancaster Valley tectonite zone, ML=Martic line, TA=Tucquan antiform, PBS=Peach Bottom structure. Lithologies: cm=Conestoga marble, m=Marburg phyllite and schist, w=Wissahickon phyllite and schist, pc=Peters Creek schist and quartzite, pb=Peach Bottom slate and phyllite, c=Cardiff conglomeratic quartzite, bmc=Baltimore mafic complex. Grenvillian massifs: hu=Honey Brook Upland, mr=Mine Ridge. Metamorphic isograds: chl=chlorite, bio-biotite, gar=garnet.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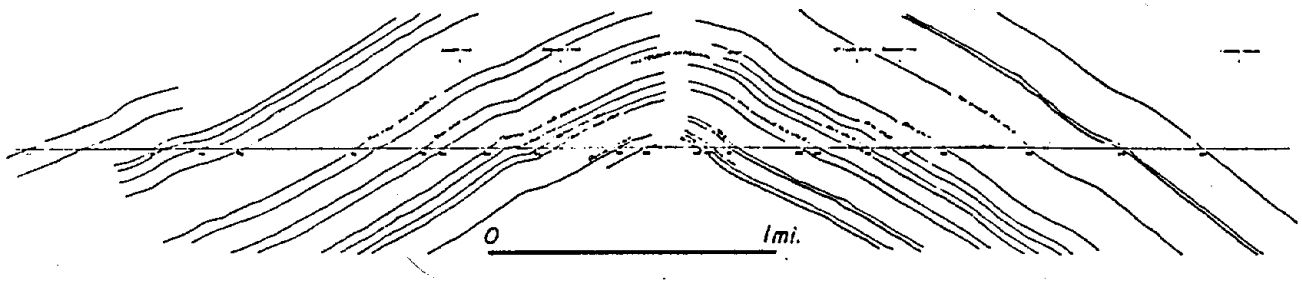


Figure III-2. Cross section of the "Tocquan Creek Anticlinal" after Frazer (1880)

described as "a great anticlinal, a stratigraphical feature so important and apparently so far reaching in effects." Knopf and Jonas (1929) redefined the antiformal structure as a double-crested open fold (the Pequea and Tucquan Anticlines), not overturned, with dips of 20° to 40° (Figure III-3) and they traced these folds across Lancaster County. Stose and Jonas (1939b) stated that the Tucquan Anticline is a double arched structure which is continuous with the Mine Ridge Anticline to the east. Later workers (Freedman and others, 1964) realized that the arched schistosity that defines the Tucquan structure is not sedimentary layering and renamed the fold the Tucquan Antiform. Freedman and others (1964) also recognized the correlation of regional D2 structures in the western Piedmont with the formation of the Tucquan Antiform and Wise (1970) proposed a kinematic model for the formation of the Tucquan Antiform that involved uplift of a "railroad tie" shaped basement block (the Mine Ridge Massif).

The axis of the Tucquan Antiform projected east of the Susquehanna River is approximately continuous with the Mine Ridge Antiform just north of Quarryville (Knopf and Jonas, 1929). Wise (1970) projected a single-crested Tucquan Antiform into the Mine Ridge Antiform; however, the Tucquan Antiform crest was projected about 2 kilometers south of the Mine Ridge Antiform axis due to nonparallelism of the regional schistosity with the sedimentary bedding which defines the Mine Ridge Antiform. West of the Susquehanna River and east of High Rock, the Tucquan Antiform has been mapped as a double-crested structure including the York Furnace Anticline of Knopf and Jonas (1929). At High Rock the two crests have been shown to merge, and a single crest traces southwestward into Harford County, Maryland (Figure III-3).

The oldest structure in the area, determined by cross-cutting relationships, is the S1 regional primary schistosity (Freedman and others, 1964). The metamorphic minerals defining

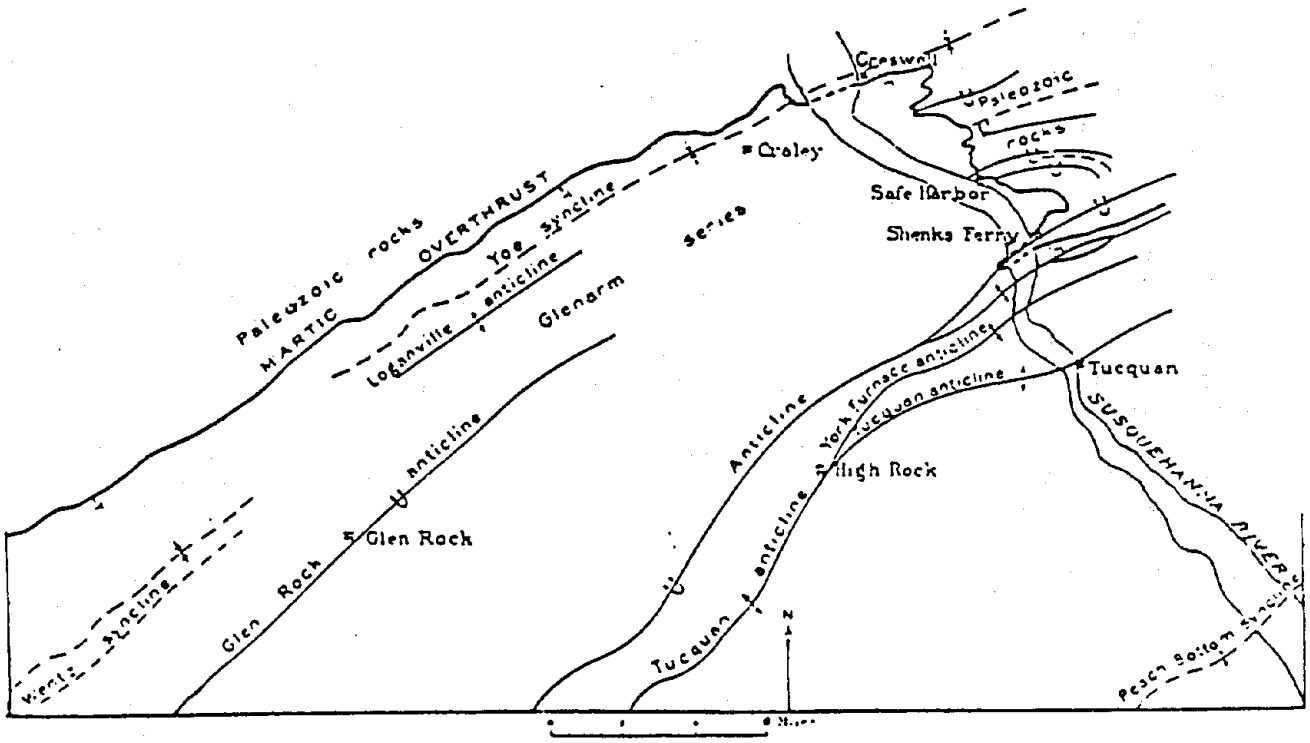


Figure III-3. Structure map of the western Piedmont after Knopf and Jonas (1929)

the S1 schistosity are associated with the first prograde episode (M1) of regional metamorphism (Faill and Valentino, 1989) and with the Taconian Orogeny (Freedman and others, 1964; Lapham and Basett, 1964). The Tucquan Antiform is defined by arched S1 schistosity (Freedman and others, 1964). The S1 schistosity is defined by parallel alignment of micas (muscovite and biotite), chlorite and chloritoid, as well as by planar aggregates of plagioclase, quartz and garnet. In most places compositional layering is parallel to the schistosity, as are layers of vein quartz. Isoclinal flow folds have axial planes parallel to the S1 schistosity with the hinge axes usually parallel to the strike of the schistosity (Freedman and others, 1964). These isoclinal flow folds, which range from millimeters to meters in amplitude, have thickened hinge areas with attenuated limbs that commonly are discontinuous. Freedman and others (1964) and Wise (1970) proposed a model of subhorizontal nappe emplacement to the northwest to explain the S1 schistosity and F1 isoclinal flow folds.

During the present study, metamorphic and structural petrology was correlated with structures observed in the field to develop a metamorphic and structural history for the formation of the Tucquan Antiform. The primary regional schistosity (S1) is deformed by regional D2 structures, some of which previously have been related to the formation of the Tucquan Antiform (Freedman and others, 1964). The D2 structures have been observed as

crenulations, discrete crenulation cleavage, internally penetrative schistosity (Valentino, 1989) and shear thrust fabrics. These structures have been correlated with a muscovite, chlorite and secondary biotite-producing episode of metamorphism associated with the formation of the Tucquan Antiform.

A detailed cross section (Figure III-2) of the Tucquan Antiform was first drawn by Frazer (1880). At the Susquehanna River the north limb of the antiform strikes 230° to 250° and dips 40° to 60° NW and the south limb strikes 040° to 060° and dips 50° to 70° SE. Frazer (1880) shows only one major crest for the Tocquan Creek Anticlinal at the Susquehanna River. A minor synformal structure dipping moderately southeast occurs on the north limb of the Tucquan Antiform in the area of Pequea, thus producing a minor antiformal structure adjacent to the north (Figure III-2). The York Furnace Anticline of Knopf and Jonas (1929) traces through the Pequea area (Figure III-3). The results of the present study concur with the concept of a single-crested Tucquan Antiform with a minor synform-antiform in the Pequea area as observed by Frazer (1880).

Wise (1970) showed that the Tucquan antiform narrows from about 27 kilometers wide at the Susquehanna River to about 8 kilometers wide at the western end of the Mine Ridge Grenvillian Massif. McCollough (1981) constructed a cross section of the Tucquan Antiform from data collected along the Patapsco River, Maryland, west of Baltimore. In this cross section the Tucquan Antiform is represented by an arch of S1 schistosity approximately 4.5 kilometers broad. The regional shape of the Tucquan Antiform suggests that the structure has a double plunge. However, the data presented by Wise (1970) show the northeastern end of the Tucquan Antiform to be plunging gently to the southwest.

The Tucquan Antiform Hinge Area

Figure III-4 is a contour plot of the poles to primary schistosity planes (S1) measured along the Susquehanna River. Immediately one can recognize the overall antiformal geometry plunging gently (<10°) in the direction of approximately 260°. The symmetry of the antiform suggests that the axial plane is subvertical. A plot of the schistosity from the hinge area shows a complex pattern of superimposed folds (Figure III-5). The contour of the plot of structural data from the northern limb of the Tucquan Antiform is continuous with the superimposed fold that trends approximately due west while the southern limb is continuous with the west-southwest trending fold (Figure III-6). Nonparallelism of the limbs and the superimposed fold geometry suggest that the Tucquan Antiform has a domal geometry.

There are numerous structures in the hinge area that developed during the second metamorphism, associated with the formation of the Tucquan Antiform. These structures have been divided into two categories: 1) subhorizontal ductile shear zones showing

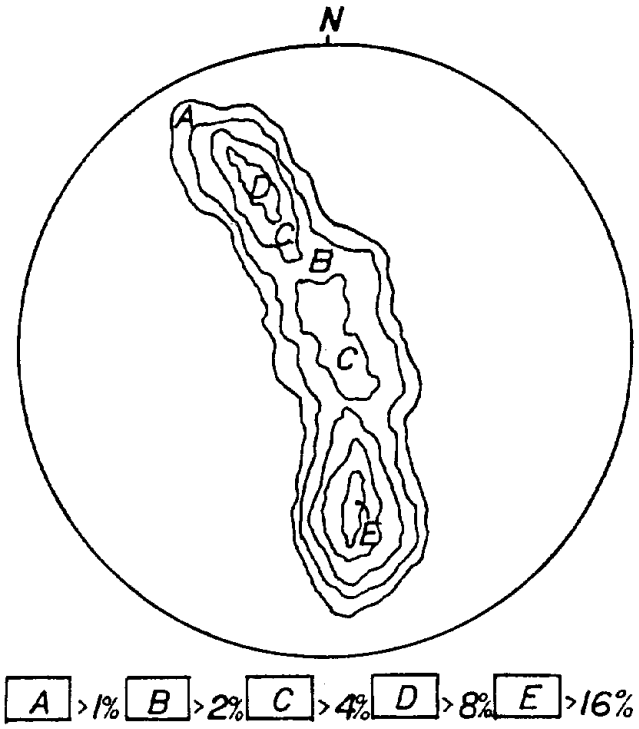


Figure III-4. Contoured lower hemisphere Schmidt net projection for the poles to S1 schistosity along the Susquehanna River transect, Lancaster County; the entire Tucquan antiform.

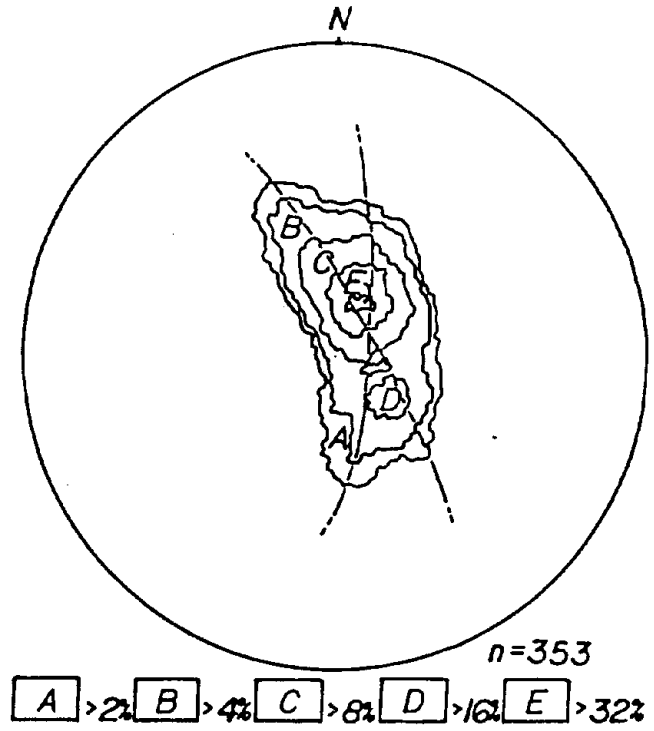


Figure III-5. Contoured lower hemisphere Schmidt net projection for the poles to S1 schistosity from the hinge area of the Tucquan antiform.

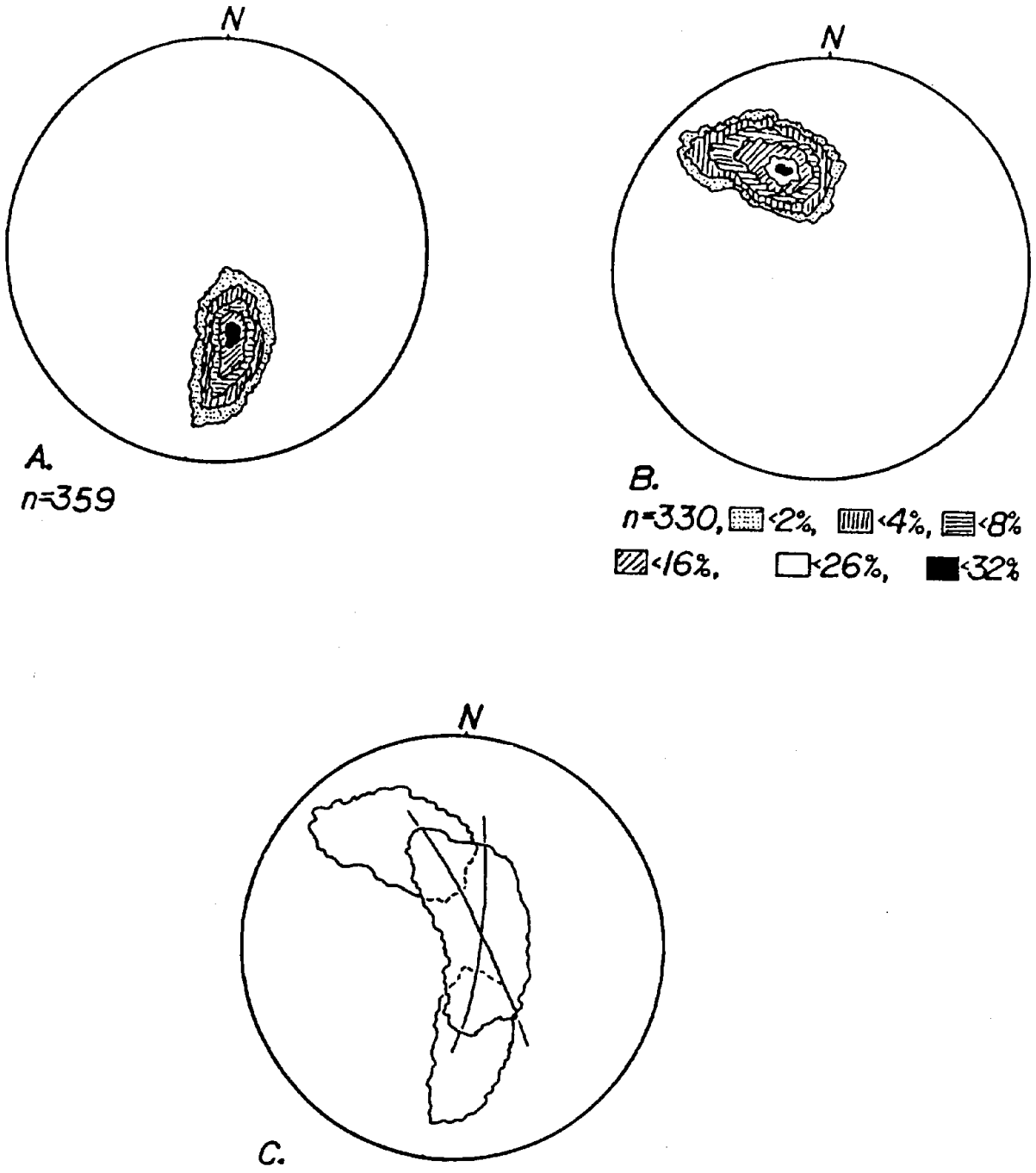


Figure III-6. Contoured lower hemisphere Schmidt net projections for A. the Tucquan antiform north limb, B. the south limb, and C. overlap of Figure III-5 with both A and B.

signs of biotite and chlorite recrystallization and 2) crenulation and crenulation cleavage with associated chlorite, muscovite and minor biotite recrystallization.

Thin sections from the Pequea area contain discrete shear surfaces parallel to the S1 schistosity and thin (millimeters

wide) shear zones cross-cutting the schistosity at a low angle, defined by weakly pleochroic biotite and recrystallized muscovite. These surfaces or thin shear zones (Figure III-7a) can be traced across the thin section (Figure III-7b) and usually are the only locations of secondary biotite in the rock. Muscovite and ilmenite, on the other hand, usually are distributed evenly throughout the rock. The reaction of ilmenite plus muscovite to produce biotite has occurred only along the discrete surfaces or in the thin shear zones, while the remaining ilmenite and muscovite in the rock is unaltered. The formation of biotite along discrete reactivation surfaces suggests that the biotite is secondary and most likely the result of localized reaction. Penetrative metamorphic processes would have allowed for a more evenly distributed reaction of ilmenite and muscovite to produce biotite.

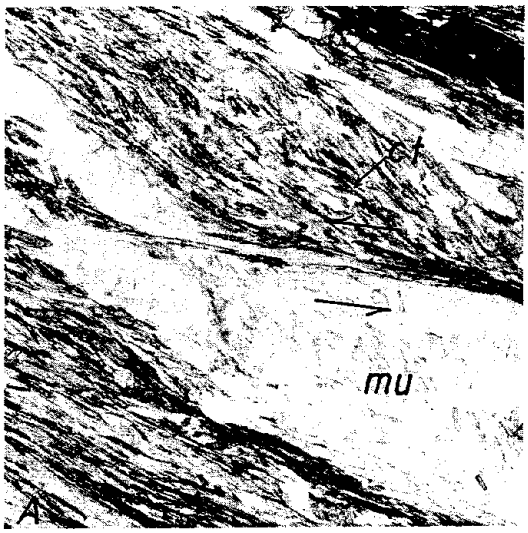
Variably developed crenulations in the antiform hinge area formed under conditions that allowed for the growth of new chlorite, muscovite and minor biotite. The S1 schistosity is crenulated with the trend of the hinge axes consistently to the northeast or southwest. Crenulations range in size from submillimeter to a few centimeters in amplitude and wavelength (Figure III-7c) and associated crenulation cleavage (S2) has an average orientation of 038° strike and 72° SE dip.

The Tucquan Antiform Limbs

The S1 schistosity steepens gradually away from the crest of the Tucquan Antiform. On the north side of the antiform the S1 schistosity strikes 240° to 260° and dips to the northwest (Figure III-6a). On the south side of the antiform the S1 schistosity strikes 050° to 070° and dips to the southeast (Figure III-6b). The angular increase in dip is approximately 7° per kilometer from the crest outward until about 75° is reached in the extreme northwest and southeast where D2 structures dominate the limbs. The overall width of the Tucquan Antiform is approximately 27.5 kilometers.

The D2 structures dominate the limbs of the Tucquan Antiform, as S2 penetrative schistosity, in the Turkey Hill area in the north and the Peach Bottom area in the south. This second deformation phase is characterized by strong penetrative foliation in the Conestoga Formation and northernmost Wissahickon Group on the north limb and Peach Bottom Formation and adjacent Peters Creek Formation on the south limb. The transition zones from S1 dominated rock to S2 dominated rock are as broad as 2 kilometers. The relative timing between S1 schistosity and S2 schistosity is easily determined in the field. The intersection of the two foliations forms a lineation which is diamond shaped in profile view. Truncation of the S1 schistosity at the S2 surface clearly indicates that S2 is later.

The S2 schistosity in the Wissahickon Group and Conestoga Formation strikes between 250° and 260°, and dips steeply between 75° and 90° to the northwest (Figure III-8a). Near the contact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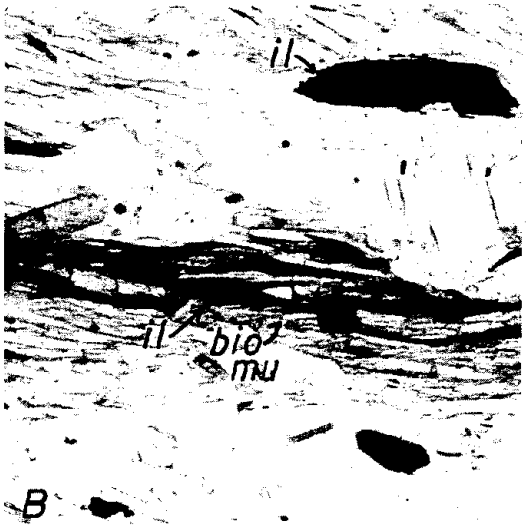
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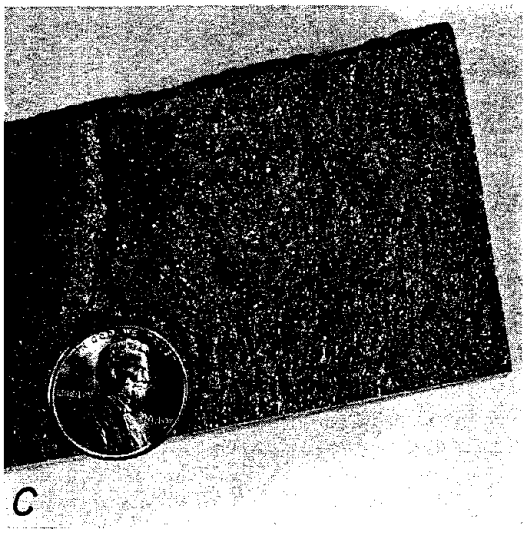
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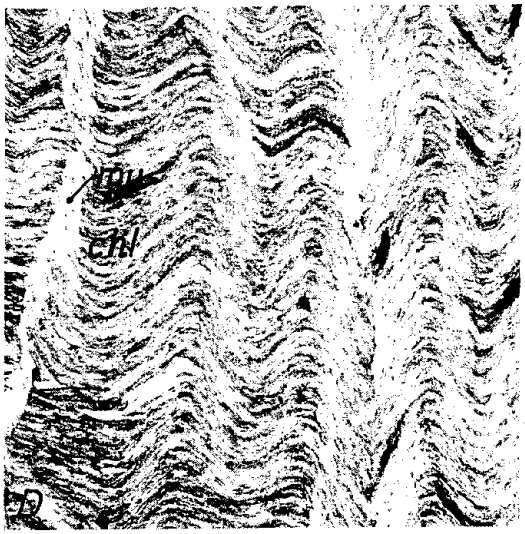
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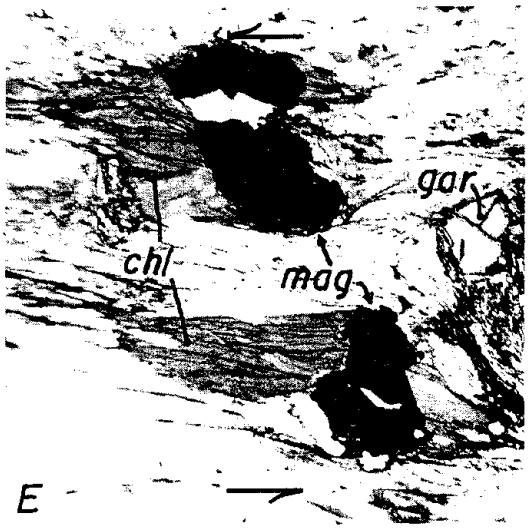
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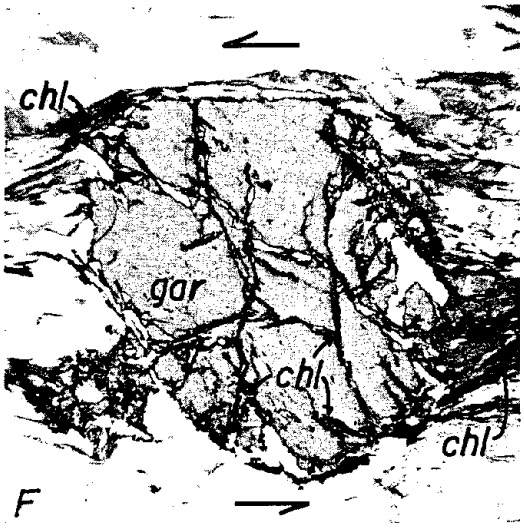
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between the Wissahickon Group and the Conestoga Formation the S2 foliation is penetrative and defined by the parallel alignment of second generation muscovite, chlorite and quartz crystals in the schist, and planar aggregates of calcite and phyllosilicates in the marble. Farther south in the Wissahickon Schist, S2 appears as moderately to weakly developed crenulation cleavage with new growth of chlorite in the hinge and muscovite on the limbs of the crenulations (Figure III-7d). Near Safe Harbor Dam (STOP 2) the S2 foliation rarely is observed. A complete gradation between internally penetrative S2 schistosity and widely spaced S2 cleavage exists between Turkey Hill and Safe Harbor.

The S2 schistosity in the Peach Bottom Formation and adjacent Peters Creek Formation strikes between 040°-050° and dips steeply 75° to 90° to the southeast (Figure III-8b). In the Peach Bottom ("slate") Formation the S2 schistosity is defined by parallel alignment of chlorite, muscovite and sericite. Metamorphic similarities between the S2 zones on the antiform limbs suggest that the muscovite and chlorite in the southern S2 zone also are second generation minerals. The Peach Bottom Formation is dominated by S2 while the foliation of the adjacent Peters Creek Formation varies from weakly developed crenulation cleavage to internally penetrative schistosity.

Taconian Isograd Distribution

Regional prograde isograds, Taconian in age (Lapham and Bassett, 1964; Wise, 1970), are distributed symmetrically about the hinge of the Tucquan Antiform (Hanscom, 1965; Faill and Valentino, 1989, 1990). The northern biotite-garnet isograd trends approximately parallel to the strike of S1 schistosity in

Figure III-7 (facing page). a. Thin shear zone in chloritoid-muscovite schist from the Pequea area; the view is looking southeast at a nearly vertical surface cut perpendicular to the schistosity; the field of view is 2.5 mm.

b. Reactivation surface in chlorite-muscovite-plagioclase schist from the Pequea area with reaction of muscovite and ilmenite to produce biotite; the view is looking southeast at a nearly vertical surface cut perpendicular to the schistosity; the field of view is 2.5 mm.

c. D2 crenulations on the S1 schistosity.

d. Photomicrograph of D2 crenulations with recrystallization of muscovite in the limbs and chlorite in the hinge areas; field of view is 2.5 mm.

e. Asymmetric chlorite pressure fringes on magnetite crystals indicating top-to-the-northeast thrusting; the view is looking southeast at a nearly vertical surface cut perpendicular to schistosity; field of view of 2.5 mm.

f. Asymmetric chlorite pressure fringes on garnet indicating top-to-the-northeast thrusting; the cracks within the garnet crystal are filled with retrograde chlorite; view is looking southeast at a nearly vertical surface cut perpendicular to schistosity; field of view is 2.5 mm.

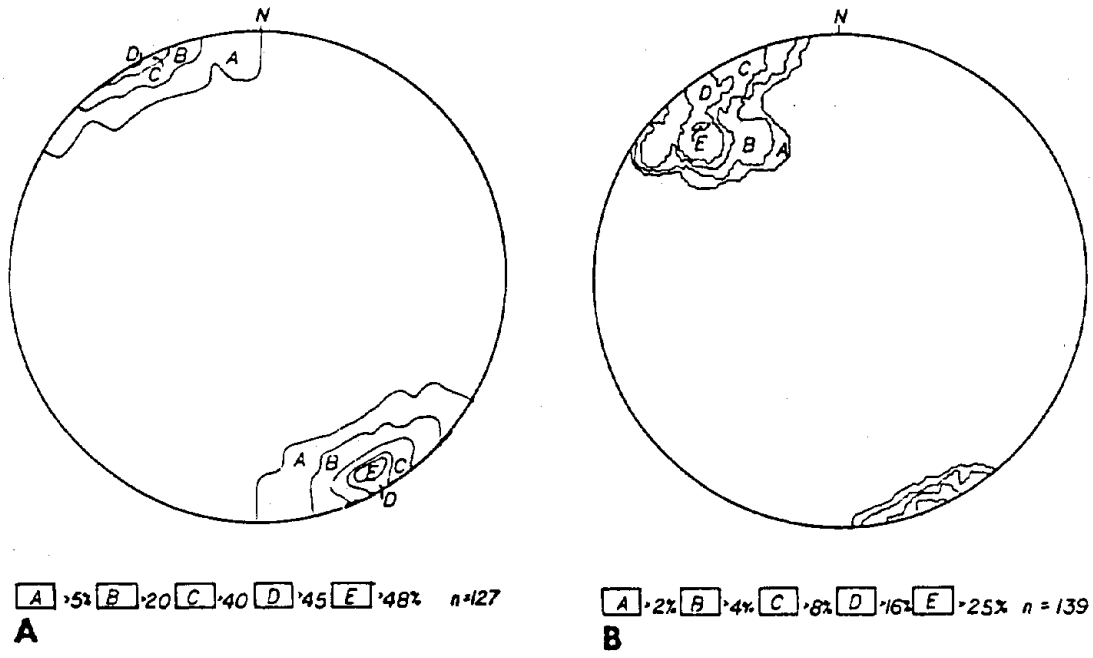


Figure III-8. a. Contoured lower hemisphere Schmidt net projection for the S2 schistosity on the north limb of the Tucquan antiform. b. Contoured lower hemisphere Schmidt net projection for the S2 schistosity on the south limb of the Tucquan antiform.

the area of Muddy Creek, York County and Martic Forge, Lancaster County. This isograd is approximately 4.9 kilometers north of the Tucquan antiformal hinge along Muddy Creek in York County (Hanscom, 1965) and approximately 4.0 kilometers north of the hinge axis at the Susquehanna River (Valentino and Faill, 1990). The southern biotite-garnet isograd also trends about parallel to the strike of S1 schistosity where it crosses Muddy Creek in York County and the Susquehanna River just south of the Holtwood Dam. This isograd is approximately 5.2 kilometers south of the hinge axis at Muddy Creek and approximately 4.8 kilometers south of the hinge axis at the Susquehanna River. The distribution of the biotite-garnet isograds about the Tucquan antiformal hinge suggests: 1) the biotite-garnet isograd surface is approximately parallel to the S1 schistosity, 2) the biotite-garnet isograd surface probably was connected over the crest of the antiform prior to erosion, and 3) differences in the isograd distance from the hinge area represents minor relief in the generally horizontal biotite-garnet isograd surface prior to the deformation that produced the Tucquan Antiform.

Microstructures Across the Tucquan Antiform

In the Pequea area discrete shear surfaces (Figure III-7b), thin subhorizontal shear zones (Figure III-7a) and asymmetric chlorite pressure fringes on plagioclase, magnetite and garnet

(Figure III-7e & 7f) indicate ductile subhorizontal shear directed toward 030°-040°. The pressure fringes commonly are retrograde after the host garnet or magnetite crystals (Figure III-7e & 7f) and the thin shear zones (Figure III-7a) contain recrystallized chlorite, muscovite and biotite. The growth of new chlorite at the expense of M1 garnet (Taconian) indicates that these microstructures are: 1) post-Taconian in age, associated with the second phase of metamorphism, and 2) cogenetic with other D2 structures in the area such as S2 regional schistosity and the Tucquan Antiform.

The direction of shear thrusting is oblique, approximately 20° to 30° counter-clockwise from the Tucquan antiformal hinge axis trend. If subhorizontal shear occurred prior to the antiform development, the trend of mineral lineations defining the direction of subhorizontal shear would systematically appear to rotate clockwise from south to north across the Pequea area. However, the direction of subhorizontal shearing varies non-systematically less than 10° across the Pequea area (Figure III-9) indicating that the subhorizontal shearing occurred after or in response to antiform development. It is interesting to note that Knopf and Jonas (1929) mapped a thrust fault on the north limb of the Mine Ridge Antiform approximately 12 kilometers east (Figure III-1) along strike of this zone of horizontal shearing in the Wissahickon Group rocks.

Secondary chlorite and muscovite recrystallization in the hinge and limbs, respectively, of D2 crenulations has occurred (Figure III-7d). Crenulations define a lineation on the S1 schistosity that trends 035°-050° on the north limb and 215°-230° on the south limb of the antiform. The axial planes of the crenulations are generally steeply dipping northwest or southeast. The orientation of the crenulation does not vary across the Tucquan Antiform, suggesting that these crenulations developed after the S1 schistosity was arched.

Crenulations are variably developed across the Tucquan Antiform with the strongest development in the northwest and southeast. In the extreme north and south the crenulation is so intense that crenulation cleavage and a new schistosity has developed (Valentino, 1989; Valentino, 1990) defined by secondary chlorite, muscovite and biotite. The Lancaster Valley Tectonite Zone (Valentino and MacLachlan, 1990) in the Lancaster-Columbia Synclinorium (Freedman and others, 1964) is dominated by the S2 schistosity, especially in the area of Turkey Hill where the S1 of the Tucquan Antiform north limb has been obliterated (Valentino, 1990). The Lancaster Valley Tectonite Zone strikes approximately 070°-080° and dips subvertically northwest. The Peach Bottom Structure located on the southern flank of the Tucquan Antiform also is dominated by the S2 schistosity. The Peach Bottom Structure strikes 040°-050° and is subvertical or steeply dipping to the southeast. These zones of S2 schistosity are equidistant from the hinge of the Tucquan Antiform (approximately 15 kilometers).

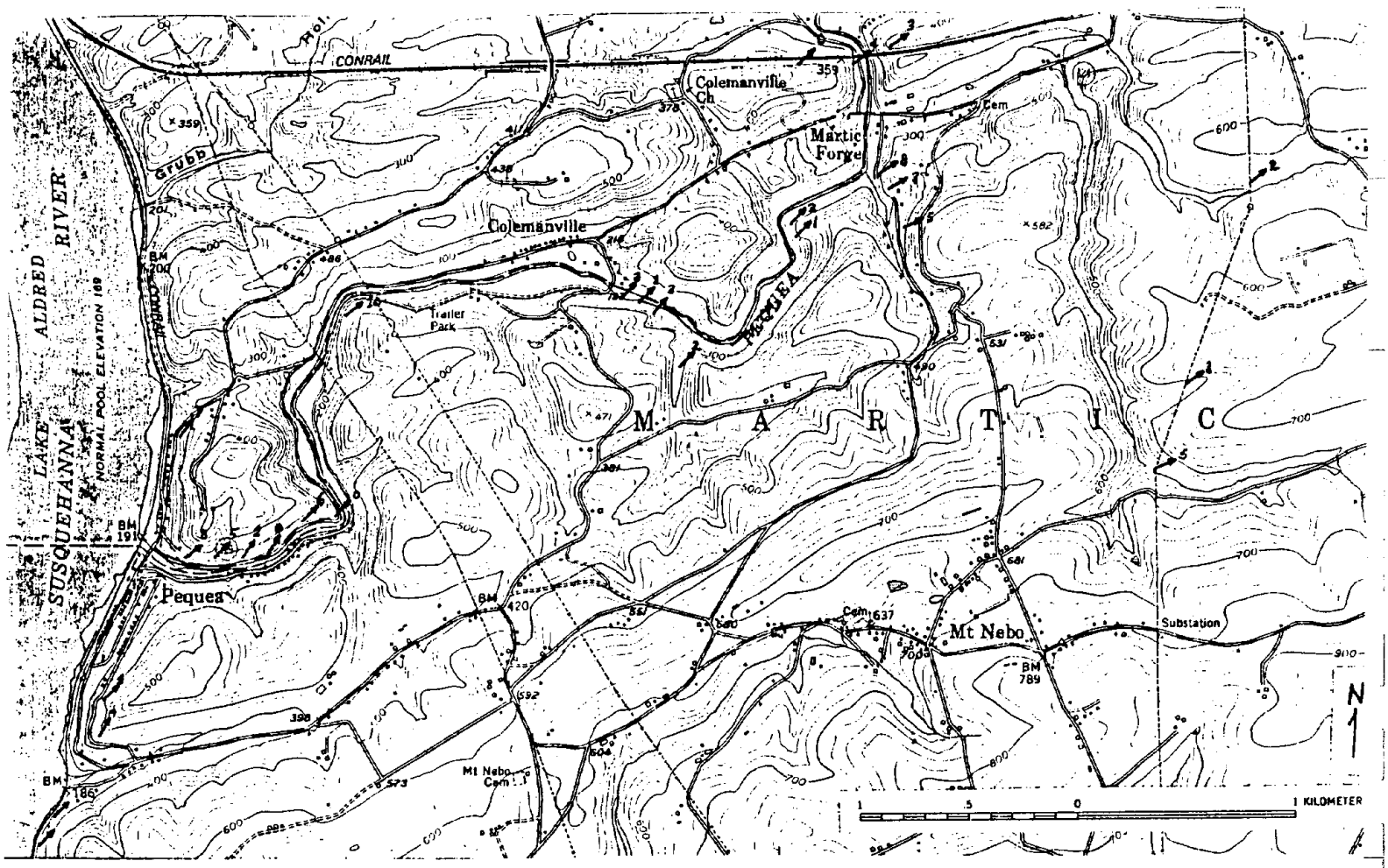


Figure III-9. Map of the Pequea area with the direction of offset associated with late shear thrusting (see text).

THE LANCASTER VALLEY TECTONITE ZONE

Definition

The Lancaster-Columbia Valley is largely underlain by marble of the Conestoga Formation with considerably smaller amounts of the Antietam (quartzite and schist), Vintage (dolomitic marble), Kinzers (dolomitic and calcitic marble and slate) and Ledger (dolomitic marble) Formations (Figure III-1). The southern border of the valley is defined by the contact between the Conestoga Formation and the schist lithologies of the Wissahickon Group.

Chickies Ridge, the type section for the Chickies Quartzite, marks the northern border of the valley.

Wise (1960) first recognized that the rocks of the Lancaster-Columbia Valley are host to a phase (D2) of extreme folding (F2) and cleavage/schistosity development (S2). Freedman and others(1964) categorized the phases of deformation in Lancaster County by detailed documentation of fold patterns at 22 study locations along the Susquehanna River. Although the earlier workers recognized that the distribution of D2 deformation was focused in the Lancaster-Columbia Synclinorium, they did not map the geographic distribution and intensity of the D2 deformation phase. The approximate boundaries or limits of D2 tectonized rocks (Figure III-1) recently have been delineated by Valentino and MacLachlan (1990).

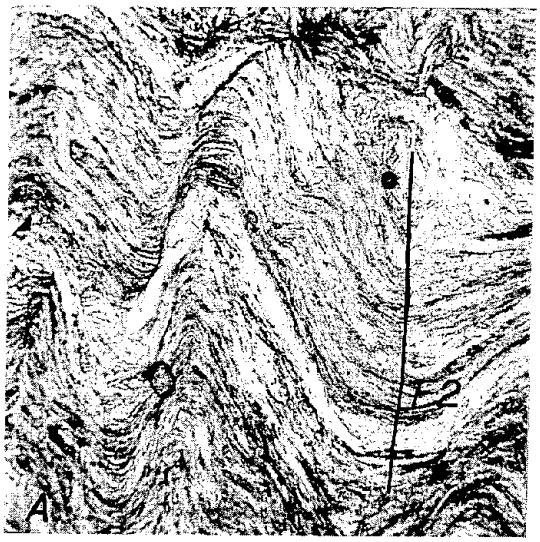
The Boundaries Of The Tectonite Zone

The Lancaster Valley Tectonite Zone lies within the Lancaster-Columbia Synclinorium (Freedman and others, 1964). Wise (1960) first described the deformation in the Conestoga Formation south of Lancaster at Williamson Park in the Guidebook for the 25th Annual Field Conference of Pennsylvania Geologists. Although numerous S-surfaces were recognized, the rock here is dominated by the S2 schistosity and F2 meso- and micro-folds. Just to the north, numerous quarries in the Conestoga Formation reveal relatively non-tectonized rock with foliation/bedding dipping moderate to steeply southeast. STOP 7 of the present field trip lies just within the northern boundary of the tectonite zone. This northern boundary aligns with the Brandywine Manor Fault to the east and possibly with the Stoner Fault to the west, and coincides with the chlorite-biotite isograd (Valentino and Faill, 1990).

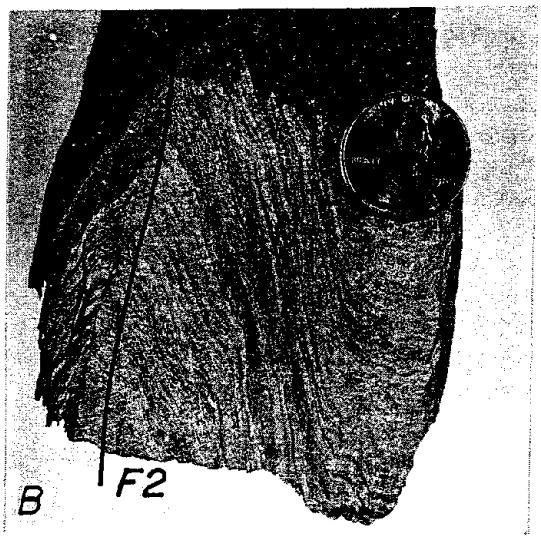
The southern boundary of the tectonite zone is defined by the occurrence of folded and cleaved marble and schist. Along the Little Conestoga Creek two exposures of Conestoga Formation clearly define the southern tectonite boundary in Conestoga marble. Over a distance of approximately 50 meters non-tectonized marble grades into marble dominated by the S2 cleavage. This locality lies approximately along the strike of the transition from S1 to S2 dominated schist in the Wissahickon-Marburg lithologies along the Susquehanna River. The M2 retrograde biotite-chlorite isograd coincides with the southern boundary of the tectonite zone (Valentino and Faill, 1990).

F2 Folds In The Tectonite Zone

Upright folds with gently east and west plunging axes can be found in just about every part of the tectonite zone. These folds generally have straight attenuated limbs and thick rounded hinge areas, and occur on the scale of a few millimeters in wavelength to a few kilometers (Figure III-10a, 10b & 10c). Although folding is a general characteristic of the tectonized rocks,



III-10A



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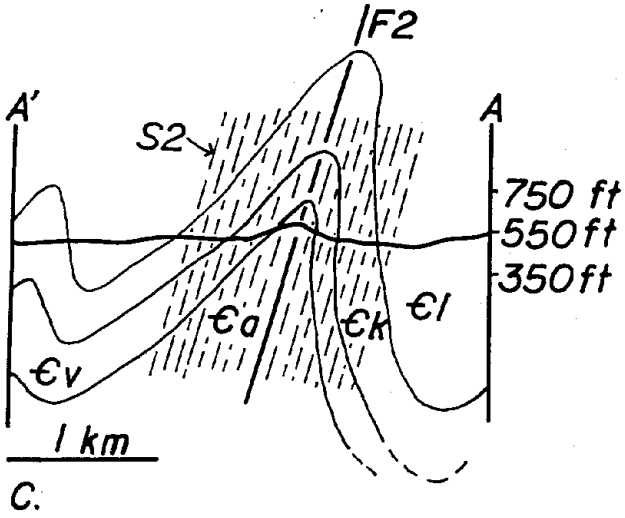
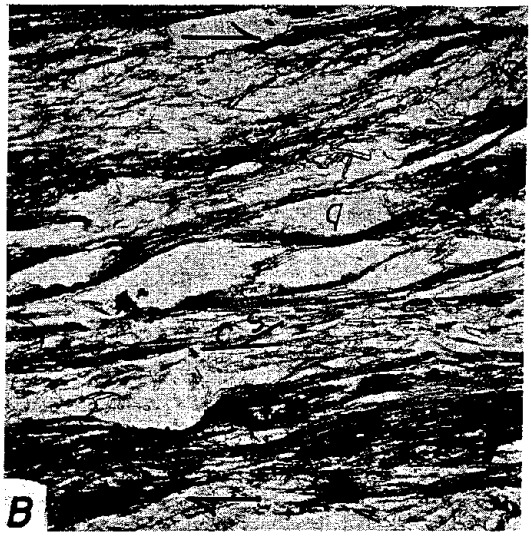


III-10C

III-15A



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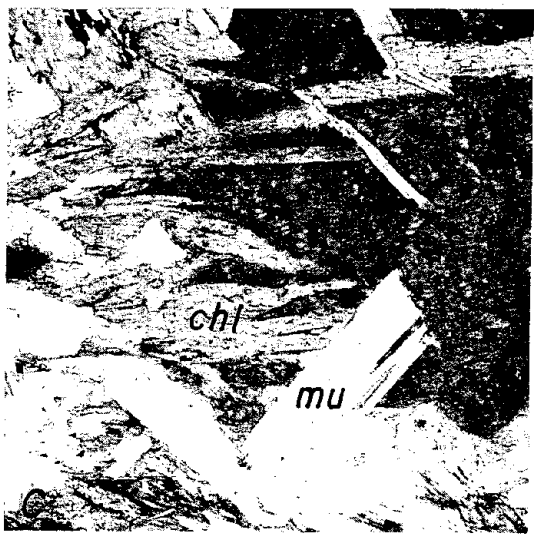


III-15C

the rocks along the southern part of the zone also have been deformed by strike-parallel shear resulting in penetrative schistosity development (Valentino 1989; Valentino, 1990). The northern part of the zone is generally characterized by F2 folds with much less S2 schistosity and cleavage development.

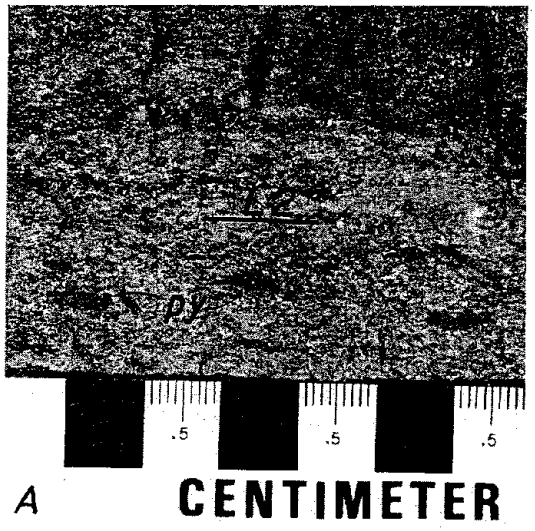
The orientation of these folds, upright with shallowly plunging hinge axes, suggests subhorizontal compression in a NNW and SSE direction. Estimations of the minimum percent shortening at sample localities have been made and the range of values is between 48% and 61% shortening perpendicular to the F2 fold axial planes with an average value of 56%. If a value of 56% minimum horizontal shortening is used to calculate the original width from the present (7.4 km), an original width value of 11.5 km is obtained. This estimated value suggests considerable collapse of the rock in the NNE-SSE direction during the D2 deformation phase.

Evidence of Dextral Shear in the Tectonite Zone

Subhorizontally oriented mineral lineations on the S2 schistosity surfaces are defined by elongate aggregates of pyrite (Figure III-11a) and quartz pressure-fringes on pyrite porphyroclasts (Figure III-11b). Steeply dipping penetrative schistosity (S2) with subhorizontally oriented extension lineations (L2) are consistent with a model of strike-parallel shearing. Near Turkey Hill, where S2 is penetrative in the Wissahickon-Marburg lithologies, strike-slip asymmetric quartz pressure-fringes (Figure III-11b) have been observed. Similar pyrite and quartz pressure-fringe microstructures have been observed from the outcrop belt along strike at the Conestoga River. Consistent strike-slip dextral motion was determined from the pyrite crystals with asymmetric quartz pressure-fringes (pyrite type:

Figure III-10 (facing page). a. Microscopic F2 folds/crenulations from the Marburg schist in the Lancaster Valley tectonite zone; field of view is 2.5 mm.
b. Mesoscopic F2 fold in Conestoga phyllitic marble from the Lancaster Valley tectonite zone.
c. Cross section of an F2 fold from the Lancaster Valley tectonite zone; see Figure III-13 for the line of the section.

Figure III-15 (facing page). a. Type I S-C mylonitic fabric in the Cardiff conglomeratic quartzite indicating dextral shear; view is looking down on a surface cut perpendicular to the S2 schistostiy and parallel to the L2 lineations; field of view is 2.5 mm.
b. Type I S-C mylonitic fabric in a sheared quartz vein from the Peach Bottom slate indicating dextral shear; view is looking down on a surface cut perpendicular to the S2 schistosity and parallel to the L2 lineations; field of view is 2.5 mm.
c. Photomicrograph of new growth of chlorite (M2) at the expense of primary (M1) biotite from the Safe Harbor area; field of view is 0.8 mm.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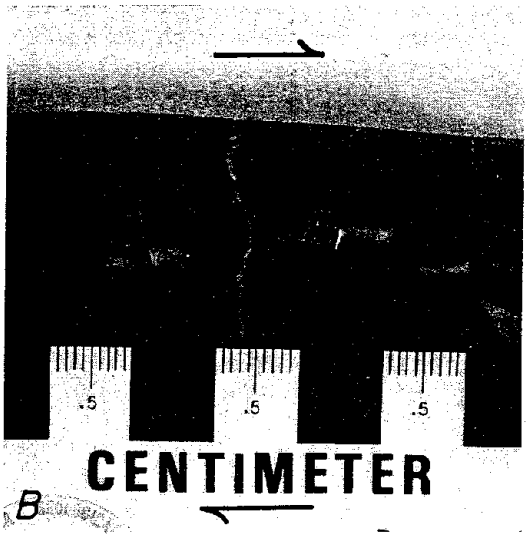
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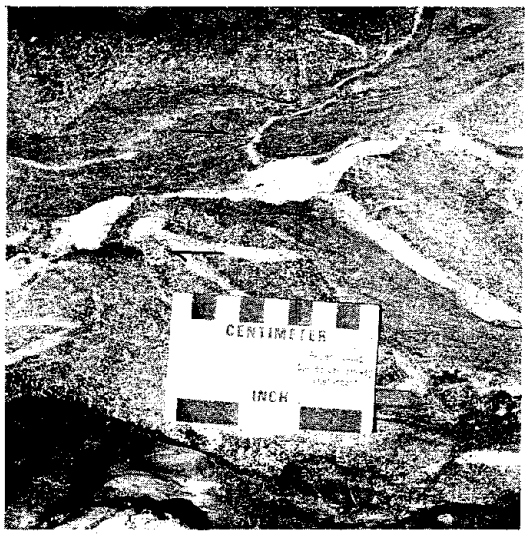
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Ramsay and Huber, 1983) that are best observed on surfaces cut perpendicular to foliation and parallel to lineations (subhorizontally oriented surfaces).

Competent layers of coarse crystalline marble in relatively more ductile phyllitic-marble matrix have formed boudins with fibrous quartz vein fill (Figure III-11c). On subhorizontally oriented outcrop surfaces these boudins often reveal quartz vein asymmetry that indicates dextral rotation of the boudins parallel to the S2 fabric. Cross-cutting veins of quartz also have been sheared dextrally parallel to the S2 schistosity (Figure III-11d).

An unusual linear prong of Wissahickon-Marburg rock extends into the Conestoga Formation to the east, along the strike of the zone of penetrative S2 schistosity near Turkey Hill. Microstructural analysis in this zone of penetrative S2 reveals consistent dextral offset. The geometry of the linear prong of Wissahickon-Marburg rock and the overlap with the zone of penetrative S2 schistosity suggests that the Wissahickon Group-Conestoga Formation contact (the Martic Line) has been locally transposed by dextral shear (Figure III-12).

Generally it appears that the evidence for strike-slip shear is restricted to the southern portion of the zone of severe deformation. The northern part of the tectonite zone may lack shear entirely and is characterized by horizontal shortening perpendicular to the tectonite zone boundaries.

Possible Sinistral Offset

The outcrop pattern of lower Paleozoic metasedimentary rocks in the area of the Brandywine Manor Fault suggests possible sinistral offset (Figure III-13). North of the fault and west of

Figure III-11 (facing page). a. L2 mineral lineations on the S2 schistosity from the Turkey Hill area defined by pyrite aggregates.

b. Polished slabs containing pyrite crystals with quartz pressure fringes associated with the S2 schistosity. These microsctructures define L2 mineral lineations and indicate dextral shear; view looking into the earth at a subhorizontal surface.

c. Dextrally rotated boudins of coarse crystalline marble in a phyllitic marble matrix (Conestoga Fm.).; view looking down on a subhorizontally oriented exposure surface

d. Dextrally sheared quartz vein in the conestoga Formation; view looking down on a subhorizontally oriented exposure surface.

e. Cardiff conglomeratic quartzite with elongate pebbles that define the L2 lineation.

f. Dextrally sheard quartz vein in Peters Creek lithology within the zone of penetrative S2 schistosity; view looking down on a subhorizontally oriented exposure surface.

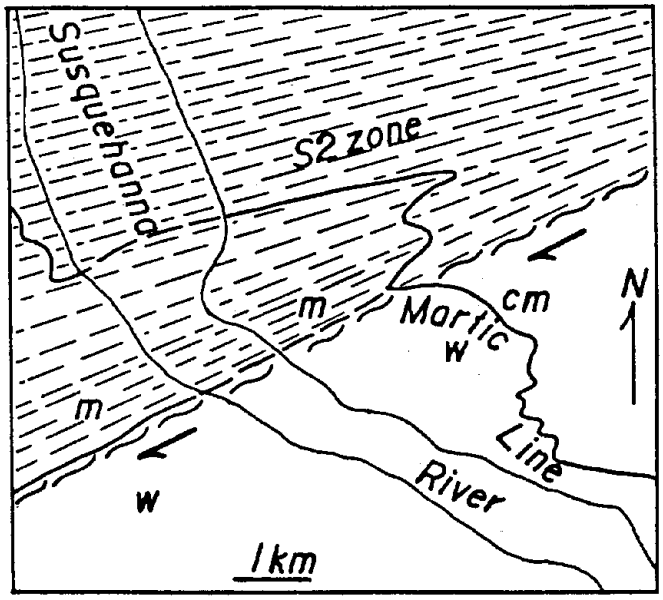


Figure III-12. Map of the Turkey Hill area including the Martic line and the zone of S2 schistosity (southern portion of the Lancaster Valley tectonite zone). Lithologies: cm=Conestoga marble, m=Marburg phyllite and schist, w=Wissahickon phyllite and schist.

the northern Honey Brook Upland Massif, the sequence of formations from west to east is as follows: Kinzers, Vintage, Antietam, Vintage and Antietam. South of the Brandywine Manor Fault the same lithologic sequence exists in the western end of the southern Honey Brook Upland Massif: Kinzers, Vintage, Antietam, Vintage and Antietam. The following list shows the horizontal width of the formations measured from the geologic map:

Formation	North of Fault	South of Fault
Kinzers	0.12 km	0.15 km
Vintage	1.75 km	1.60 km
Antietam	2.70 km	2.58 km
Vintage	2.58 km	2.30 km
Antietam	3.10 km	3.30 km

Reconstructing the sequence of lithologies suggests that sinistral displacement has taken place across the Brandywine Manor Fault; the magnitude of displacement is approximately 17 km. The same situation appears to exist across subordinate faults south of the Brandywine Manor Fault. The total sinistral offset across all three faults is approximately 19.7 km. Reconstruction across the faults produces the lithologic distribution of Figure III-14. The map-pattern fold geometry in the reconstruction is consistent with F2 folds found elsewhere.

Crawford and Hoersch (1984) proposed "scissors" type offset on the Brandywine Manor Fault to explain the juxtaposition of amphibolite facies gneiss of the southern Honey Brook Upland Massif

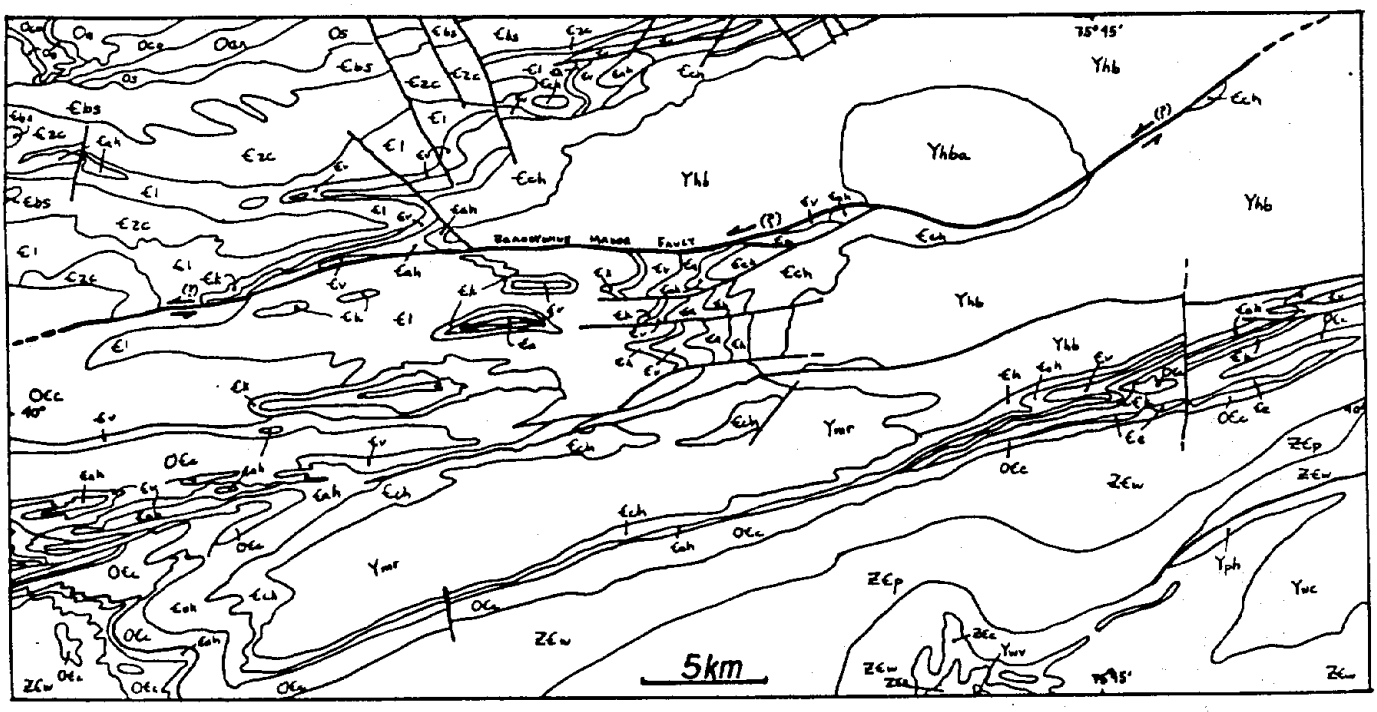


Figure III-13. Geologic map of the Brandywine Manor fault, the northern boundary of the Lancaster Valley tectonite zone.

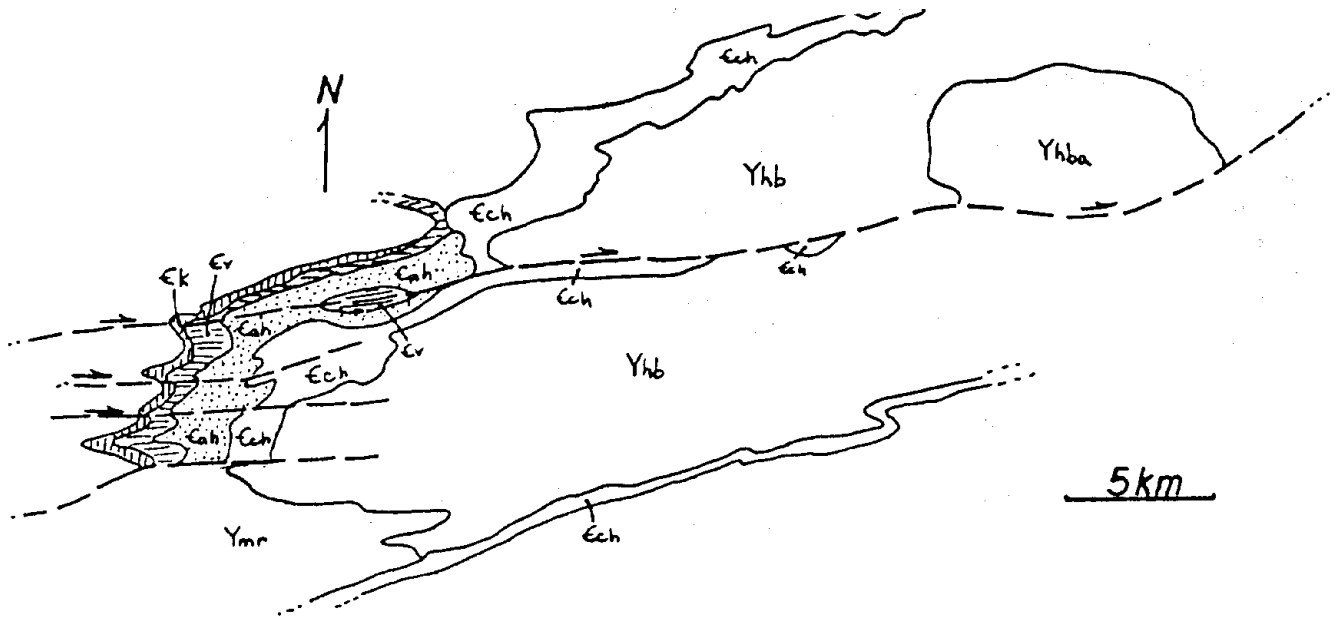


Figure III-14. Possible structural reconstruction of the Brandywine Manor fault.

with the granulite facies gneiss of the northern Honey Brook Upland Massif, with the northern block moving up relative to the southern. Offset of this nature across the Brandywine Manor Fault is not supported by the distribution of the lower Paleozoic metasediments as explained above.

THE PEACH BOTTOM STRUCTURE

General Description

Traditionally the Peach Bottom Structure has been defined as a syncline comprised of a core of black slate (the Peach Bottom Formation) with conglomeratic quartzite (the Cardiff Formation) and schist (the Peters Creek Formation) on the limbs (Knopf and Jonas, 1929; Stose and Jonas, 1939; Agron, 1950; Freedman and others, 1964; Wise, 1970). Higgins (1972) proposed that the structure is anticlinal based on preserved graded beds in the Peters Creek Formation that suggest the rocks south of the "syncline" are right-side-up where a synclinal interpretation would require that they be overturned. Recent mapping by the Pennsylvania Geological Survey has shown that the Peach Bottom Structure is not confined to the interpreted syncline, but has a width of approximately 5 kilometers at the Susquehanna River. This structure is characterized by a zone of S2 schistosity, F2 folds, abundant shear indicators, and M2 metamorphism, similar to the Lancaster Valley Tectonite Zone. The Peach Bottom Structure and Lancaster Valley Tectonite Zone are approximately equidistant from the Tucquan Antiform hinge axis, to the southeast and northwest respectively.

The Peach Bottom Structure Boundaries

The northern boundary of the Peach Bottom Structure is defined by the abrupt appearance of semi-penetrative S2 schistosity and F2 upright folds in the area just south of Drumore along the Susquehanna River (Figure III-1). The southern boundary is defined by the appearance of S2 schistosity and F2 folds in the area just north of the town of Peach Bottom, Lancaster County. Across both the northern and southern boundary the S1 regional schistosity is deformed by the development of S2 schistosity and F2 folds indicating a post-Taconian age for the Peach Bottom Structure.

F2 Folds And Dextral Shear In The Peach Bottom Structure

Evidence for subhorizontal compression and strike-slip deformation are present within the Peach Bottom Structure. The total width of the deformation zone is approximately 5 kilometers and includes the Peach Bottom slate belt. Most of the deformation is characterized by upright F2 folds and axial planar S2 schistosity that strikes 040°-050° and dips 70°-90° southeastward. The plunges of the F2 hinge axes are dependent on the original orientation of the S1 schistosity that was folded; however, the hinge axes generally plunge 05°-35° to the northeast. In

rock where F2 folds are well developed, estimates of minimum shortening perpendicular to the axial planes range from 40% to 70% If an average minimum shortening of 55% is integrated across the width of the deformation zone (about 5 kilometers wide), the conclusion is that the rock has been subhorizontally shortened a minimum of 6 kilometers.

Although the deformation zone is dominated by the F2 folds and subhorizontal compression, there also is evidence for strike-slip deformation parallel to the S2 schistosity. The zone of penetrative second schistosity (S2), approximately 1.5 kilometers broad, coincides with the Peach Bottom slate belt. This penetrative S2 also dominates the adjacent Cardiff conglomeratic quartzite and adjacent Peters Creek lithologies. The exposures of Cardiff conglomeratic quartzite in Lancaster County are dominated by elongate quartz pebbles (Figure III-11e). These deformed quartz pebbles define a subhorizontally oriented lineation that trends approximately parallel to the strike of the S2 schistosity. The combination of subhorizontal extension lineations and steeply dipping schistosity suggests a model of strike-parallel shearing along the S2 schistosity in this penetrative zone.

During recent mapping of the Peach Bottom Structure, numerous dextrally sheared quartz veins were observed at outcrops (Figure III-11f) in the Peters Creek Formation. Type I S-C mylonitic structures (Lister and Snoke, 1984) were observed in the Cardiff Formation that indicate dextral offset (Figure III-15a). In addition, dextral type I S-C mylonitic structures are developed in sheared vein quartz in the Peach Bottom Formation (Figure III-15b). The magnitude of displacement across this penetrative zone is unknown at this time; however the width of the zone (about 1.5 km) suggests considerable displacement.

Agron (1950) and Southwick (1969) mapped a fault on the north side of the Peach Bottom slate belt to explain the absence of Cardiff Formation. Freeman and others (1988) proposed dextral offset on this fault zone to explain the distribution and shape of ultramafic bodies to the north and west of the slate belt. Recent work by Krol and others (1990) has revealed a 2 kilometer broad zone of phyllonite in Harford County, Maryland, directly along strike with the zone of penetrative S2 in the Peach Bottom area. Krol and others (1990) proposed dextral offset across this zone based on microscopic kinematic analysis.

CORRELATION OF STRUCTURES WITH POST-TACONIAN METAMORPHISM

Metamorphic History

The Tucquan Antiform, Lancaster Valley Tectonite Zone and the Peach Bottom Structure have identical metamorphic histories. Thin subhorizontal shear zones and asymmetric shear structures, associated with northeast directed subhorizontal shear, are defined by secondary chlorite, muscovite and minor biotite (Figures III-7a, 7b, 7e & 7f). The F2 folds and crenulations found in the

Lancaster Valley Tectonite Zone and the Peach Bottom Structure are accompanied by chlorite and minor biotite recrystallization in the hinge area and muscovite recrystallization in the limbs (Figure III-7d). New muscovite and chlorite growth is restricted to the cleavage while the rock between cleavage surfaces generally remains unaltered. Primary M1 biotite in the Wissahickon Group shows signs of retrogression to chlorite in the area just south of Turkey Hill (Figure III-15c). The production of M2 chlorite at the expense of M1 biotite defines an isograd (Figure III-1) that coincides with the southern boundary of the Turkey Hill Shear Zone (Valentino, 1989; Valentino, 1990), the sheared southern portion of the Lancaster Valley Tectonite Zone. The Peach Bottom Shear Structure is located within a relatively narrow (3 km) chlorite grade zone (Figure III-1) that extends across the Piedmont (Faill and Valentino, 1989). Steep metamorphic gradients near this chlorite zone have been interpreted to be the result of the second episode of metamorphism.

The prograde regional metamorphism in the western Piedmont is interpreted to be Taconian (Lapham and Bassett, 1964). The second metamorphism associated with the Tucquan Antiform, Lancaster Valley Tectonite Zone and Peach Bottom Structure has overprinted the Taconian metamorphic minerals, indicating a post-Taconian age for the metamorphism and structures. Lapham and Bassett (1964) dated second generation muscovite and obtained an average age of 330 Ma. This date suggests that the (M2) metamorphism and (D2) structures are associated with early Alleghanian deformation. Dextral strike-slip shearing in conjunction with subhorizontal compression are consistent with the Alleghanian deformation style observed in the southern Appalachian Piedmont Province.

Structural Model

The early investigations by Frazer (1880), Knopf and Jonas (1929) and Stose and Jonas (1939) were primarily concerned with identification of structures and documentation of lithologies. The first model for the regional D2 structures was proposed by Freedman and others (1964). Basement uplift was held to be primarily responsible for the arching of the S1 schistosity to form the Tucquan Antiform and for the formation of the S2 crenulation cleavage. Wise (1970) constrained the uplifted basement to the shape of a "railroad tie." This conclusion was reached by the pattern of folded S1 schistosity over the Tucquan Antiform in Lancaster and York Counties. Thrusting along the northern margin of the Mine Ridge Anticline associated with the Tucquan Antiform development, and correlation of the doming of the Woodville Massif in Chester County with the D2 deformation phase also was proposed by Wise (1970).

The D2 structural models developed by Freedman and others (1964) and Wise (1970) concentrated on rock movement directions; however, a mechanism for rock movement was never addressed. The combination of compressive and strike-slip structural components, observed during this investigation, formed over a relatively

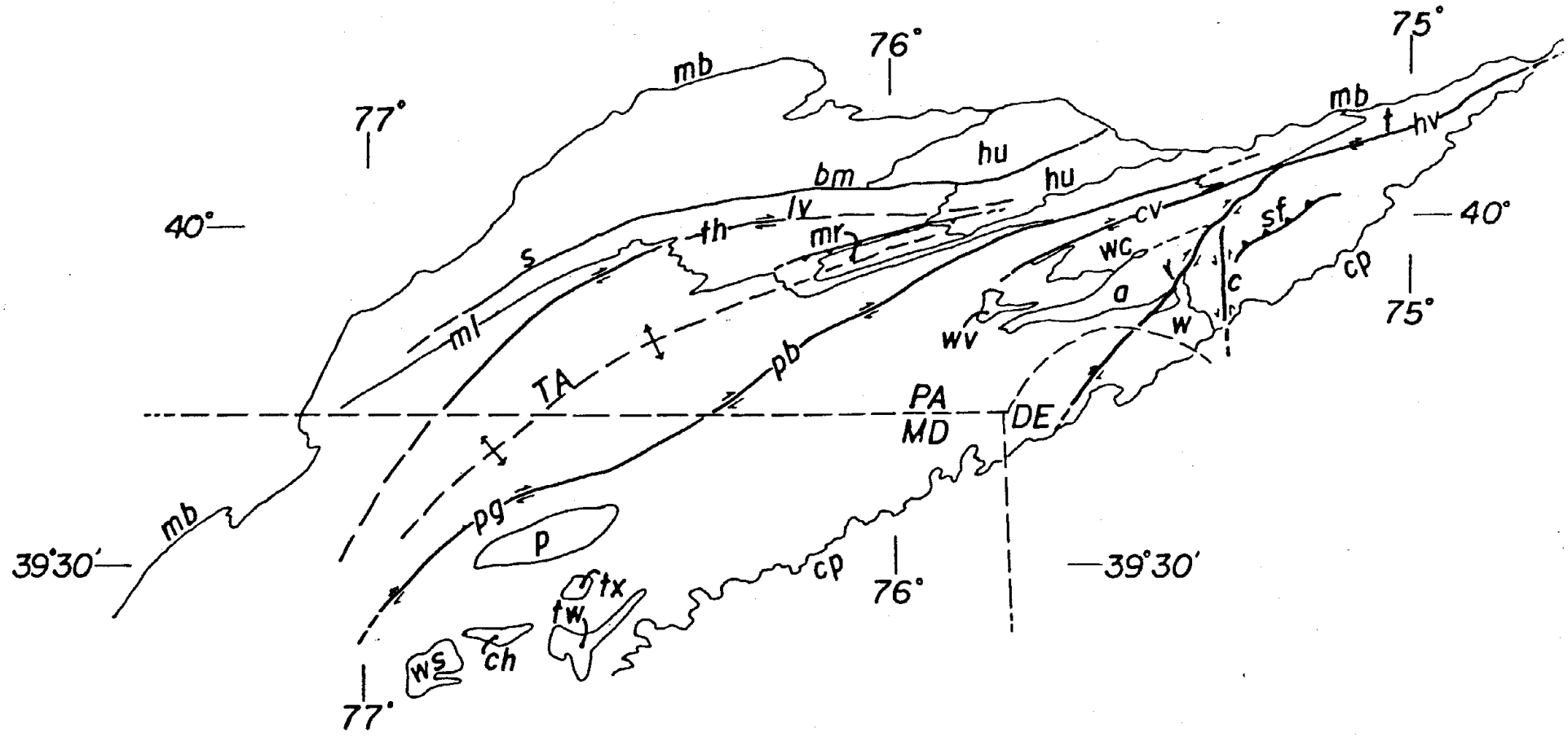
brief time (as suggested by identical metamorphic histories) suggests a model of transpressional deformation for the region. In other parts of the Pennsylvania Piedmont, evidence for late transpression recently has been documented. Gates (1989), in the State Line district, recognized a pattern of late conjugate strike-slip shear zones, folding of the Peters Creek Formation and reactivation of early structures into dextral strike-slip faults consistent with transpression. Bormack (1989) proposed a model of transpressional dome formation based on conjugate strike-slip shear zones and east-northeast directed shear thrusting for the Woodville Dome. Wise (1970) correlated the Woodville Dome with the same structural event that formed the Tucquan Antiform.

Alleghanian (?) Structures In The Pennsylvania Piedmont

The Lancaster Valley Tectonite Zone, Tucquan Antiform and Peach Bottom Structure have been shown to cross-cut rocks bearing Taconian metamorphism and structures. Lapham and Bassett (1964) dated individual D2/M2 micas and concluded an approximate age of 330 Ma for the regional D2 deformation. Faill and Valentino (1989) demonstrated that the Taconian metamorphic isograds of the western Piedmont were deformed by this late stage of D2 deformation and that the retrograde chlorite-biotite isograds on the margins of the Tucquan antiform are associated with the second regional metamorphism. Folds in the Lancaster Valley Tectonite Zone and Peach Bottom structure are characteristic of a large component of NNW-SSE subhorizontal compression, as is the domal Tucquan Antiform. A component of D2 dextral strike-slip shear was observed primarily in the southern half of the Lancaster Valley Tectonite Zone and the Peach Bottom Structure.

Along strike of the Peach Bottom Structure to the southwest, Freeman and others (1988) proposed dextral offset on a fault zone based on the three-dimensional geometry of ultramafic bodies determined by magnetic survey. The Pleasant Grove Shear Zone (Figure III-16) recently has been mapped by Krol and others (1990) and is characterized by dextral shear. The Pleasant Grove Zone is the along-strike equivalent to the Peach Bottom Zone to the southwest (Figure III-16). Similarly, along strike to the northeast, Baker (1987) proposed a dextral ductile shear zone in the Octoraro phyllonite parallel to the Martic Line. Baker (1987) correlated the ductile shearing with an episode of metamorphism (M2) that produced chlorite from biotite and garnet. Farther east, Myer and others (1985), Hill (1987), Song and Hill (1988), and Hill (1989) proposed dextral offset and chlorite-grade secondary metamorphism along the Martic Shear Zone.

The Cream Valley-Huntington Valley Shear Zone is the border fault between the northern margin of the West Chester Grenvillian Massif and Wissahickon lithologies in the Cream Valley and also is the border fault between the southern margin of the Trenton Grenvillian Massif and Wissahickon lithologies north of Philadelphia (Figure III-16). Armstrong (1941) mapped zones of mylonite along these faults and also recognized local retrograde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metamorphism associated with the mylonite development. Hill (1989) proposed that this retrograde metamorphism is the same as the retrogression along the Martic Zone and also proposed dextral offset on the Cream Valley-Huntington Valley Zone (Figure III-16) based on microstructural analysis.

R. Valentino (1989) compiled structural and metamorphic data from the Philadelphia Terrane and proposed that the dextral Rosemont Shear Zone (Valentino, 1988) and the sinistral Crum Creek Shear Zone (Figure III-16) are post-Taconian map-scale conjugate shear structures. A southern embayment of the Taconian metamorphic isograds mapped by Wyckoff (1952) corresponds directly with the boundaries of the mapped Crum Creek Shear Zone (Faill and Valentino, 1989). Offset of these isograds clearly demonstrates the post-Taconian nature of these conjugate shear structures (Figure III-16).

It appears that the extent of post-Taconian deformation characterized by retrograde metamorphism and dextral shear is not confined to western Piedmont structures and that all of the above-mentioned structures possibly comprise a regional scale transpressional shear system (Figure III-16). Alleghanian deformation in the southern Appalachian Piedmont is characterized by dextral strike-slip faults and transpressional domes (e.g., Bobyarchick, 1981; Gates, 1987). It is likely that the post-Taconian D2 deformation observed in the Pennsylvania Piedmont is the northern extension of the Alleghanian deformation observed in the southern Appalachians.

Figure III-16 (facing page). Map of Alleghanian(?) structures in the Piedmont province. Strike-slip faults: s=Stoner, bm=Brandywine manor, th=Turkey Hill, pg=Pleasant Grove, pb=Peach Bottom, cv=Cream Valley, hv=Huntington Valley, r=Rosemont, c=Crum Creek. Thrust faults: sf=Springfield, m=Mine Ridge thrust. Grenvillian massifs: ws=Woodstock, ch=Chattolance, tw=Towson, tx=Texas, p=Phoenix, mr=Mine Ridge, wv=Woodville, a=Avondale, wc=West Chester, hu=Honey Brook Upland, t=Trenton. mb=Mesozoic basin, cp=Coastal plain, lv=Lancaster Valley tectonite zone, ml=Martic line, w=Wilmington complex. PA=Pennsylvania, MD=Maryland, DE=Delaware.

IV. SAPROLITE AND LANDSCAPE EVOLUTION IN THE PIEDMONT

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INTRODUCTION

The Holtwood area (Figure IV-1) in southern Lancaster and York Counties is part of a surficial geology mapping project being done cooperatively by the Pennsylvania Geological Survey and the Maryland Geological Survey. The total project includes all of the York, PA-MD 1:100,000 scale topographic quadrangle. The Conestoga, Holtwood, Delta, and Bel Air Quadrangles (Figure IV-1) cover the initial area where mapping units and procedures for rapid mapping are being established. Thus, it is appropriate on this field conference to consider some of the surficial materials of the area and their relationships to the landscape and its history.

The Piedmont has long been considered an area of landscape stability and longevity. Recent work on saprolite within the Piedmont, however, as well as work elsewhere, has given rise to a moderate controversy regarding the longevity of landforms and the age of saprolite. This chapter reviews the characteristics and origin of saprolite, the relation between saprolite and Piedmont landscape, and some aspects of Piedmont landscape evolution in Lancaster and York Counties, Pennsylvania.

SAPROLITE

Description

The term saprolite originally was applied to rocks in North Carolina by Becker (1895, p.302), who wrote, "The surface rocks are decomposed, and almost everywhere to a considerable depth. Perhaps 50 feet would be a fair estimate of the thickness of the rotten layer, for which I have suggested the name saprolite...." Becker did not elaborate on the characteristics of saprolite nor did he discuss its origin beyond use of the words "decomposed" and "rotten."

Modern use of the term saprolite generally implies the following characteristics (Pavich, 1985, p.308): "...it is isovolumetric with the underlying bedrock, as indicated by the retention of texture and fabric of the parent material, and it exhibits gradational chemical and mineralogical changes of composition going from the parent to the geomorphic surface." In addition, Carroll (1970, p.19-20) says that there is little or no "movement of alteration products. Leaching has changed feldspars to clay minerals and oxidation of ferrous iron to ferric iron has given the saprolite a brownish color...." Saprolites are typically soft and are easily dug with a shovel or cut with a knife.

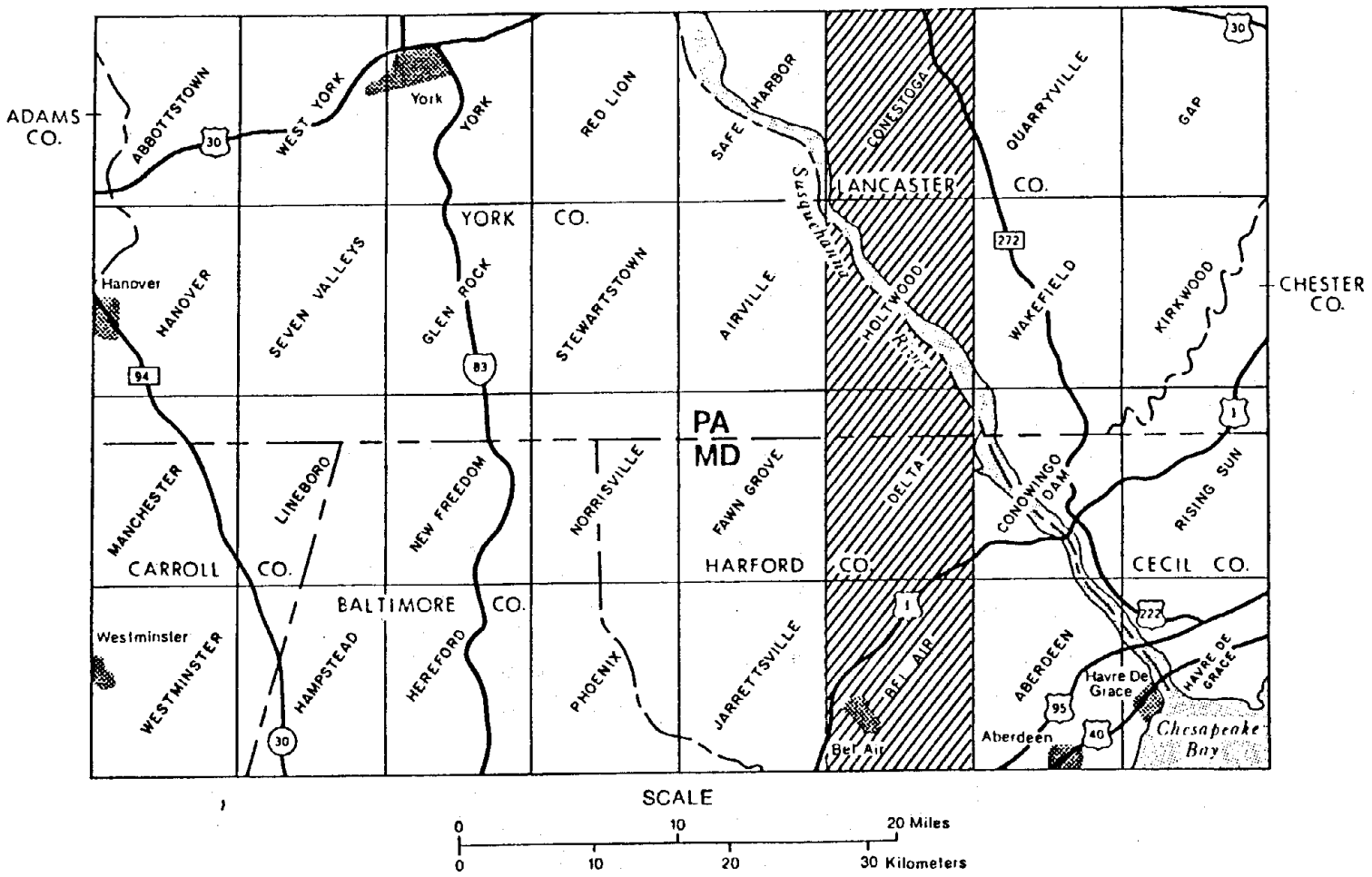


Figure IV-1. Index map of the York, PA-MD 1:100,000 scale topographic gradrangle showing principal cultural features, 1:24,000 scale topographic map names, and the area of initial work (diagonally lined).

Immediately above fresh bedrock is a layer of weathered bedrock (Figure IV-2) which is variable in thickness and must be broken with a hammer. This weathered rock "is discolored brown or yellow with hydrated iron oxides, especially along partings. Clayey alteration of minerals in the rock can be seen with a microscope, but the minerals are still firm.... In dense rocks this weathered layer is thin; in porous types it may be many feet thick...." (Hunt, 1972, p.150-155). Pavich and others (1989, p.25) note that solution movement and weathering in this zone "is restricted to relatively large joints and to fractures of high permeability."

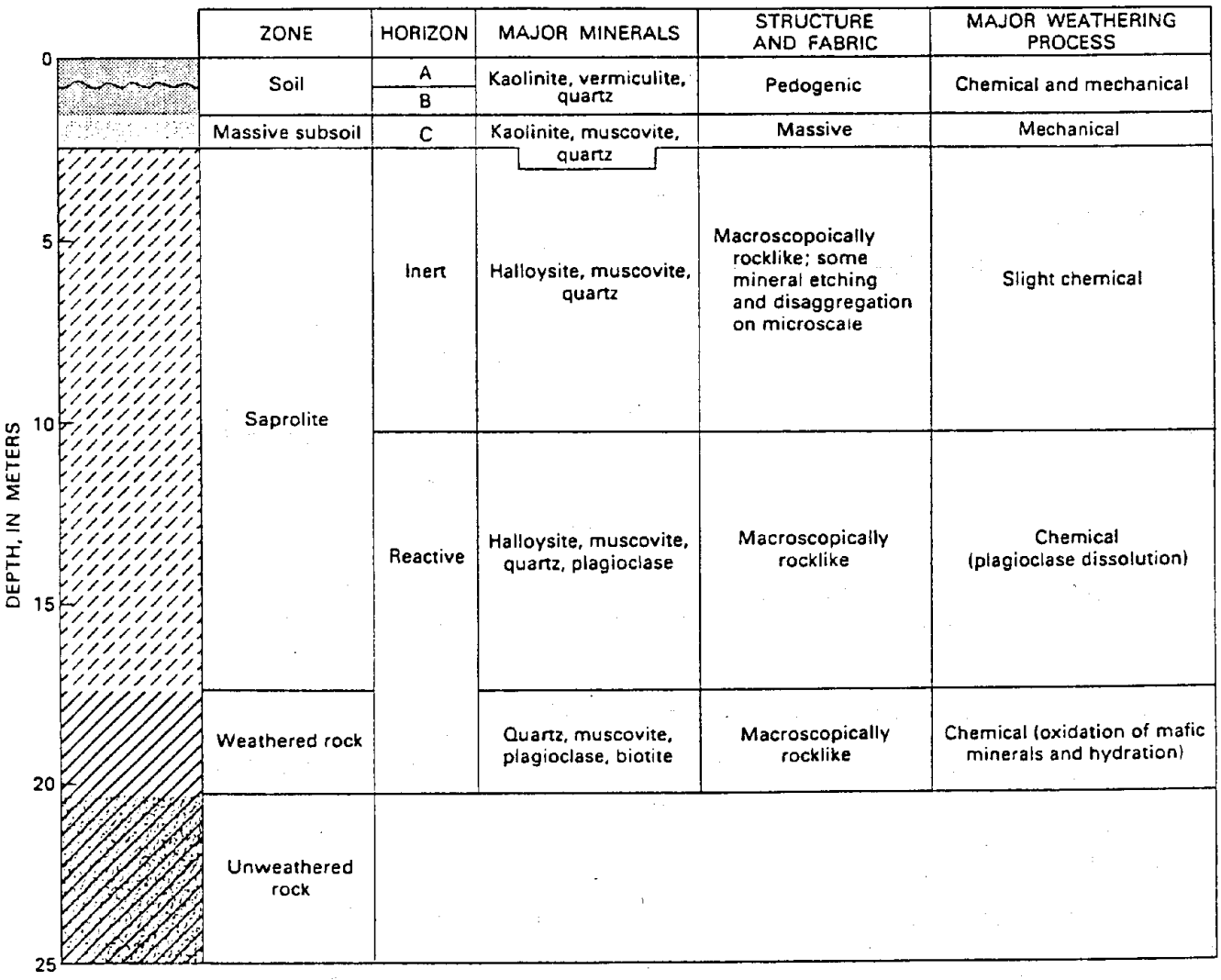
Above the weathered bedrock is the saprolite or structured saprolite (Figure IV-2). This zone preserves the structure of the parent rock because there has been no mechanical disruption, but the mass has been chemically altered so that its density is only half that of the original rock (Hunt, 1972). "The properties of the saprolite are not uniform as a function of depth from the geomorphic surface. The zonation of these properties is related to the differences in primary mineral stabilities and to the difference in duration of weathering between the bottom and top of the saprolite...." (Pavich and others, 1989, p.25). Although changes in density are transitional, changes in mineralogy and chemical composition are distinct with mainly inert minerals occurring in the upper part of the saprolite (Pavich and others, 1989).

Above the structured saprolite there generally is a layer of massive saprolite, a zone similar in appearance to structured saprolite except that it lacks the original rock structure. The boundary with the underlying structured saprolte is gradational. Pavich and others (1989) refer to this zone as massive subsoil (Figure IV-2) and indicate three criteria for its recognition: 1) disruption of original grain-to-grain contacts between resistant residual framework minerals, 2) an upward increase in bulk density because of volume decrease, and 3) a decrease in mechanical strength. This massive zone, generally within 6 ft (2 m) of the surface, results from mechanical disruption of the structured saprolite by burrowing organisms, roots, seasonal wetting and drying, and frost action.

The uppermost zone is the soil proper, which includes the pedogenic A and B horizons. The soil is a zone of extremely active physical and chemical processes where mass and volume are constantly reorganized and quartz and muscovite, which are relatively stable in the upper saprolite, undergo significant chemical alteration. The mechanical processes affecting the soil zone are the same as those acting on the massive subsoil, but the processes act with greater intensity.

Origin

Saprolite is produced by the complex interactions of chemical weathering caused by ground water. The zone of maximum weathering occurs at the rock-saprolite interface (Figure IV-3).



ZONE	HORIZON	MAJOR MINERALS	STRUCTURE AND FABRIC	MAJOR WEATHERING PROCESS
Soil	A / B	Kaolinite, vermiculite, quartz	Pedogenic	Chemical and mechanical
Massive subsoil	C	Kaolinite, muscovite, quartz	Massive	Mechanical
Saprolite	Inert	Halloysite, muscovite, quartz	Macroscopoically rocklike; some mineral etching and disaggregation on microscale	Slight chemical
Saprolite	Reactive	Halloysite, muscovite, quartz, plagioclase	Macroscopically rocklike	Chemical (plagioclase dissolution)
Weathered rock		Quartz, muscovite, plagioclase, biotite	Macroscopically rocklike	Chemical (oxidation of mafic minerals and hydration)
Unweathered rock				

Figure IV-2. Generalized weathering profile of thick regolith developed on upland quartzofeldspathic rocks (Pavich and others, 1989, Figure 17, p. 34).

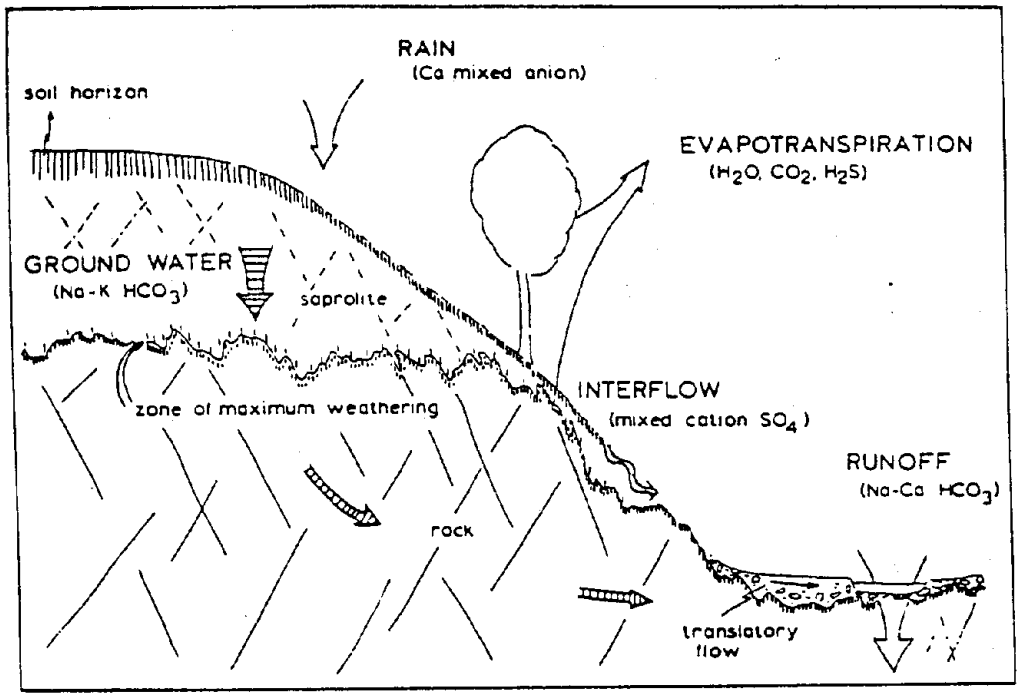


Figure IV-3. General weathering model for saprolite development (Cleaves and others, 1970, Figure 7, p. 3028).

"The chemical weathering is a constant-volume process, whereby 50 to 60 percent of the original rock mass is removed as dissolved solids in percolating groundwater. In the transition from rock to saprolite, original rock minerals are replaced by secondary minerals of lesser density, bulk density decreases and porosity increases...." (Cleaves, 1974, p.1).

"As the water reacts with the minerals some of the reactants are removed in solution and eventually discharged into the surface water (alkali cations, alkaline earth cations, bicarbonate, and dissolved silica). Other reactants are reconstituted as oxides and clay minerals. Mineral weathering sequences...are: plagioclase alters to kaolinite and gibbsite; biotite to vermiculite and kaolinite...and muscovite to illite and kaolinite...." (Cleaves, 1983, p.48-49). These secondary minerals occupy the space of the original minerals, but have lower densities. The rock framework is maintained by the quartz which is not dissolved sufficiently to change the rock volume. Color is added to the saprolite by ferric oxides produced during the weathering process.

There are three primary controls on the formation of saprolite: 1) rock type, 2) rock structure, and 3) climate.

Rock Type

Extensive chemical weathering, even in the same watershed, does not always produce saprolite. Cleaves and others (1974)

and Cleaves (1983) have pointed out that chemical weathering of rock types such as serpentinite is a non-isovolumetric process because there is no quartz to form a framework. Thus, when antigorite (the only significant weatherable mineral in serpentinite) is weathered, there is little accumulation of secondary minerals and the rock is slowly denuded by solution. The same thing happens in relatively pure carbonate rocks which have no framework minerals and weather to a structureless mass. Locally in the Holtwood area the Conestoga Formation has a sandy facies in which quartz provides a framework and so the marble weathers to form thick saprolite.

Pavich and others (1989) note that saprolite in Fairfax County, Virginia, is thickest on quartzofeldspathic metapelite, metagraywacke, and foliated granite; thin on diabase; and virtually nonexistent on serpentinite. Saprolites developed on sandstones in Pennsylvania (Sevon, 1975; Berg, 1975; Berg and others, 1981) appear to be moderately thick.

The Wissahickon Formation in the Holtwood area is variably classified as mainly mica schist, chlorotoid mica schist, and garnet mica schist. Table IV-1 shows the estimated compositional variation of Wissahickon rocks in the Holtwood area.

Mineral	Mean (%)	Range (%)	Absent*
Muscovite	44	7 - 70	0
Quartz	22	1 - 50	0
Chlorite	18	7 - 30	0
Chloritoid	4	0 - 60	24
Plagioclase	3	0 - 37	13
Ilmenite	3	0 - 10	1
Magnetite	2	0 - 5	10
Biotite	1	0 - 10	11
Garnet	1	0 - 20	22
Other	1	0 - 20	

* - Number of samples

Table IV-1. Compositional data derived from modal estimates for 36 thin sections of rocks of the Wissahickon Formation collected from between Pequea Creek to the north and Holtwood Dam to the south in Lancaster County, Pennsylvania. Data from unpublished work of David Valentino, Pennsylvania Geological Survey.

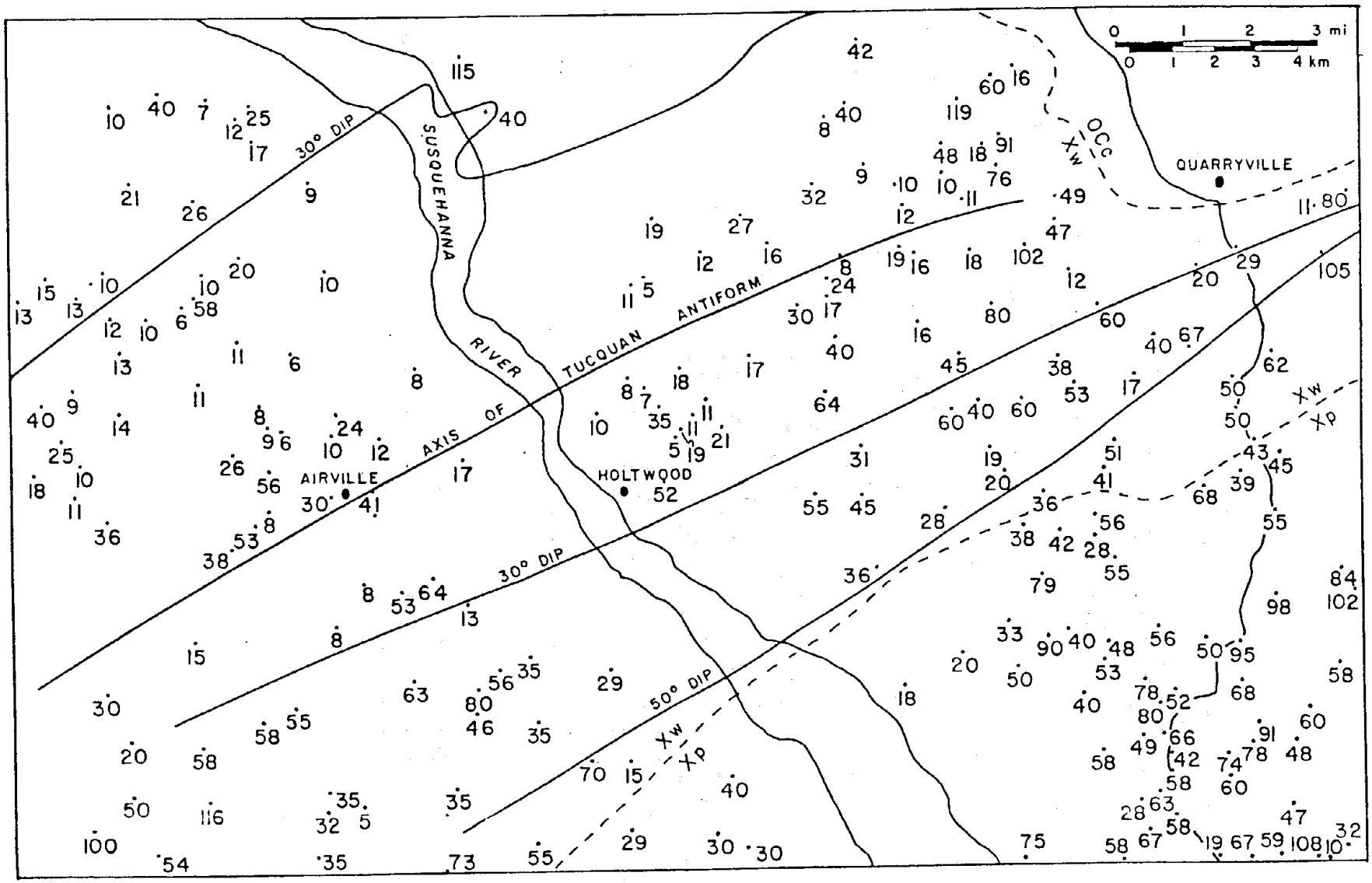
No analyses of mineralogical changes which occur during the transformation of this fresh rock to saprolite have been made in the Holtwood area, but the alterations should be similar to those described for areas to the south in Maryland and Virginia (Cleaves, 1974; Pavich, 1986; Pavich and others, 1989).

Rock Structure

Both Cleaves (1973) and Pavich (1986) suggest that rock structure is important in saprolite formation because of its relationship to water movement. The steeply dipping foliation of most Piedmont rocks enhances anisotropic movement of water, primarily downward. In contrast, Schoenberger and Aziz (1990) demonstrate that there is no significant difference in saturated hydraulic conductivity of saprolite regardless of orientation relative to primary foliation. However, their work treated samples taken in the soil zone and the saprolite just below the soil at the top of sequences of thick saprolite and thus may not reflect the importance of orientation during the early stages of weathering of rock to form saprolite. Lateral movement occurs mainly along brittle fractures and at the interface between weathered rock and saprolite where the major permeability change occurs. If rock structure is a control on saprolite development, then saprolite should be thickest in areas of steeply dipping primary foliation and thinnest in areas of horizontal primary foliation.

This relationship seems to be demonstrated in the Holtwood area where the primary foliation in the schistose rock flattens and reverses dip direction across the Tucquan antiform. Information about the position of the axis of the Tucquan antiform and the variation in dip of primary foliation was combined with depth-to-bedrock data to produce a generalized saprolite thickness map for the Holtwood area (Figure IV-4). The depth data were retrieved selectively according to the following generalized criteria: all wells on uplands and the upper parts of side slopes above any major change in gradient were used and all wells in valley bottoms and on the lower parts of side slopes below any major change in gradient were not used. The raw data on the east side of the Susquehanna River generally support the hypothesis. The data on the west side of the Susquehanna River are more scattered for reasons not yet understood. When the raw data are smoothed (Figure IV-5), the trend is well displayed and the hypothesis appears to be well supported, although the exis of the antiform is at the soth edge of the area ofminimum depth-to-bedrock. The area north of the axis has variation, but generally low dips of primary foliation. An objection to the validity of the demonstration is that the saprolite is thinnest east of the Susquehanna River in the area

Figure IV-4 (facing page). Map showing depth to bedrock at selected locations in the Safe Harbor, Conestoga, Quarryville, Airville, Holtwood and Wakefield quadrangles (Figure IV-1). The depth to bedrock (data from water well information on file at the Pennsylvania Geological Survey) presumably reflects the thickness of saprolite. The map also shows the axis of the Tucquan antiform, generalized zones of dip of primary schistosity, and generalized bedrock geology (Freedman and others, 1964; Wise, 1970; Berg and others, 1980; D. Valentino, pers. comm.)


0 1 2 3 mi
0 1 2 3 4 km
QUARRYVILLE
AIRVILLE
HOLTWOOD
SUSQUEHANNA RIVER
TUCQUAN ANTIFORM
AXIS OF
30° DIP
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50° DIP
OCC
Xw
Xp

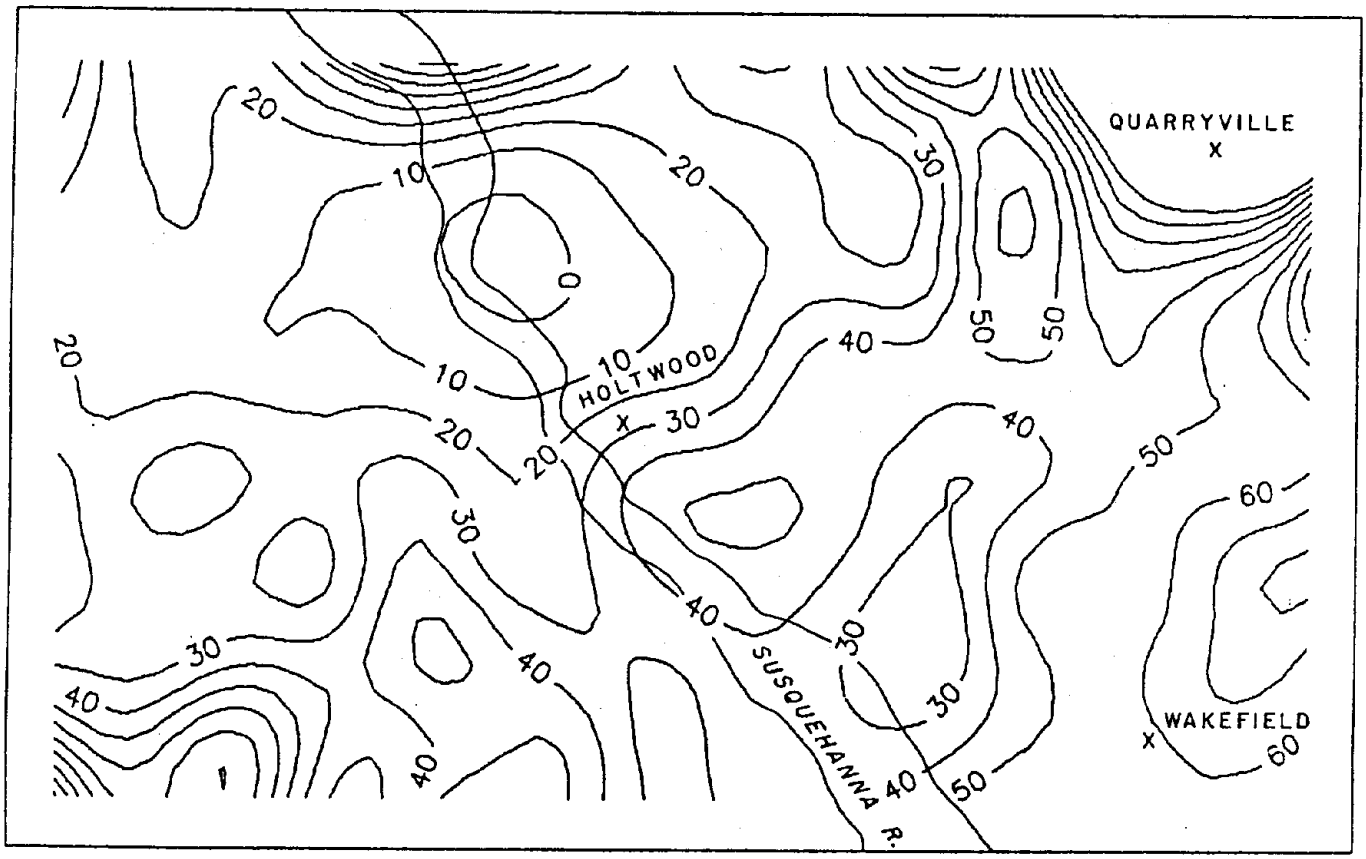


Figure IV-5. Contour map of the depth to bedrock data presented in Figure III-4. The data were computer-processed using SURFER, Version 4 (Golden Software, Inc.) for maximum smoothing.

of highest elevation where local relief is greatest, slopes are steepest, and more erosion of saprolite may have occurred.

Climate

Climate affects principally the rate at which saprolite forms. The amount of precipition is the primary control factor, but temperature is also very important. "At constant volume, slow movement of water to channel favors more concentrated weathering solutions and more rapid denudation than does rapid flow at shallow depths...." (Dethier, 1986, p.505). Furthermore, "chemical denudation increases with runoff because larger volumes of...water are available to displace mineralized pore waters and to flush readily soluble constituents from particle surfaces...." (Detheir, 1986, p.521). Thus, in general, the more humid the climate, the more rapid the rate of saprolitization.

Temperature is important because of its effect on chemical reaction rates. Jenny (1941, p.143) points out that, "For every 10 °C rise in temperature the velocity of a chemical reaction increases by a factor of two to three. The rule [Van't Hoff's] holds for a large number of chemical reactions, particularly slow ones and applies equally well to numerous biological phenomena." This is true for everything except carbonate rocks, because cold water is able to hold more CO_2 than warm water and, therefore, $CaCO_3$ should dissolve more readily in cooler climates than in warmer climates (Birkeland, 1974).

Table IV-2 presents some basic climate data which indicate that, based on temperature alone, the rate of saprolitization in equivalent rocks should be twice as rapid in South Carolina as in Pennsylvania. The additional precipitation farther south also should increase the rate.

Place	Mean Annual Temperature (in °C)	Mean Annual Rainfall (in mm)	Groundwater Temperature (in °C)	Number of Wells
York, PA	11.7	1032	12.6	171
Washington, DC	13.9	1036	13.3	46
Richmond, VA	14.3	1119	14.5	58
Greensboro, NC	14.6	1072		
Atlanta, GA	16.4	1197	18.8	4
Charleston, SC	18.3	1250	22.4	7
Jacksonville, FL	20.8	1355	23.6	273

Table IV-2. Air temperature, rainfall, and groundwater temperature data for selected places in the Atlantic coastal states. Climate data from NOAA and groundwater temperature data from the U. S. Geological Survey.

Cleaves (1989) recently has drawn attention to the strong relationship between rate of saprolitization and soil CO_2 concentrations. He points out that the soil CO_2 reacts with water to form carbonic acid, the primary weathering agent. Because carbonic acid is more soluble in cooler than warmer waters, the presence of more carbonic acid in cooler temperature regimes should promote more weathering.

However, an additional factor is that CO_2 concentrations "depend upon the rate of plant decomposition, microbial and root respiration, and the diffusion rate of CO_2 in the atmosphere...." (Cleaves, 1989, p.166). Thus, a prime factor in generation of soil CO_2 is litterfall and its subsequent decomposition. Spurr and Barnes (1973) point out that litterfall quantity in equatorial forests is twice that in warm temperate forests and three times that in cool temperate forests. In addition, microflora and microfauna in equatorial forests decompose litter at a rate 6 to 10 times that in temperate zones. Waring and Schlesinger (1985) report that release of C as CO_2 due to soil microorganism activity is parabolic relative to moisture, with an optimum moisture of about 40-45 percent. The release changes dramatically with temperature, the amount nearly doubling for every 10 °C increase. They also show that the rate of decomposition of fresh litter in South Carolina is about twice that in Pennsylvania.

Another factor, possibly of considerable importance, is the soil (or saprolite) thickness. According to Stallard (1985, p. 296-297), "For a given set of conditions (lithology, climate, slope, etc.), there is presumably an optimum soil thickness which maximizes the rate of bedrock weathering." If the soil is too thin, some or much of the water supplied by precipitation is lost to runoff. Water infiltrates and circulates slowly through thicker soils, especially where the land is forested. If soils are too thick, water residence times at the base are long and weathering is slowed.

Thus, the rate of saprolitiziation varies with the complex interplay of precipitation, temperature, soil CO_2 concentrations, and soil thickness, all of which are climate-dependent. Theoretical calculations of the modern rate of saprolitization by Cleaves (1989) give rates of 25-48 m/my (82-157 ft/my). Other rates of chemical denudation calculated through mass balance studies in different small drainage basins are presented in Table IV-3. The rate of saprolitization must have varied considerably in the Holtwood area during the past 2 million years because of the differences between glacial and interglacial climates.

Rate (m/my)	Area	Rock	Reference
2	Maryland	Schist	Cleaves and others, 1970
2.2	Maryland	Serpentinite	Cleaves and others, 1974
1.2	Maryland	Schist	Cleaves and others, 1974
2	Virginia	Sandstone	Alifi and Bricker, 1983
10	Virginia	Shale	Afifi and Bricker, 1983
37	N. Carolina	Mixed	Afifi and Bricker, 1983
3.3	Virginia	Granite	Pavich, 1986

Table IV-3. Rates of chemical denudation calculated through mass balance studies in small drainage basins in the North American Atlantic coastal states.

LANDSCAPE EVOLUTION IN THE PIEDMONT OF SOUTHERN LANCASTER AND YORK COUNTIES, PENNSYLVANIA

Introduction

The Holtwood area (Figure IV-1) comprises a rolling topography which tends towards the extremes of 1) broad, relatively flat uplands in drainage divide areas far removed from the Susquehanna River and 2) steep-sided, narrow valleys in deeply dissected areas near the Susquehanna River. The most striking feature of the area is the Susquehanna River which flows in a narrow, deep, steep-sided gorge. Local relief generally decreases with increasing distance from the Susquehanna River and is least in the southeastern part of the area. Except for the more dissected parts of the landscape, the area is underlain by saprolite of variable thickness (Figures IV-4 and IV-5) which comprises the unconsolidated surface material of greatest volumetric proportion. When the landscape is viewed from an upland vantage, there is an apparent visual accordance of uplands. These uplands commonly have been referred to as remnants of the Harrisburg Peneplain. Except for the small area of carbonate rocks around Quarryville in the northeast, the Holtwood area is underlain by schists of the Wissahickon Formation or the Peters Creek Formation (Figure IV-4).

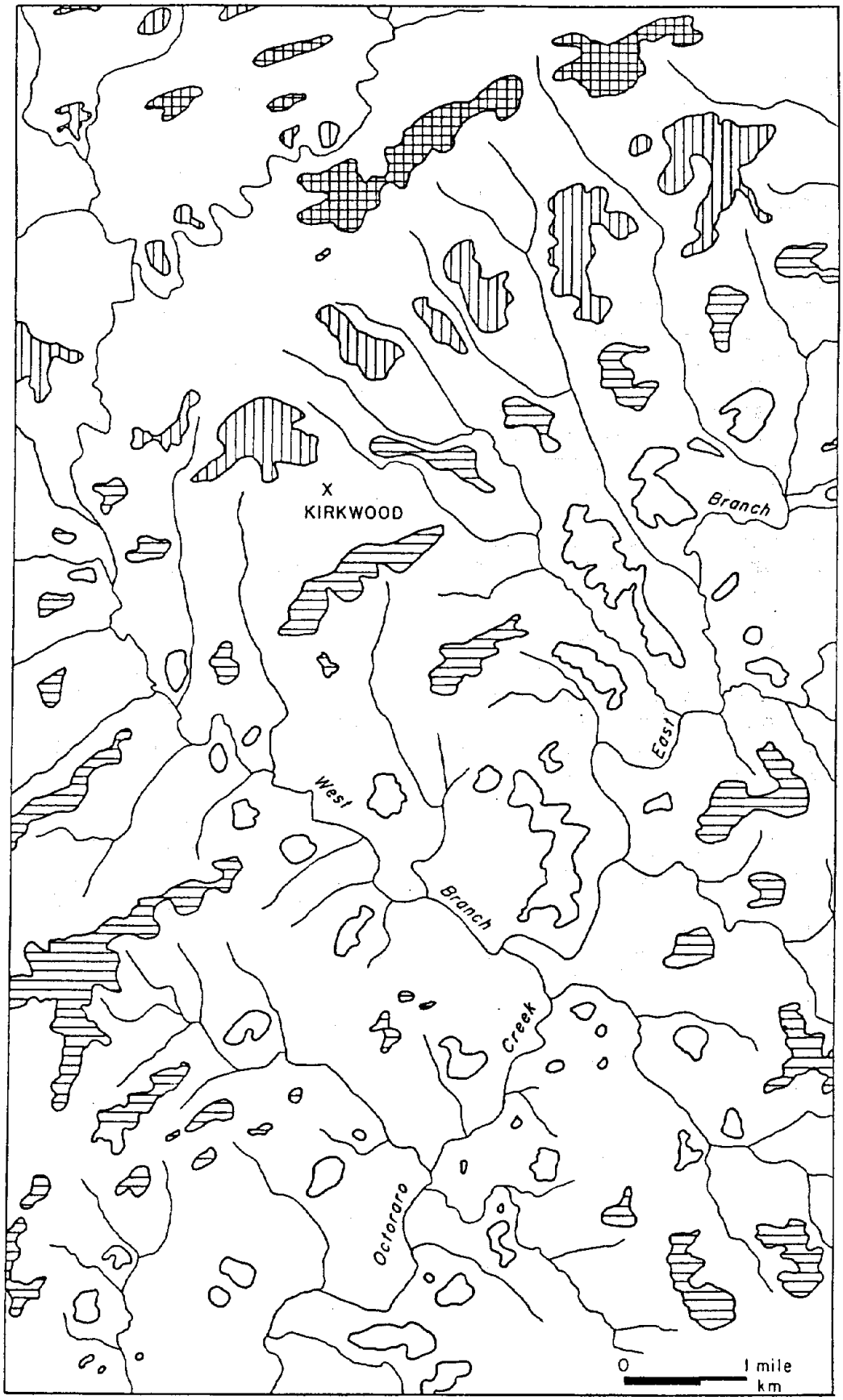
All of the larger stream valleys, and even many of the very small intermittent stream valleys, have some alluvium on the valley bottom. In general this alluvium is not thick and much of the upper part probably was deposited after the land was cleared for cultivation. The alluvium occurs in alluvial plains which range in width from very narrow to several hundred feet. All of the tributaries to the Susquehanna River in the Holtwood area show a similar pattern: they are entrenched with narrow valley bottoms and steep valley walls near the Susquehanna and change character dramatically a short distance upstream where the valleys bottoms are broad and flat and slopes on the valley walls are moderate to gentle.

Almost all of the slopes have a blanket of colluvium of variable thickness. The steep-sloped, deep valleys near the Susquehanna River often have bedrock exposed along the upper valley walls, but may have boulder colluvium covering part of the valley wall, particularly the base. Low to moderate gradient slopes coming from uplands far removed from the Susquehanna River have thin to thick deposits of colluvium which comprise mainly transported saprolite and weathered bedrock. In the headwater areas of many small drainages the colluvium grades imperceptibly downslope into alluvium. In stream valleys where the floodplain is well-defined, there is generally a distinct change in slope where the floodplain alluvium merges with colluvium at the margin of the valley bottom. The upslope terminus of colluvium commonly is obscure. These deposits were formed mainly during the Pleistocene and Holocene Epochs. Although these deposits constitute a significant volume of surficial material in the Holtwood area, they are not discussed further here.

Most of the history of landscape development in the Piedmont of Pennsylvania is lost. Slingerland and Furlong (1989) showed that during the Alleghanian orogeny the Piedmont was part of an orogenic highland which reached heights of 2.2-2.8 mi (3.5-4.5 km) and a width of 155-185 mi (250-300) km. During and subsequent to the Alleghanian erosion of this highland sediment was fed to a vast alluvial plain to the west and drainage from or across the Piedmont was to the west and northwest. Since the end of the Alleghanian Orogeny at least 6 mi (10 km) of material has been eroded from the Piedmont in the Holtwood area (Jamieson and Beaumont, 1988). We may presume that much of this erosion occurred prior to Late Triassic and Jurassic rifting, but the amount is not known.

Drainage was reversed during Late Triassic-Jurassic rifting and early development of the Atlantic Ocean, and the Susquehanna and Schuylkill Rivers began to flow across the Piedmont to the Atlantic (Sevon, 1989a). Conglomerates in the Triassic rocks of the Gettysburg-Newark Basin occur in positions comparable to the present courses of those rivers, suggesting that the rivers have not changed their positions appreciably since the Triassic. The events that occurred between then and more recent times are open to speculation. Thompson (1988) recently discussed some of the features of the Susquehanna River in the Holtwood area and their possible origins.

Figure IV-6 (facing page). Map showing topographic steps in the Kirkwood and part of the Gap quadrangles (Firgure IV-1). Each step has relief o fless than 40 feet and corresponds to an area of relatively flat appearance both in the field and on the topographic map. Lower contour for each of the steps is as follows: cross-hatched - 700 feet; vertical lines - 600 feet; horizontal lines - 480 feet; open - 400 feet.


X
KIRKWOOD
Branch
East
West
Branch
Creek
Octororo
0
1 mile
km

Earlier Work

The only published account of the geomorphology of the Holtwood area is that of Knopf and Jonas (1929). They departed from the vogue of their time by suggesting that the topography of the area could not be fit into a scheme of three erosion cycles (peneplains) as was generally done by other workers elsewhere in Pennsylvania. They recognized numerous flat-topped uplands developed on beveled upturned rocks as remnants of old surfaces of low relief. They believed, however, that a succession of these flat surfaces stair-stepped from Mine Ridge to the ocean and that these steps were caused by a "reoccurence of numerous slight uplifts that caused repeated interruptions to the continuity of baseleveling...." (Knopf and Jonas, 1929, p.117). Such uplifts would be caused, they presumed, by isostatic adjustment to removal of material by erosion. They also suggested (p.118) that "The evidence...suggests that no erosion cycle recorded in the region between Blue Mountain and the Coastal Plain is older than late Tertiary." Current opinion that all parts of the landscape are being lowered by erosion conflicts with their belief "that on existent divides remnants of surfaces cut during previous cycles are more or less immune to the destructive agencies of the present cycles." (p. 97).

The essence of the work of Knopf and Jonas lay in the recognition of stepped or terrace-like surfaces descending from an upland such as Mine Ridge. They particularly noted the drainage divide between the Susquehanna and Schuylkill Rivers as a prime example which shows seven such steps between Mine Ridge and the Atlantic Ocean.

Field examination of the drainage divide itself is inconclusive in the sense that there appear to be innumerable flat-topped surfaces of limited lateral extent. However, step-like, flat-topped uplands which may be comparable to what Knopf and Jonas saw are very evident just west of the Susquehanna-Schuylkill drainage divide in the Gap and Kirkwood Quadrangles (Figure 1). Here in the upper reaches of the Octoraro Creek drainage basin are four well-defined topographic steps (Figure IV-6) starting at Mine Ridge at an elevation of 700 feet (214 m) with each lower step about 100 feet (30 m) less in elevation. Some latitude was taken in defining the steps in that up to 40 feet (12 m) of local relief was allowed for a defined upland. For most uplands the local relief is about 20 feet (6 m) and appearance in the field is of a gently rounded nearly flat upland. A similar but not as well-defined pattern of steps occurs to the west in the Wakefield Quadrangle (Figure IV-1). No search for such steps has been made elsewhere in the area.

Campbell (1933) considered the Piedmont area of southeastern Pennsylvania to be representative of the Harrisburg Peneplain which he had defined earlier (1903). In his 1933 paper (p.571-573) he did a logical estimation of the age of the peneplain surface by calculating the time necessary to erode the Susquehanna River Gorge from Turkey Hill south to its mouth. He concluded

that the peneplain had reached its full development by the Miocene and that dissection of the surface had occurred since that time. He did not relate the surface to saprolite.

Campbell (1929; 1933) also hypothesized warping of the Harrisburg Peneplain to create the Westminster Anticline (Figure IV-7). He based this hypothesis on the presence of what he interpreted to be uplifted terrace gravels along the Susquehanna and Potomac Rivers. He firmly believed that upland surfaces in the area of his proposed anticline could represent only the remnants of a former peneplain. Stose (1929) correlated the same terrace gravels but did not show any warping and he suggested (1930) that Campbell had not correlated the gravels correctly. No correlation studies of gravels along the lower Susquehanna River have been undertaken since then. Stose and Jonas (1939b) did not mention the Westminster Anticline in their report on York County. They did, however, define the Glen Rock Anticline across part of York County and the axis of the Glen Rock anticline is in part coincident with that of the Westminster Anticline (Figure IV-8).

Is the Westminster Anticline real? When viewed with the concept of a peneplain surface as a reality, topography in the area outlined by Campbell (1933) for the anticline (Figure IV-7) presents a strong visual impression that the uplands define a crest area with other uplands gradually becoming lower to either side of the crest. These uplands are all underlain by saprolite. West of the Susquehanna River a drainage divide between

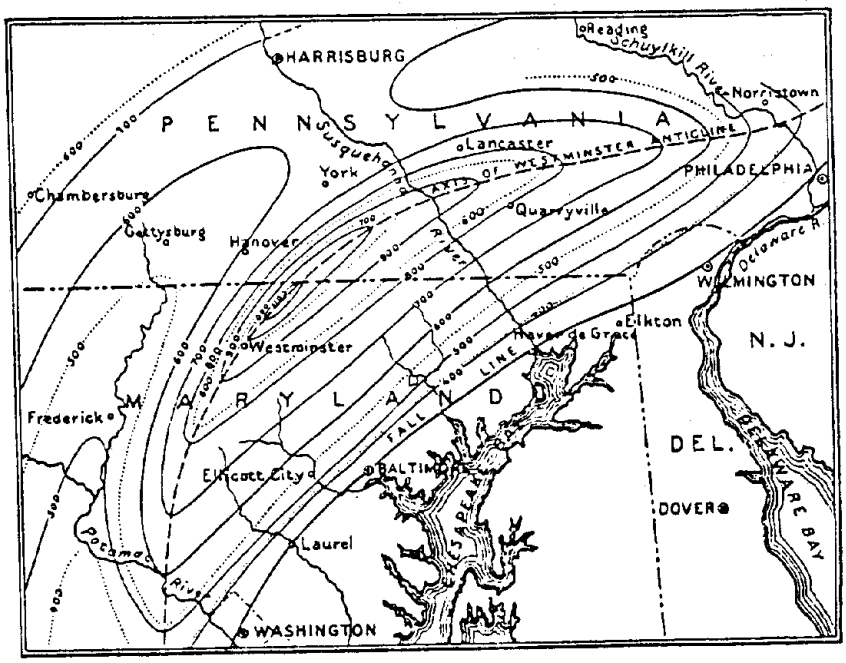


Figure IV-7. Map of the Westminster anticline. Dotted lines show deformation of the Bryn Mawr berm, and solid line show the deformation of the Chambersburg peneplain. From Campbell, 1933, Figure 3, p. 568.

north-flowing and south-flowing streams wanders back and forth across the axis of the anticline. The highest uplands, however, occur along the drainage divide and the most extensive areas of uplands occur either north or south of the axis, sometimes a far as 2.5-3 mi (4-5 km) (Figure IV-8). Therefore, the axis of the Westminster Anticline and the contoured deformation of the Harrisburg Peneplain as indicated by Campbell (Figure IV-7) bear little relationship to the existing topography. In addition, if erosion has proceeded in a stair-step fashion as suggested by Knopf and Jonas, then no surface other than that along the crest of the drainage divide could be a remnant of the oldest erosion surface. All lower surfaces would be younger and could not be correlated with the highest surface to define the form of a warped surface.

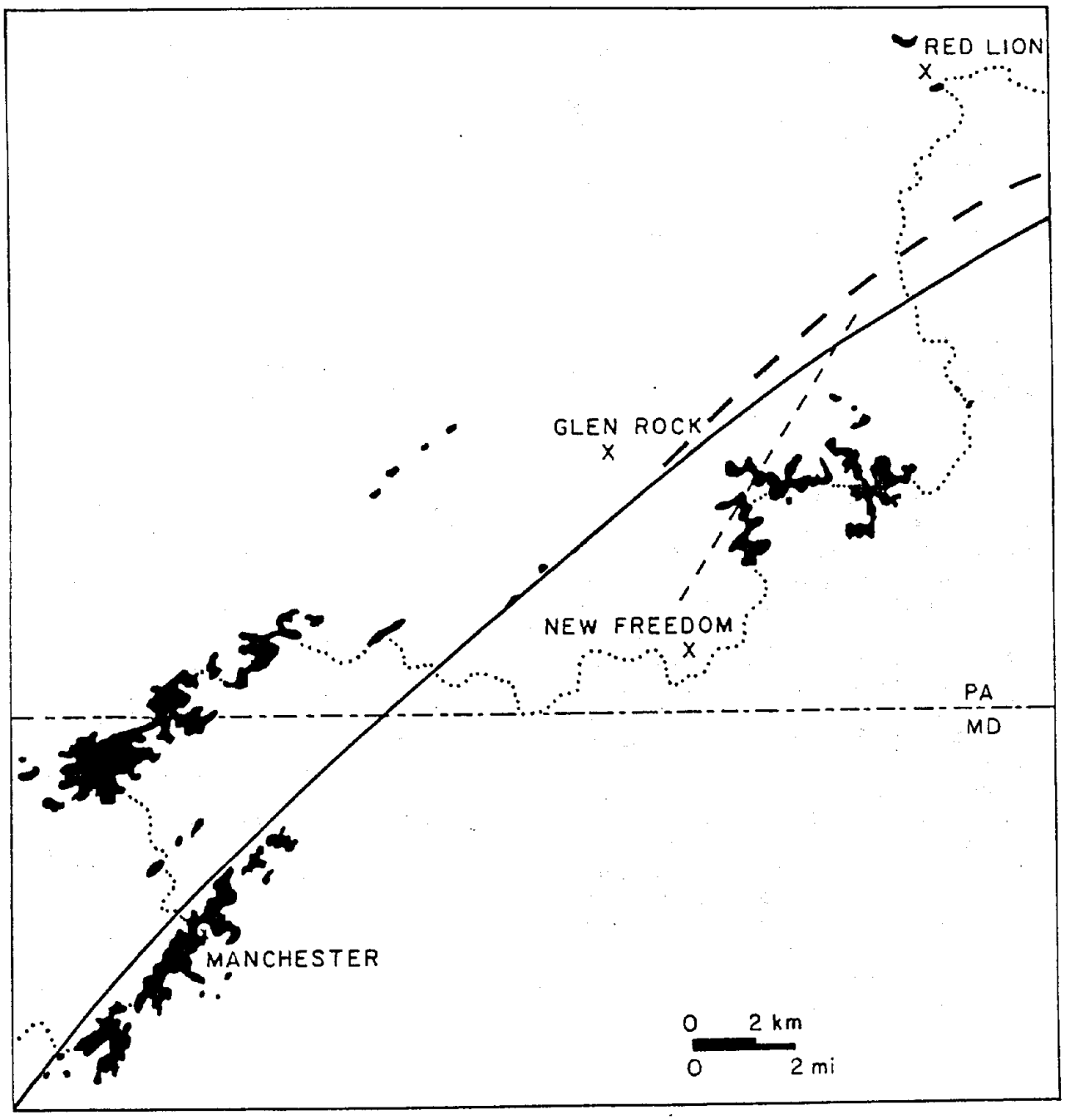
Finally, there is no observable deformation of the Wissahickon Schist relative to the anticline. Dip of primary schistosity appears to maintain a fairly uniform northward dip across the axis of the anticline. The only structure close to the axis of the Westminster Anticline is a structural bench (Figure IV-9) whose axis is diagonal to that of the anticline. Thus, there seems to be no real evidence for the Westminster Anticline. The same observations suggest that there is no Glen Rock anticline.

Recent Work

The concept that the Piedmont landscape, particularly as exemplified by relatively flat uplands underlain by thick saprolites, is an old, dissected peneplain was recently succinctly restated by Cleaves and Costa (1979) and Costa and Cleaves (1984). They argued that the main time of planation and saprolite development was during Late Cretaceous and early Tertiary time, and that erosional incision of the landscape has occurred mainly since the Miocene.

Pavich (1985, 1986, 1989a, 1989b), on the other hand, has argued that modern rates of saprolite production are sufficiently rapid to allow development of thicker saprolite than presently exists on Piedmont rocks. He also shows through ^{10}Be analysis that soil residence time is relatively short and that there must be erosional loss from the upper surface, even in the flat upland areas where erosion is generally considered to be minimal to nonexistent. Pavich argues that a balance between saprolite production and upland erosion exists in the Piedmont and that

Figure IV-8 (facing page). Map showing axis of the Westminster anticline (solid line, Campbell, 1939), the Glen Rock anticline (heavy dashed line, Stose and Stose, 1939b), drainage divide between north-flowing and south-flowing streams (dotted line), and the areas of highest upland topography (black areas). Base map was the York, PA-MD, 1:100,000 scale quadrangle.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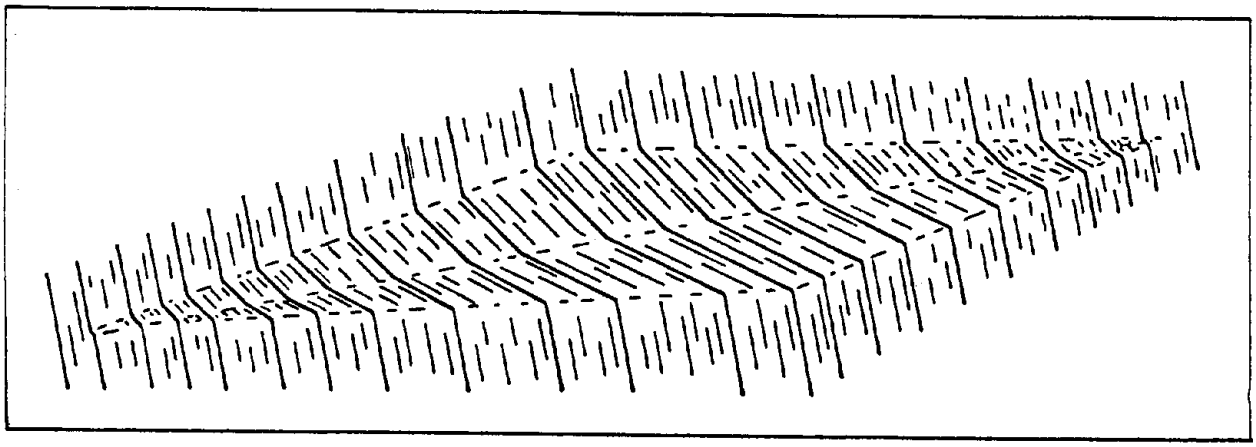

Figure IV-9. Hypothetical drawing of a structural bench displayed by primary schistosity in the Wissahickon Formation in southern York County. View looking at the bench from the north. See Figure IV-8 for trend and location.

this approximates a state of dynamic equilibrium. He believes that there is neither evidence nor necessity for a peneplain surface capped by thick saprolite. Additional fuel for controversy was recently provided by Poag and Sevon (1989) who calculated the volumes of Mesozoic and Cenozoic sedimentary deposits of the U. S. middle Atlantic continental margin. Their data indicate that a very large amount of sediment has been eroded from the Appalachian source area since the end of Early Miocene time, following a long period of erosional quiescence. Braun (1989) used these data for a backfill estimate and argued that at least 0.7 mi (1.1) km of material must have been eroded from the Appalachian source area since the end of the Early Miocene in order to account for the volume of continental margin sediment. His analysis indicates that about 500 ft (150 m) of material would have been removed in the Pleistocene. If the offshore data and Braun's analysis of them are reasonable estimates of past erosion, then the upland parts of Piedmont landscape cannot be as old as previously thought by many, the ideas of Pavich have considerable merit, and the insight of Knopf and Jonas is remarkable.

SUMMARY

The evolution of Piedmont landscape in the Holtwood area is not yet totally understood. It appears that only small parts of the upland topography, those at major drainage divides, have any possibility of being remnants of a former peneplain. Even those remnants are presumably being lowered steadily and may not even appoximate the elevation of the original surface. We have estimates of the age of that proposed surface which range from Cretaceous to Pleistocene. The thick saprolite present throughout the area apparently is not the result of ancient weathering, but is

a product of continued weathering under many different climates. The entrenchment of the lower reaches of streams tributary to the Susquehanna River is part of the present erosion cycle which was initiated at an unknown time in the past. The message about this erosion cycle has not reached the uplands far removed from the mouths of the streams. The landscape appears to be evolving at a very slow rate, but available data suggest that the rate of change is much faster than we previously thought.

Do we now know any more about the development of landscape in the Holtwood area and the Piedmont of Pennsylvania in general than did Knopf and Jonas in 1929? It is difficult to tell. Lacking a method for absolute dating which can be applied to surfaces such as those which occur in the Piedmont, we still rely on indirect methods. However, as more and more different types of data are evaluated, we are able to ask better questions and we hope that better answers will be forthcoming. Consider this an interim report of progress.

V. GEOMORPHOLOGY OF THE LOWER SUSQUEHANNA RIVER GORGE

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Elizabethtown College

INTRODUCTION

The Susquehanna River provides drainage for approximately 27,000 square miles (70,000 sq. km), most of which is in Pennsylvania. It consists of a major trunk formed at the confluence of its North and West Branches then joined downstream by its largest tributary, the Juniata River. Together with the combined discharge (av. 20,000-30,000 cfs) of minor tributaries, the Susquehanna subsequently flows southward through a major gorge to its mouth in Chesapeake Bay. Except for the contributions by streams entering the gorge itself, all water passing through originates in diverse physiographic provinces including Appalachian Plateau, Ridge and Valley and Piedmont. Extreme climatic fluctuations characterizing the Pleistocene also must have affected discharge characteristics through the gorge.

The Susquehanna Gorge begins near Washington Boro at Turkey Point (Figure V-1) and extends southward for 35 miles (56 km) to Perryville, Md. In Pennsylvania, the river acts as a physical boundary between Lancaster and York Counties and has thereby profoundly influenced both culture and history. The gorge depth (to present water surface) varies irregularly from 200 feet (60 m) to 515 feet (157 m), the latter being at a point known as Pinnacle Hill, or simply "the Pinnacle." The river width at Washington Boro is approximately 1.6 miles (2.6 km); it narrows in the gorge to 0.23 miles (0.37 km) at the Pinnacle, thus reducing that dimension by a factor of seven. The river bed gradient through the gorge is approximately 6 ft/mi (1.15 m/km) which is considerably steeper than average upriver trunk gradients of 2.7 ft/mi (0.5 m/km). Precipitous walls are composed of Piedmont metamorphics, including the Wissahickon and Petters Creek Schists.

Three gorge hydroelectric impoundments--Conowingo (Md.), Holtwood and Safe Harbor--take advantage of the local river gradient, and their waters have obscured much of the geologically interesting bedrock river bed. The uniqueness of generating electricity by a singe-site combination of hydropower and coal-fired steam is discussed by Inners et al. (1978). Muddy Run, a gorge tributary, is the site of a hydroelectric pump storage operation. Its dam creates a reservoir some 400 feet (122 m) above the main river.

The gorge, with its craggy walls, impounded waters, steep ravines with tumbling streams and adjacent plateau-like highlands, is a prime area for recreation. The electric companies permit aquatic activities on the lakes and have provided facilities for picnicing, hiking and camping. A state park, Susquehannock, also is available for outdoor activities. Furthermore,

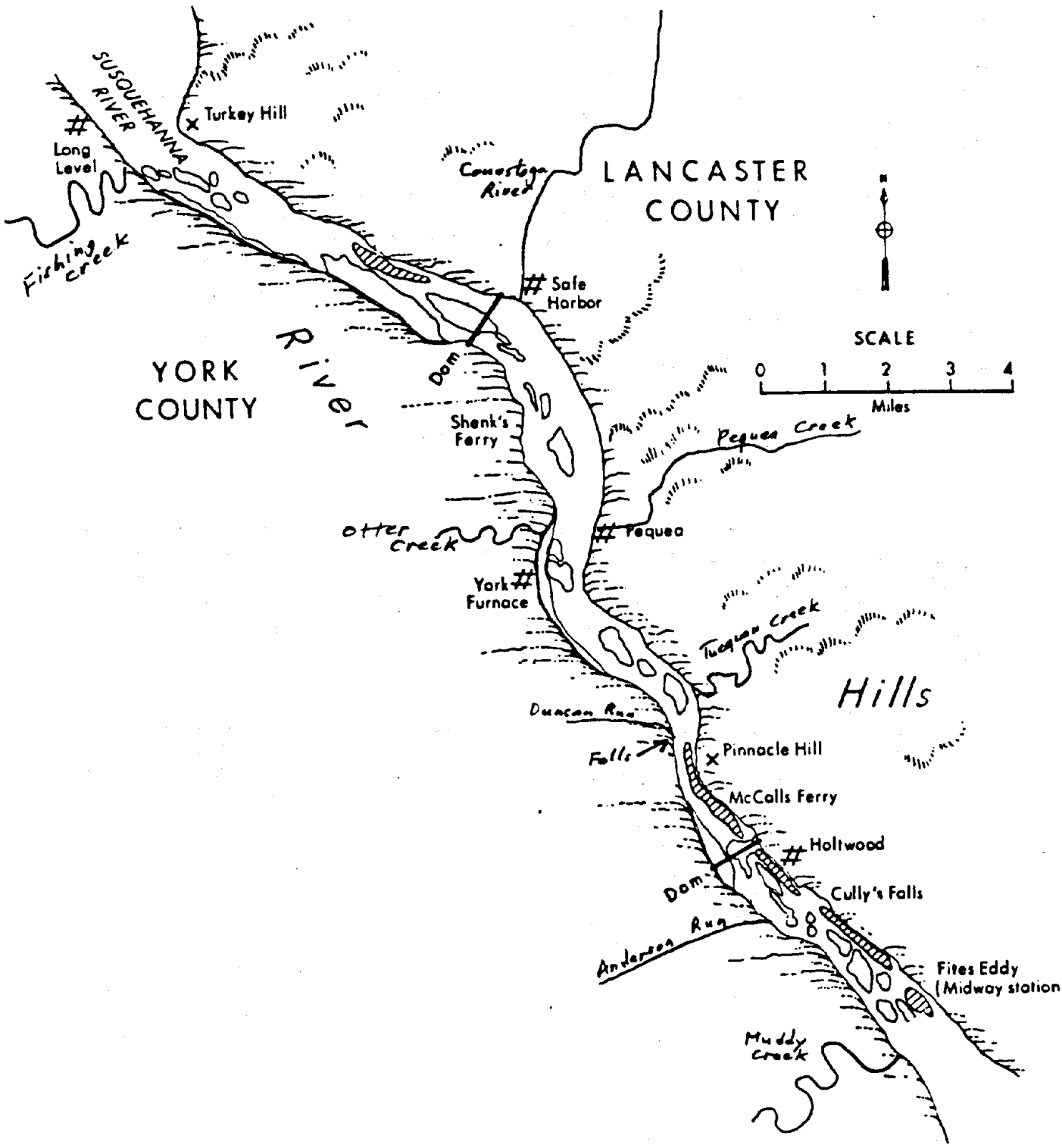


Figure V-1. General map of the lower Susquehanna River gorge. (From AEG 1978 Field Trip Guidebook)

Tucquan Glen is preserved as a natural area. Primary access is gained by PA Route 372 which crosses the central gorge area on the Norman Wood Bridge. Walking this bridge provides excellent views of the river bed, especially at times of low water. Additional public overlooks are available on the Lancaster County side. These include Safe Harbor, the Pinnacle, Holtwood and Susquehannock State Park. Trails are abundant on both sides of the river, and there are several public boat launching facilities and marinas.

Past and current studies of the general geology of the gorge and surrounding areas is discussed by Sevon in Chapter IV of this guidebook. In addition to general reports, several special topics have been investigated and reported on. These include "the deeps" (Mathews, 1917), terrace gravels (Stose), the Westminster anticline (Campbell, 1933), a paleofalls system (Thompson, 1985), river bed erosion (Sevon and Thompson, 1987) and some comparative hydraulic studies (Thompson, 1988).

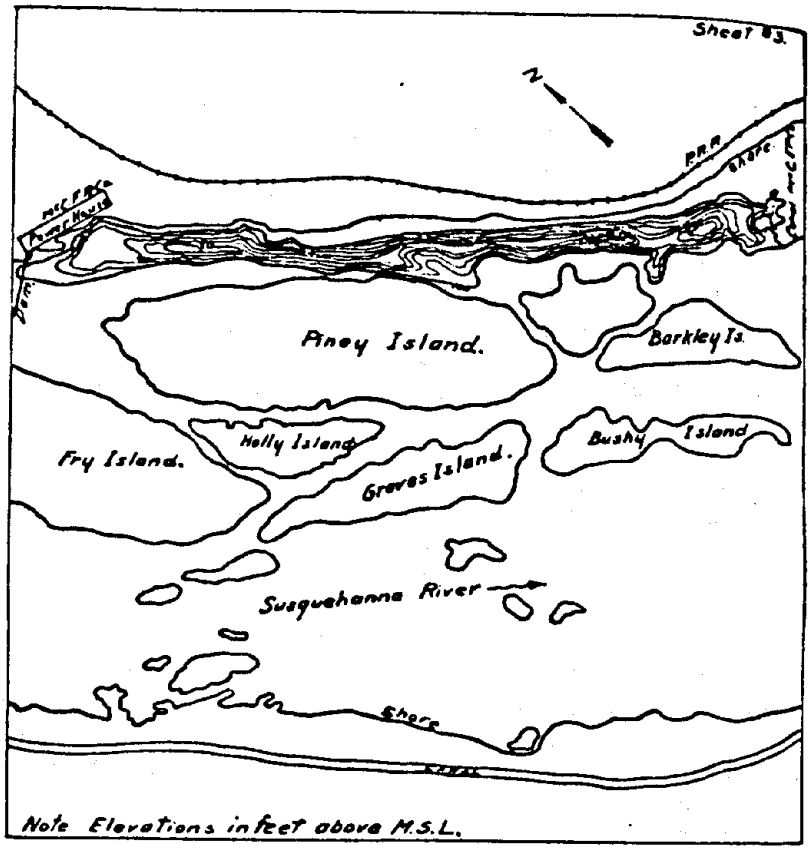
GEOMORPHIC FEATURES IN THE LOWER GORGE

The Deeps

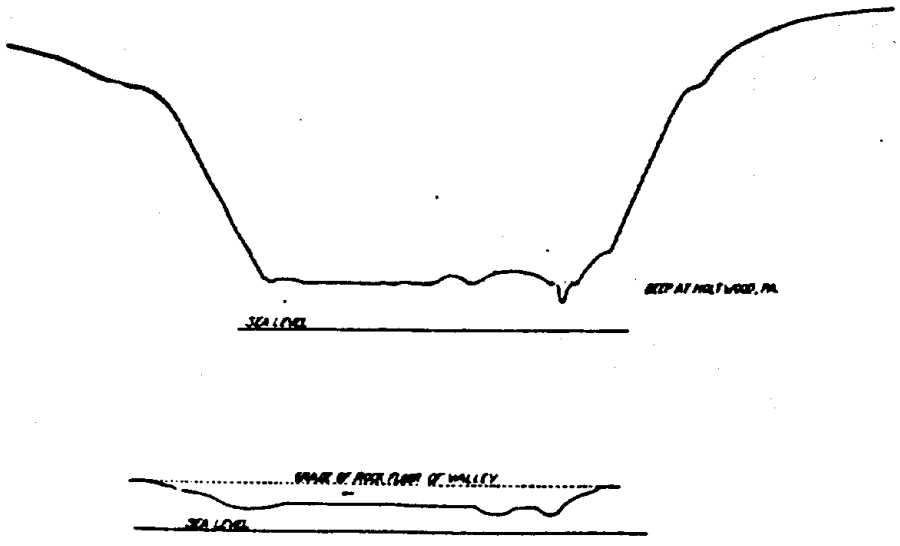
Pre-construction engineering studies for lower Susquehanna hydroelectric impoundments have provided river bottom survey information of better than usual detail (Figure V-2). Based on these data and on observations made of cofferdam-protected river bed reaches drained during dam construction, E. B. Mathews (1917) wrote:

> The portion of the survey under present consideration extends from Turkey Hill, 3 miles south of Washingtonboro, Pennsylvania, to tide near Port Deposit, Maryland. Throughout the entire distance the river flows in a flat-bottomed rock gorge with stream-cut walls, which rise to the general level of the Piedmont Upland. The river bottom is generally studded with numerous rocky islets, which rise but a few feet above the normal river surface, and a few steep-sided islands, whose wooded tops may reach 100 feet above the water. Under ordinary conditions the bed of the river is covered with less than 15 feet of water, and in dry season may be largely exposed as a rock floor from one-half to one and one-half miles in breadth. Within this flat bottom of the broad gorge the survey discovered six long spoon-shaped depressions, some of them over 100 feet deep, with their deepest portions extending below tide level.

Mathews further described each "deep" in detail, including one which had been observed while it was drained in preparation for becoming the present tail race for the Holtwood Dam. He continued:



-Map of "Deep" at Holtwood, Pennsylvania
Scale: 1 inch = 1,000 feet



-Profiles of "Deep" at Holtwood, Pennsylvania
Scale: Horizontal, 1 inch = 2,000 feet; vertical, 1 inch = 500 feet

Figure V-2. From Mathews, 1917

This "deep" lies close to the left bank of the Susquehanna, between it and Piney Island, and has been utilized by the engineers as a tail race for their power plant. During construction of the dam it was exposed by a diversion of the water to a depth of nearly 50 feet. The water surface was about 110 feet above tide and the rock floor about 100 feet. From the latter rise Fry and Piney Islands to a height of 140 and 160 feet respectively. The hills above the power plant rise rapidly to an elevation of over 500 feet. This depression is a gorge of 4,000 feet length, with a width of from 200 feet to 300 feet within the rock floor of the river, which at this point is about 100 feet above tide. The general level of the bottom of the gorge is 60 feet above tide, or 40 feet below its rim, and shows three local depressions (Figure V-2). That opposite the upper end of Piney Island reaches to 50 feet above tide, while the two at the lower end, opposite Barkley Island, reach 40 feet above tide. The rock barrier between it and the foot of Culleys Falls was removed, so that it is now continuous with the "deep" described later. The withdrawal of the water gave exceptional opportunity for studying the walls. Everywhere were deep vertical pot-holes of varying diameter and perfection, so closely placed that they suggested the fluting of a pipe organ or the fracture of a block by the use of "plug and feathers." Some of the pot-holes extended below water level, while other showed nests of boulders part way down the side of the gorge.

Mathews went on to summarize his observations:

> Their peculiarity lies in their extreme ratio of length to breadth, their depth of cutting (at times below sea level), and their bottom profiles, which rise downstream and do not persist as canyons.

It seems obvious that the "deeps" have an origin in the hydrodynamics of fluvial bedrock erosion. Investigators yet wonder about the particular conditions which have fostered their development. Speculations run rampant, and good questions are wanting. Certainly to be considered are such factors as flood turbulence and frequency, water depth, bed and suspended loads and the possibility of ice influence. Are the "deeps" genetically related to present day river conditions, or are they the product of hydrodynamics developed under different climatic regimes?

Certain facts are known. The "deeps" are uniquely located (Figure V-1) in the gorge area. They are not connected as a continuous channel, thus ruling out the possibility that they are merely relicts of a narrow water course carved during times of low sea level. According to silt monitoring measurements (L. Brethauer, former superintendent at Safe Harbor Dam, pers. comm.), even under ponded conditions they are not becoming filled with sediment. Finally, the "deeps" are all on the eastern side of the river. This fact has caused the present writer to

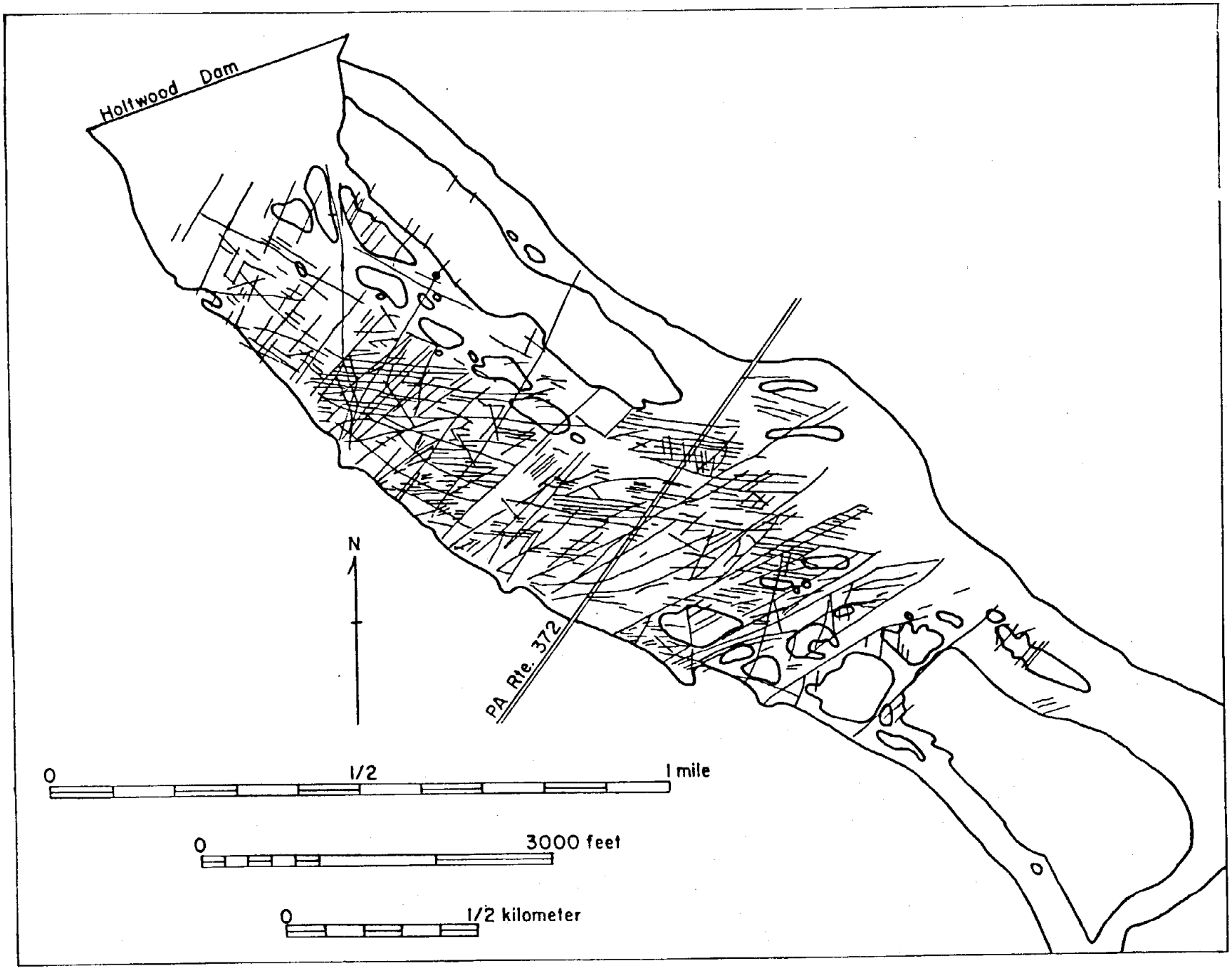


Figure V-3. Fracture pattern of rock exposed in the bed of the Susquehanna River below the Holtwood Dam. Interpretation by W. D. Sevon from aerial photograph.

hypothesize that they may be influenced by the noticeably warmer microclimate on that side. This would, in cold times, lead perhaps to the only location of open channel, thus producing what W. Sevon (Pa. Geological Survey, pers. comm.) has termed "ice focusing" of turbulent conditions.

Bedrock Islands

Upriver from Columbia, the Susquehanna River is typified by shallowness and scattered alluvial islands. Some bedrock islands appear as resistant ledges of upturned formations or arise from the influence of igneous intrusions; an example is Hill Island near Goldsboro. In contrast, the gorge area is studded exclusively with islands composed of unremoved portions of local bedrock. These islands are shaped in plan to suggest hydrodynamic process influence and are dramatically modified by joint-controlled channeling (Figure V-3). A casual inspection reveals several distinct levels of their summits, some with nearly accordant heights. Also, their heights tend to increase in the downstream direction. These height characteristics have been interpreted (Thompson, 1985) as relict portions of ancient flat river bed levels produced and abandoned due to flood erosion of a migrating falls/rapids system not unlike the present Great Falls of the Potomac. Regardless of historical sequence interpretations, the islands clearly exist because channeling has lowered the presently active river bed.

Closer scrutiny of the islands produces two additional features of interest. First, there are multitudes of potholes. These vary in size from tiny to enormous, the largest observed being nearly 7 meters (23 ft) deep and 3 meters (10 ft) in diameter at the base. It has been suggested (Sevon and Thompson, 1987) that these potholes (Figure V-4) have aided river bed erosion by weakening jointed sections to the point where hydraulic plucking could have removed whole blocks at a time, leaving dissected pothole voids on the channel margins.

Second, large rounded boulders, often exceeding a meter in diameter, are found scattered about on the island tops. These are proposed to be "fluvial erratics" deposited as bedload when present island tops were portions of ancient river beds (Thompson, 1985). The boulders were subsequently stranded as channeling lowered the active bottoms. Though some boulders are of local schist derivation, most are from identifiable sources up river as shown below:

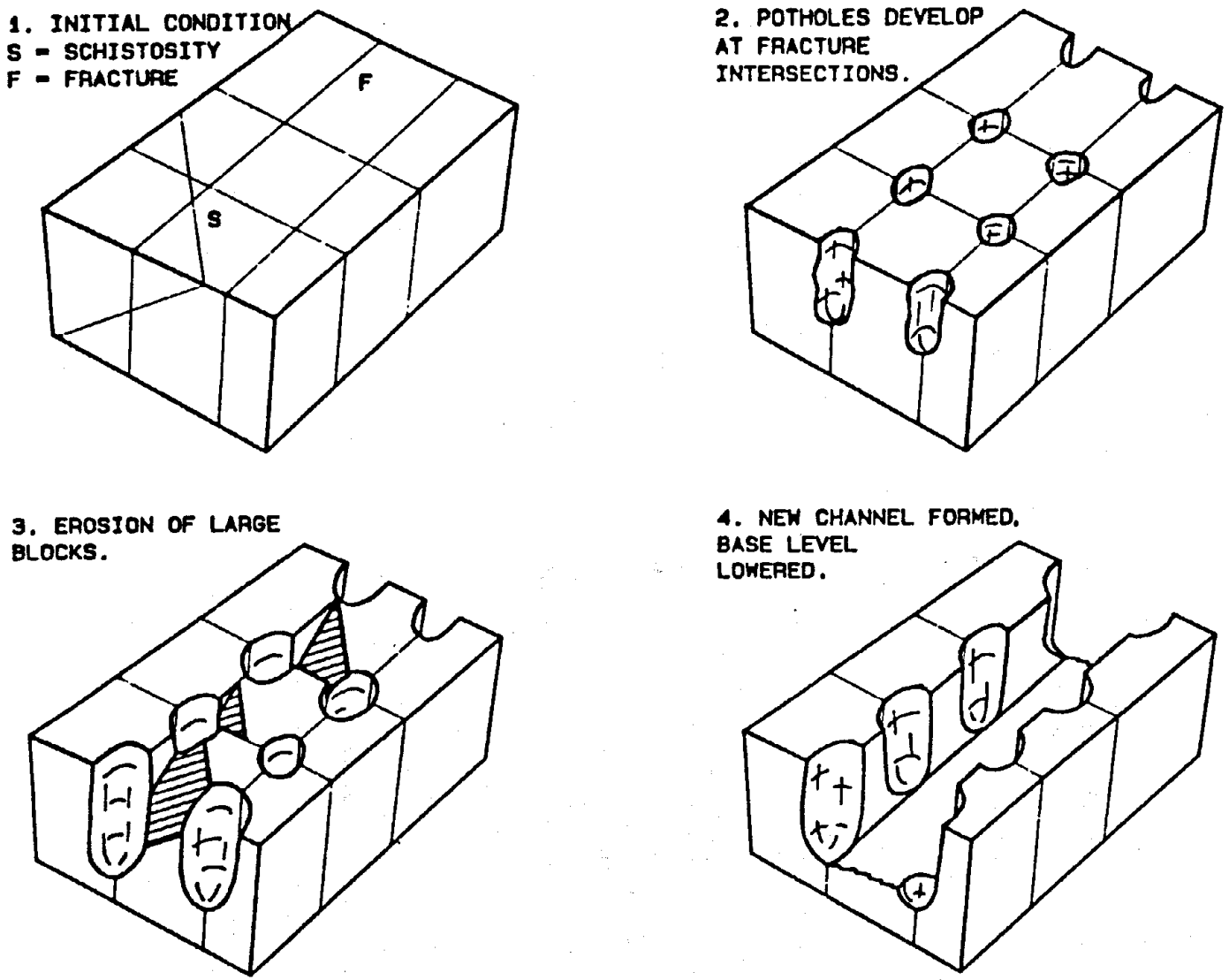


Figure V-4. River bed erosion by pothole-assisted block removal (from W. D. Sevon).

Boulder Lithology	Source Formation	Upriver Distance in miles (km) from Peavine Island
diabase	Mesozoic intrusions	
	Safe Harbor	9 (14.5)
	Conewago Falls	34 (55)
quartzite	Chickies Fm.	21 (34)
conglomerate	Pocono Fm.	55 (88.5).

It is often suggested that these erratics are the result of ice rafting. This writer does not think that to be the case. The boulders are almost exclusively found trapped in depressions, including potholes, where they would have become lodged during rolling; rafted boulders would not be dropped so selectively. Rounding further suggests normal abrasion; rafted boulders should be plucked and angular. Much more research on this question obviously is needed.

Tributaries

The tributaries which feed into the gorge area range from small to moderately large. These include unnamed creeks less than a kilometer long plus longer named streams such as Otter Creek (22 km), Pequea Creek (77 km) and the Conestoga River (107 km). These tributaries display, without exception, convex-up longitudinal profiles in their lower reaches. In moderate to longer tributaries, the upper reaches display gentle gradients and well-developed floodplains. As the streams approach the main river, they steepen and in some cases tumble and fall directly into the gorge. To a limited extent, these features may be observed in Anderson Run, a 4 km long stream paralleling Rt. 372 immediately west of the Norman Wood Bridge (STOP 4). In addition to the profile characteristics described above, some of the streams, such as Tucquan and Otter Creeks, are contained in what appear to be incised meanders. It is problematic whether the incision is inherited from ancient floodplain meanders or is a reflection of bedrock structural control. The latter alternative is favored by C. Scharnberger (Millersville University, pers. comm.) who, with his students, has mapped and compared joint patterns with stream patterns.

One analysis of tributary profiles (Thompson, 1985) concluded that the convex-up profile character was generated when high-energy glacial meltwaters from the upper parts of the Susquehanna drainage basin invaded the gorge area, causing rapid bed erosion and attendant lowering. This left the tributaries "hanging," unadjusted to the main trunk channel, each with a nickpoint which would migrate upstream at a rate reflecting local discharge and bedrock resistance. In support of this conclusion, it is easily observable on tributary profiles that the inflection points where normal concavity changes to convexity are

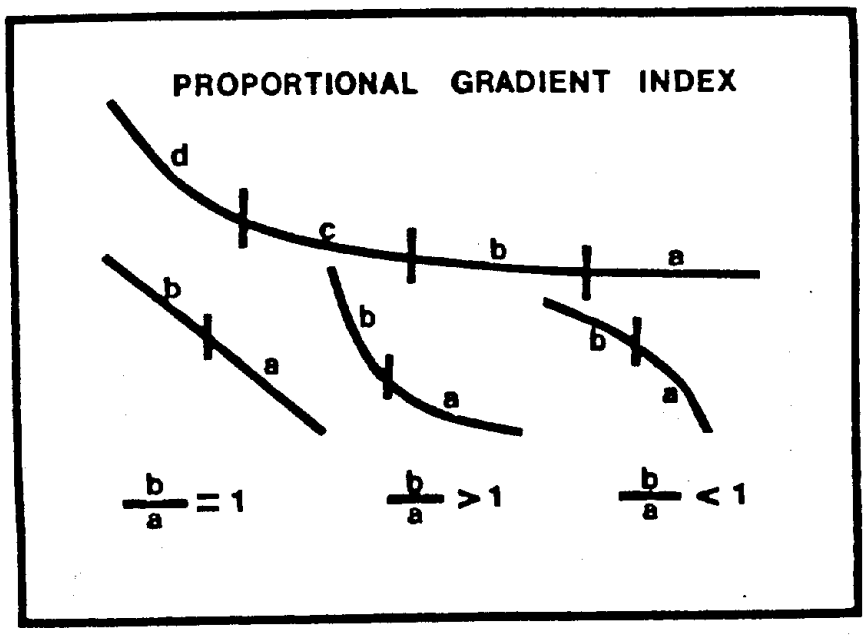


Figure V-5. Determination of PROPORTIONAL GRADIENT INDEX (Thompson, 1985). The long profile of any stream is divided into quartiles, and the gradients of the lower two (a & b) are then calculated. Their gradients are set in the ratio form: a/b, thus yielding a dimensionless index. A PGI > 1 is concave-up or normal; a PGI < 1 is convex-up or abnormal. The departure value from 1.0 is a relative index of concavity or convexity. This system treats all sizes of streams equally by selective emphasis on the lower 50%, thus making it proportional.

located a distance upstream from the mouth in direct proportion to the size of the streams themselves.

To further analyze profile curvature, a method called Proportional Gradient Indexing (PGI) was devised (Thompson, 1985, 1988). The PGI is designed to examine stream reaches most likely to be affected by lowering of local base levels; that is, the downstream quarter (25%). Thus, the same proportion of the total length of each stream is analyzed for comparative purposes. The PGI for any given stream is established by first determining the gradients of its two lower quartiles. The gradients are then placed in ratio form in such a way that the departure from a ratio of 1.0 is a relative index of concavity or convexity (Figure V-5). The results of this analysis are given below in comparison with another major river, the Potomac.

The foregoing conclusion, that the Susquehanna River was rapidly modified by Pleistocene conditions, recently has been bolstered and criticized. John Shaw (1989) of Queen's University, Kingston, Ontario, has proposed that outbursts of subglacial meltwater, discharging at rates approaching 10 million cubic meters per second, not only would account for drumlin formation and for Canadian bedrock scour features, but also would

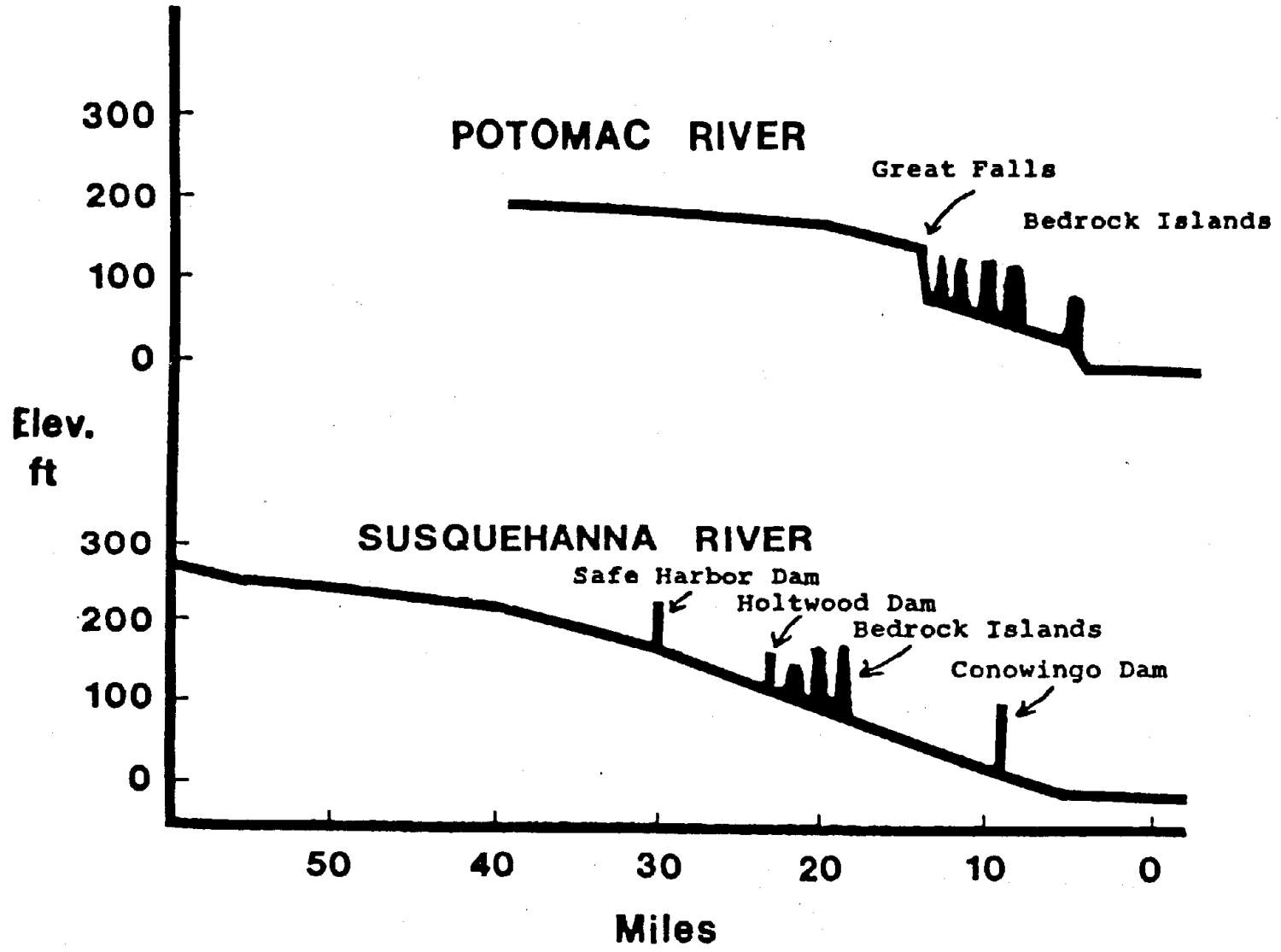


Figure V-6. Longitudinal profiles of the Potomac and Susquehanna Rivers.

have left their marks on drainways to the south. He suggested the Susquehanna River as one probable recipient of these very high discharges. Other researchers (Mullins and Hinchey, 1989) have suggested that the valleys of the New York Finger Lakes may have been carved, not by ice directly, but by streams of highly pressurized subglacial meltwater. These lake channels aim directly for the Susquehanna watershed. Duane Braun (1990) of Bloomsburg University holds an opposing position. His research area lies in the middle portion of the Susquehanna River valley, between the sites of proposed high discharge outbursts and the sites of proposed erosive results of such events. Channel restrictions at Bloomsburg should have metered the postulated floods, thus producing slackwater deposits and armored expansion bars. He finds none. Instead, he reports the presence of loess, colluvium and pre-Wisconsinan glacial deposits, all of which should have been washed away by large floods.

COMPARATIVE ANALYSES

Potomac River

The time and space scales of many geological phenomena are beyond the means of typical controlled laboratory experimentation. An alternative means of investigation is comparative analysis, one example or situation serving as a control for another. In the case of the geomorphic investigations of the lower Susquehanna River, the Potomac River has been chosen as the control (Figure V-6). It is similar to the Susquehanna River in that it heads in highlands to the west, traverses identical physiographic provinces and, by proximity, drains watersheds of similar present climate. Though the Potomac has a smaller discharge, its primary departure from Susquehanna characteristics is that it had no Pleistocene continental glaciers in its drainage basin.

This writer initially hypothesized that the Susquehanna gorge area once held a falls/rapids system similar to the Great Falls of the Potomac, and that the former was mostly destroyed by fluvial erosion intensified during Pleistocene time. The bed of the Potomac River also has been modified through time, though much less rapidly. In the case of the Great Falls, the assumption is made that, as the falls nickpoint migrates upstream, tributaries formerly entering the river adjusted to local base level will be left to plunge into the lengthening gorge. The average gradient of any tributary so bypassed would be increased. Figure V-7 clearly demonstrates that, for the Potomac River, tributaries upstream of the falls have gradients less than those that enter the gorge below the falls.

This analysis, applied to the Susquehanna River, produces results that are less certain, with several influences being suspect (Figure V-8). First, no clear dividing line, i.e., falls location, exists. Second, if Campbell (1929, 1933) was correct in detecting and interpreting warped terrace gravels, some

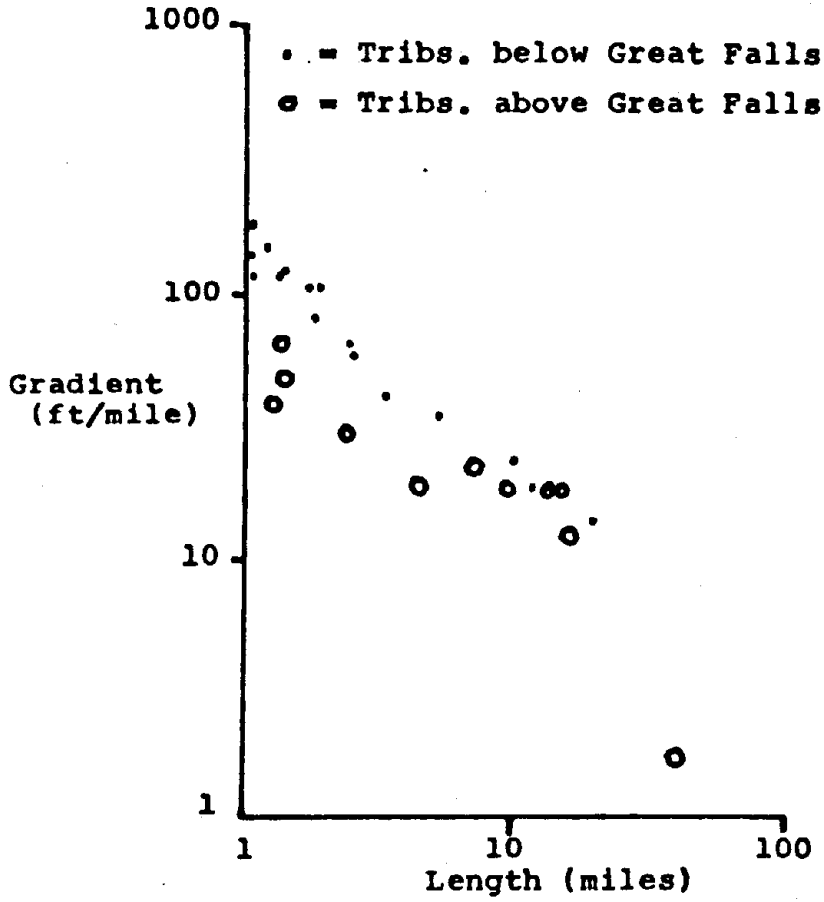


Figure V-7. Stream gradient vs. total length for tributaries of the Potomac River near Great Falls, Va.

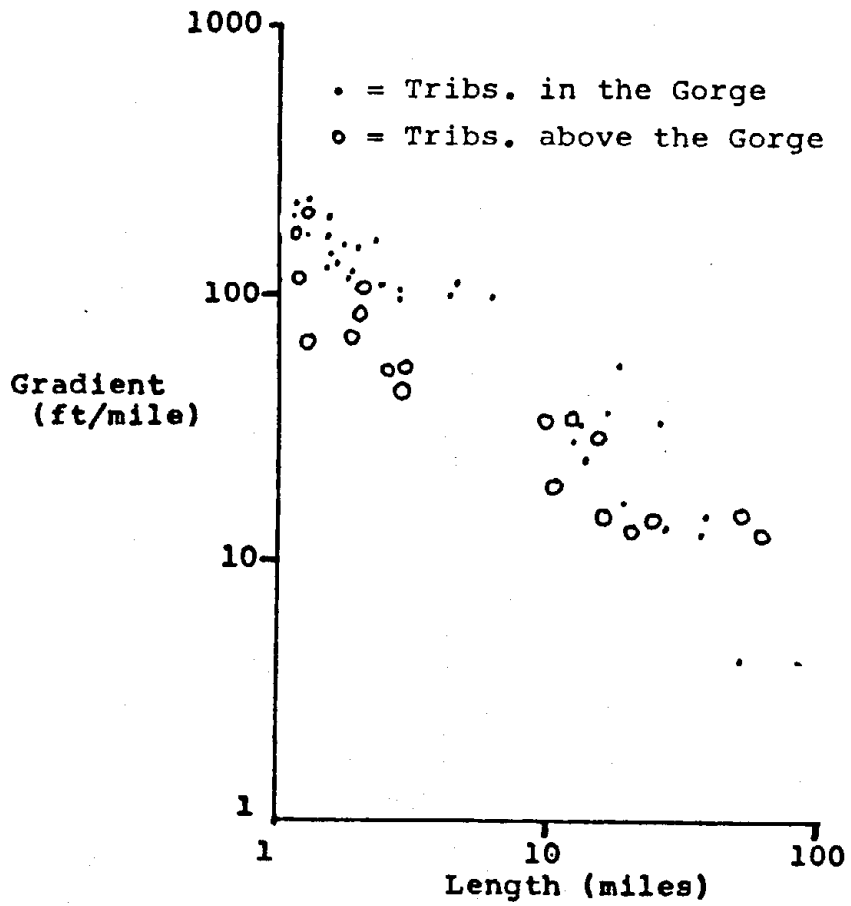


Figure V-8. Stream gradient vs. total length for tributaries of the Susquehanna River near Holtwood, Pa.

Pleistocene arching has occurred. And a third complication is that inundated tributary mouth elevations could be found only by extrapolation to trunk profiles drawn so as to connect points of known elevation in a logical fashion. Despite these complicating factors, a trend similar to the Potomac model can be demonstrated. This is interpreted as evidence that a well-defined nickpoint, probably in the form of a falls/rapids system, did previously exist in the gorge of the Susquehanna. Its most likely location was the area below Holtwood, now studded with bedrock islands formed as the falls/rapids were destroyed by intensified fluvial channeling processes. This nickpoint also is reflected in the profile of the Susquehanna and Tidewater Canal (see Chapter VI), where, in a distance of 1.5 miles, four locks were required to lift boats 34 feet. The increased rate of erosion, compared to that of the Potomac River, is attributed to intensified hydraulic activity associated with glacial meltwaters. Flood magnitudes and frequency remain in question.

The bedrock islands show several levels of accordant heights, a possible result of periodic erosion intensification. This pulsed model would correlate nicely with periodic glacial activity and would, by simultaneous falls migration and general river bed lowering, produce the highest island tops (ancient river beds) in the downstream direction. Hennery Island extends 200 feet (61 m) above the river bed, leading this writer to envisage a Pleistocene downcutting minimum of 200 feet. It is unlikely that rising island heights in the downstream direction are a result of Campbell's (1939) Westminster anticline, because he places its axis at Safe Harbor, 10 miles (16 km) upriver from Hennery Island. If the supposed upwarp did anything, it reduced the height differential between Holtwood and Hennery Island.

A final comparison with the Potomac River uses the Proportional Gradient Index (PGI) method for assessing concavity versus convexity of longitudinal tributary stream profiles. Concave streams are assumed to be normal, and those which are convex are assumed to be out of adjustment with their recipient trunk. Comparative results of the PGI analysis are shown in Figure V-9.

As expected, the Potomac PGI shifts from a value less than 1 to a value greater than 1 at the site of the Great Falls. This reflects the effects on tributary profiles as they are subjected to abrupt base-level lowering by main-trunk falls migration. For the Susquehanna River, the PGI values are consistently less than 1, and display a marked further reduction in the gorge at Holtwood. This suggests that the entire lower Susquehanna River was subject to rapid bed lowering with pronounced changes concentrated in the gorge proper, probably by reduction of a falls/rapids zone. The regional effect of this river-bed reduction was to leave the energy-deficient tributaries "hanging." If tectonic arching occurred in the Pleistocene, as suggested by Campbell (1939), this would mean that the original effects were even greater than those presently measurable.

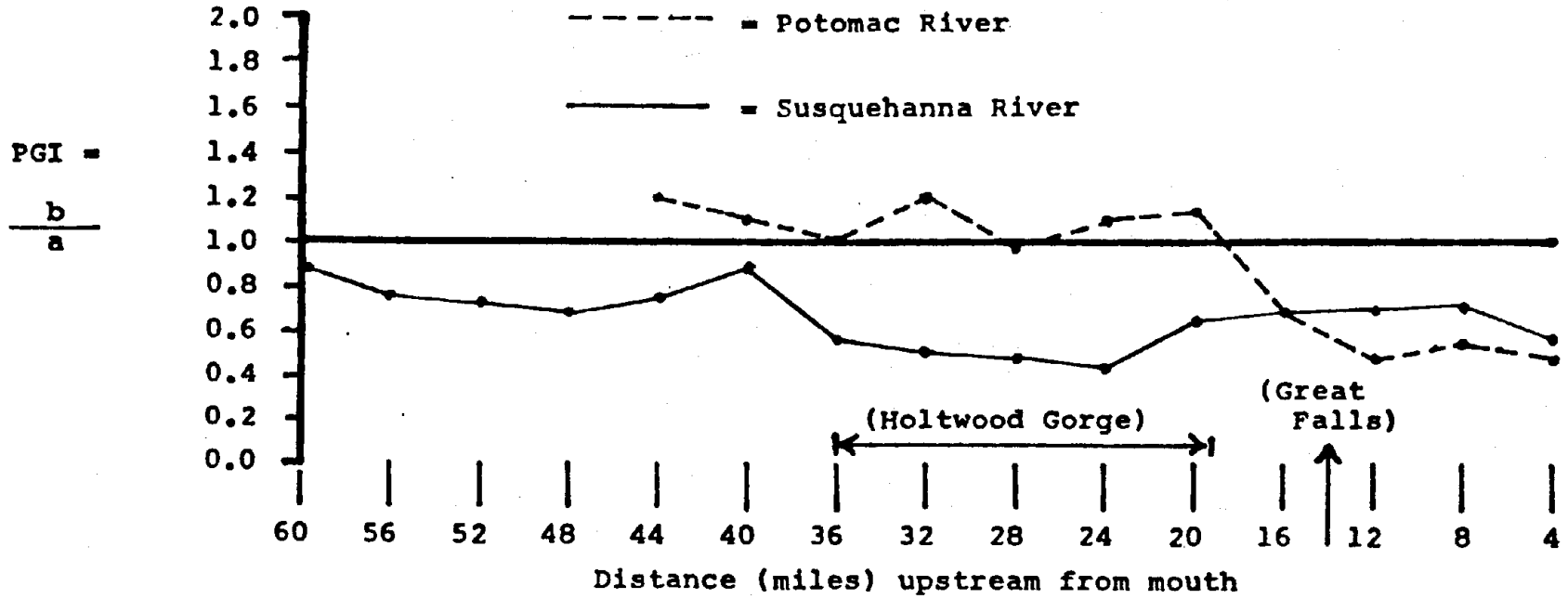


Figure V-9. Average Proportional Gradient Indicies (PGI's) for the Potomac and Susquehanna Rivers. Each point represents the average PGI for all tributaries entering the 4 mile reach on the adjacent downstream side.

The Channeled Scablands

The specific features of the bizarre landscape called the "channeled scablands" of Washington were, in 1923, first cast into a flood-produced scenario, an outrageous hypothesis promoted by J. Harlan Bretz of the University of Chicago. The resulting controversy and the long story of documentation of catastrophic flooding have been adequately presented by Victor Baker (1978). Even today there are arguments, not about the flood origin itself, but about the number of floods, reaching perhaps forty or more (Waitt, 1980, 1984, 1985). Because the scabland features generally are regarded as having formed from the effects of catastrophic flooding, they have been compared by this writer with features of the lower Susquehanna River valley. This effort was made in order to gain insight into the question of the hydraulic magnitudes needed to create the Susquehanna features.

Hydraulic erosion features in the scablands include gigantic potholes, dry waterfalls, longitudinal grooves, rock terraces and dramatic channeling in the form of coulees. The bedrock is highly jointed columnar basalt, layered from a series of extrusive events. There are, in addition, fluvial deposits, including megaripples, pendant bars, slackwater silts and fluvial erratics. These are well described by Baker and Nummedal (1978). Also present are loess-mantled interfluves, untouched by anastomosing floods.

It is difficult to make valid comparisons between scabland erosion features and those of the lower Susquehanna. In the scablands, the tremendous energies of the flood waters acting on strongly jointed bedrock have produced features that seem to have resulted more from plucking than from scouring. Nevertheless, some similarities suggest a common origin for some features, especially the potholes.

Years of pothole observation have led this writer to the generalization that potholes are concentrated in certain locations, mostly dependent on situations which would create water turbulence. Contributing factors are bedrock obstructions and periodic flood discharge conditions. Most intriguing is the fact that the largest of the potholes are found, not in the main river bed (the "deeps" possibly excepted), but on ledges and terraces above the river. This is true for the scablands and is even more spectacularly developed at Taylors Falls, Minnesota, on the St. Croix River (pers. observation). That potholes also are concentrated at locations of convex-up profile crowning is easily observable at Conewago Falls on the Susquehanna River (Sevon, 1989), the Great Falls of the Potomac, and at the falls of the James River in Richmond, Virginia. It is, therefore, suggested that these characteristics, useful in predicting pothole locations, are significant clues to their origin.

Potholes are bedrock voids, more or less circular, cylindrical to slightly conical in shape, and usually are oriented in

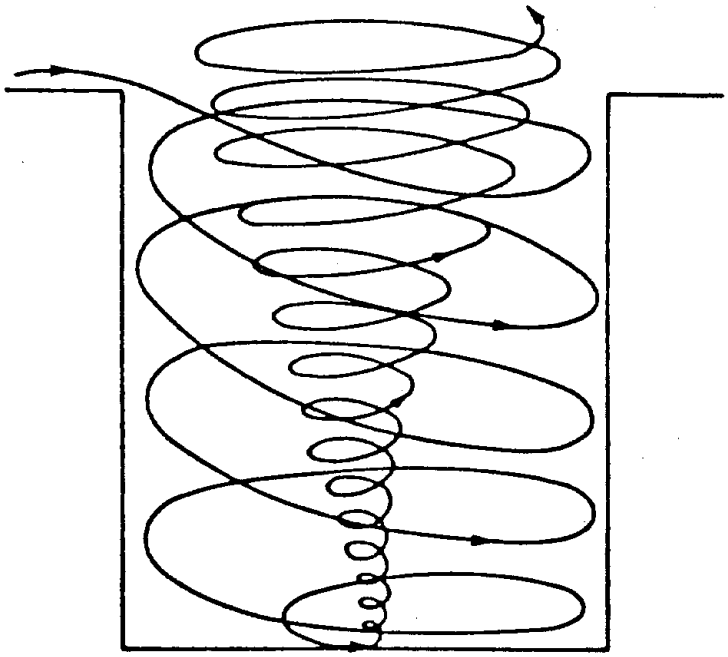


Figure V-10. Hydraulic vortex in a pothole (after Gray, Sevon).

the vertical direction. Susquehanna gorge potholes, however, commonly display axes deviating systematically from vertical. Potholes are attributed conventionally to abrasive action generated as "tools" are swirled about by rapidly rotating currents. Questions significant to pothole enlargement include the relative importance of tool grain size, solution, fluid transfer of rotational energy, cavitation, viscosity, and water depth above the void. To this list may be added the location, as discussed above, and the physics of rotational systems.

Norman Gray, at the University of Connecticut, currently is investigating hydraulic factors of pothole development and has concluded that the distribution of water pressure gradients within a pothole is significant in maintaining a vortex (Gray, 1988). His investigation focuses on vortices within rock-bound potholes, the walls of which serve to sustain pressure gradients (Figure V-10). This model, however, fails to explain the origin of a water-bound vortex which might initiate a new pothole. Furthermore, Gray indicates that a pothole vortex contains vertical components wherein rotating water descends along the pothole wall and rises in the center. This could help explain why fine-grained sediment is so seldom found in potholes. Sand and finer materials are lifted out, with pebbles being left as lag. It could also explain spiral flutes observed on some pothole walls. Vertical components of vortex flow have been suggested previously (Thompson, 1988) as the underlying cause of upwelling boils observed on the surfaces of rivers flowing in turbulent regimes. Studies in the scablands (Baker et al. in Graf, 1987) suggest that strong vortices with upwelling ("kolks"

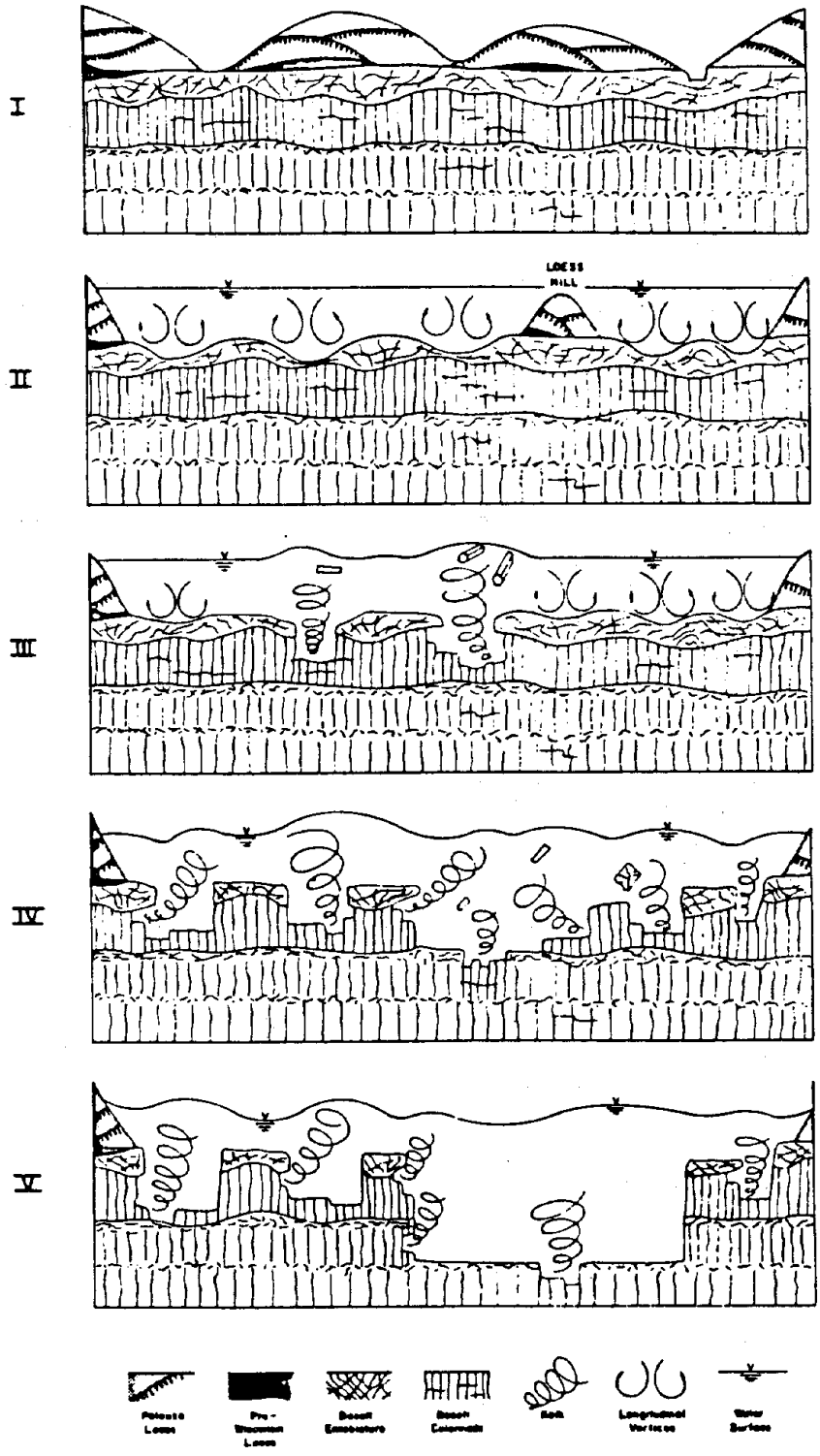


Figure V-11. Flood erosion by "kolks" in the channeled scablands (from Baker and others in Graf, 1987).

of Matthes, 1947) were responsible for plucking basalt columns loose from the bedrock and then lifting them out of the enlarging voids (Figure V-11). This may be further evidence that vortices contain a strong upwelling component.

It should be pointed out that the commonly experienced bathtub drain model of vortex development is contrary to the situation described above. An atmospheric tornado is a better approximation, and the conditions governing tornado development and behavior may be more applicable to non-draining hydraulic vortices. Rather than being the cause of rising air, tornadoes are formed from its pre-existence in convective thunderstorms. Their characteristic high wind velocities originate as slowly rotating air enters the tornado base, thus experiencing a decreased radius of curvature. The conservative property of angular momentum (the "ice skater effect") is then manifested as an increase in wind velocity.

With regard to pothole formation, it is the contention of this writer that a combination of flood discharge energies, water depth, and bedrock configuration leads to localized upwellings which, in turn, generate a series of small but violent vortices capable of spinning abrasive materials against river bedrock. Conditions promoting upwelling are suspected to exist when flood depths (undetermined at present) pass over nickpoints, be they small bottom obstructions or broad zones of bottom convexity. Also conducive to upwelling are ledges where water flows from one level down to another. One critical question concerns whether vortices act as a string series of flow-imbedded, short-term hydraulic activities, or in a standing, long-term manner. Dangerous conditions preclude direct underwater observations; however, some ingenious methods may be devised in the future to answer this question.

Once the general location for pothole development has been attained, exact location becomes critical. Sevon (1989) has argued that vortices, no matter how strong or numerous , need a place to initiate abrasive erosion, and those places usually are due to structural weaknesses, especially joint intersections. Whether or not this is true remains to be rigorously tested in the field. Intense scabland jointing and the magnitude of potholes (greater than 10 meters in diameter) make diagnostic observation there impossible.

A final comparison of flood-produced scabland landforms with those of the lower Susquehanna is quite interesting, especially because it involves the shapes of islands produced by erosion of quite disparate bedrock material, i.e., schist and columnar basalt. The shapes of scabland islands are discussed by Baker and Nummedal (1978) in their guidebook to the region. The optimal shaping there, they conclude, originated from hydrodynamic forces acting as flood waters rushed across the lower elevations of loess-mantled plains, leaving remnants characteristically streamlined, though otherwise unscathed due to their superior elevations. They analyzed several parameters, including

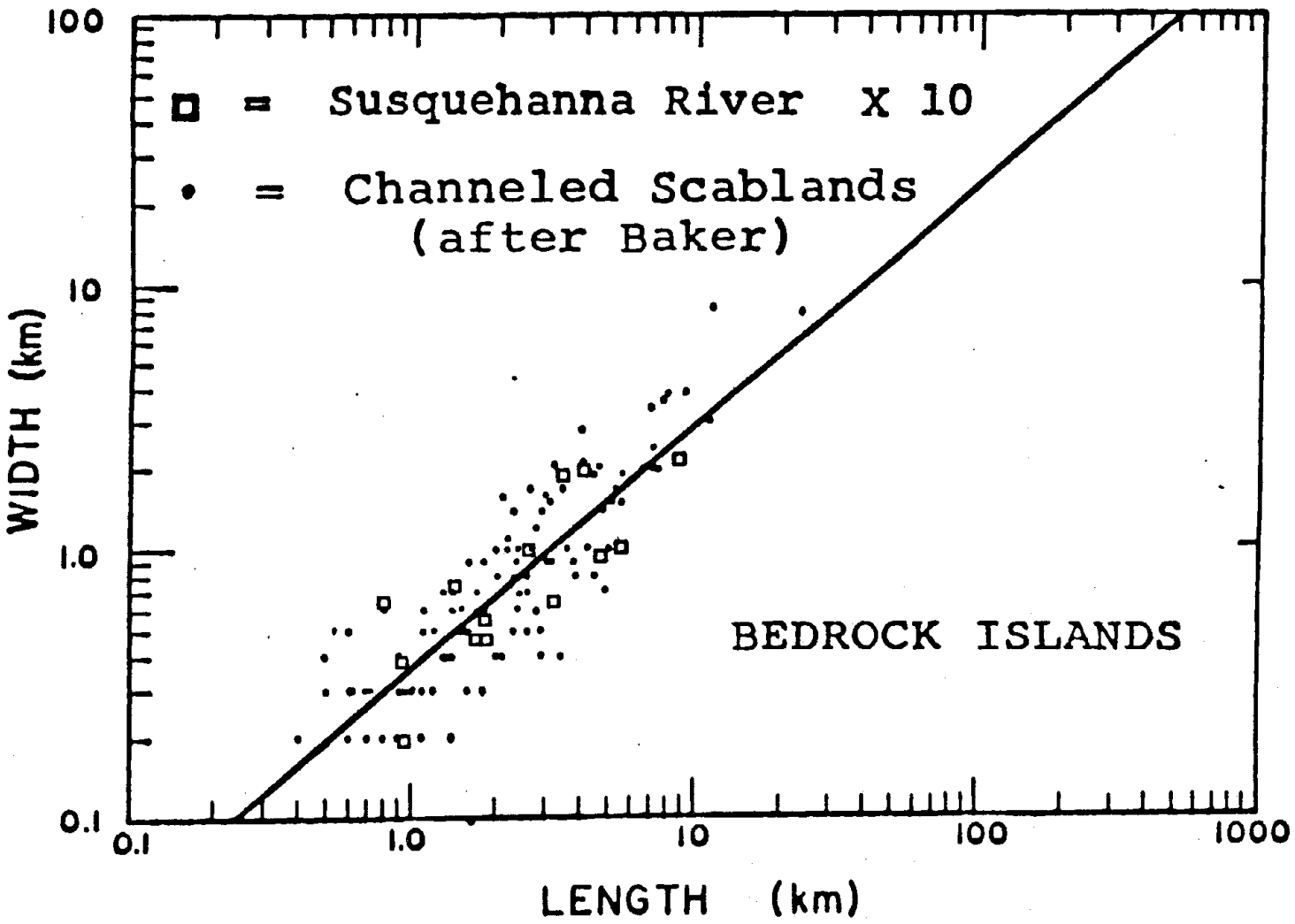


Figure V-12. Plot of length vs. width for erosionally formed islands: channeled scablands and lower Susquehanna River.

linear and areal measurements. Figure V-12 is a graph of their results on which length versus width data for Susquehanna gorge bedrock islands have been plotted for comparison. The Susquehanna island data fall well within the deviation envelope for the scabland data. This is taken as further evidence that Pleistocene flooding of greater than present-day magnitudes was primarily responsible for shaping the islands.

SUMMARY

It has been said facetiously that the first rule of science is, "If it happens, it is possible." The Susquehanna River gorge at Holtwood is the setting for a concentration of unusual and/or extreme fluvial landforms. They are there, and geological curiosity dictates a search for cause(s). It is this writer's contention that these gorge, tributary and island features are genetically related to a single cause: frequent, intensely erosive flood discharges, with Pleistocene climatic conditions being the most likely suspect. Much observational, analytical and theoretical work remains to be done at this and other morphologically similar sites.

VI. LOCK 12 AND THE SUSQUEHANNA AND TIDEWATER CANAL

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Millersville University

INTRODUCTION

From earliest colonial days the Susquehanna River has offered both the promise of easy transportation from tidewater to the interior and been a barrier to it. As is being illustrated by this field trip, the geomorphic history of the lower Susquehanna on its route across the Piedmont metamorphic rocks of York and Lancaster counties and adjacent Maryland has resulted in a deeply incised valley studded with rock islands and rapids. Today, however, in most areas this essential character is largely obscured to casual inspection by the presence of the slackwater pools of Conowingo Lake and Lakes Aldred and Clarke impounded respectively behind the Conowingo, Holtwood, and Safe Harbor power dams constructed in the first part of this century. Immediately below the Holtwood Dam, in the vicinity of STOP 4, the original character of the channel is apparent, as are the means accomplished in the 19th century to make the lower Susquehanna navigable.

The difficulties and dangers of navigating the Susquehanna River, which in earlier times was almost always in a downstream direction during high water by crude, expendable rafts and vessels carrying the products of the hinterland, were overcome in the 19th century by construction of the Susquehanna and Tidewater Canal. This canal, built between 1836 and 1839, was forty-five miles long and extended along the west bank of the river from Havre de Grace, Maryland to Wrightsville, Pennsylvania. In this distance it ascended 233 feet vertically by means of 28 lift locks. Because of the usual deteriorations of time, but especially due to flooding behind the power dams, relatively little physical evidence of the canal now remains. In the Lock 12 Historic Area, immediately downstream from the Holtwood Dam in the vicinity of the Norman Wood Bridge (PA Route 372), several locks have been preserved, their remains stabilized by the Pennsylvania Power and Light Company. Lock 12 at our lunch stop (STOP 4) is the most easily accessible of these.

While travelers aboard canal packet boats often reported squalid conditions, poor food, and other factors endemic to a slow and primitive means of transportation, the park-like setting of Lock 12 gives a feeling for the special aura of canal travel and the sense of well being that it could engender. These feelings were well recorded by Charlies Dickens in reporting on his 1842 travels on the Pennsylvania "Main Line" canal system:

> Between five and six o'clock in the morning we got up, and some of us went on deck. The washing accommodations were primitive. There was a tin ladle chained to the deck, with which every gentlemen who thought it necessary to cleanse himself (many were superior to this weakness) fished the dirty water out of the canal, and poured it into a tin

basin, secured in like manner. There was also a jack-towel. And, hanging up before a little looking glass in the bar, in the immediate vicinity of the bread and cheese and biscuits, were a public comb and hair-brush ....
And yet, despite these oddities ... there was much in this mode of traveling which I heartily enjoyed .... Even the running up, bare-necked, at five o'clock in the morning, from the tainted cabin to the dirty deck; scooping up the icy water, plunging one's head into it, and drawing it out all fresh and glowing with cold; was a good thing. The fast, brisk walk upon the towing path between that time and breakfast, when every vein and artery seemed to tingle with health; the exquisite beauty of the opening day, when light came gleaming off from everything; the lazy motion of the boat, when one lay idly on the deck, looking through, rather than at, the deep blue sky; the gliding on at night, so noiselessly, the shining out of the bright stars, undisturbed by noise of wheels or steam, or any other sound than the liquid rippling of the water as the boat went on; all these were pure delights .... (Dickens, 1842).

HISTORY OF IMPROVEMENTS ON THE LOWER SUSQUEHANNA

Following permanent settlement, and prior to the canal era of the 19th century, the amount of freight that could be transported by various craft on the rivers of eastern North America progressed rapidly from the capacity of individual canoes to that of specially designed batteaus of 2 to 10 tons, to larger poled keelboats (called Reading or Durham boats) of 8 to 20 tons, to arks or flatboats of 10 to 50 ton capacity. On the lower Susquehanna such large arks were being run downstream to tidewater by 1790 (Baer, 1981). These vessels, as well as similar rafts of logs or squared timbers, were broken up upon arrival at their destination on the Chesapeake because a return upstream, even by smaller craft, was nearly impossible under prevailing natural conditions.

The earliest improvement to navigation on the Susquehanna by means of canals was at the Conewago Falls south of Harrisburg, near Three Mile Island. At that point the Susquehanna drops 19 feet in a short distance as it passes over the southern edge of the outcropping Triassic-Jurassic diabase sill there. The one-mile-long Conewago Canal around the west (York County) end of the falls was completed in 1797 and was operated until about 1840, after which it served primarily as a mill race until its dismantlement in 1885. Two brick-lined locks at the lower end provided a controlled descent of 20 feet for boats and rafts of up to 15 tons capacity, larger craft having to shoot the falls as before.

Across the river on the eastern shore, construction of the Pennsylvania State Canal System, beginning in 1826, was stimulated by the enormous success of New York's Erie Canal. This new canal allowed navigation along the Susquehanna River above Columbia by canal boats of 75 ton capacity by 1832. The Eastern

Division, running from Columbia on the Lancaster County shore to well above Harrisburg, connected with the state system's Juniata Division extending up the river to Hollidaysburg, and via the Allegheny Portage railroad and the Western Division ultimately to Pittsburgh. It also gave access, via the Susquehanna and North Branch Divisions of the state system, to the anthracite coal fields of the Wyoming Valley. The Eastern Division, which rendered the short west bank Conewago Canal obsolete, utilized 8 lift locks to rise 55 feet in the 43 miles between Columbia and the mouth of the Juniata. The history of the extensive Pennsylvania canal system, its branches, and related internal improvements has been described by Klein (1901), Shand (1965), and McCullough and Leuba (1973).

South of the terminus of the Eastern Division, below Columbia, the Susquehanna drops 233 feet in 45 miles, a gradient nearly five times as great at that between Columbia and the Juniata. It was realized that the large volume of goods moved by the Pennsylvania Canal, part of which were hauled overland to Philadelphia by the Philadelphia and Columbia Railroad (built by the state as an integral part of its canal system), could be more cheaply transported to tidewater via the Susquehanna if similar improvements could be made to the river below Columbia. Such efforts traditionally had been vigorously opposed by Philadelphia merchants fearing a consequent loss of business to rival Baltimore (Livingood, 1947). This opposition changed to support, even before completion of the Pennsylvania Canal, when Philadelphia itself gained access to the Chesapeake with completion, in 1829, of the 14-mile-long Chesapeake and Delaware Canal cutting across the neck of Coastal Plain between the two bays.

Two private companies were organized, and chartered, by their respective states to build a canal southward along the lower Susquehanna below Columbia: the Susquehanna Canal Company in Pennsylvania and the Tidewater Canal Company in Maryland. The entire stock of the Maryland company, however, was held by the Pennsylvania corporation and the resulting canal was known and operated as the Susquehanna and Tidewater Canal. The canal was built on the west bank of the river with locks numbered sequentially downstream starting in Pennsylvania (numbers 1 to 19) and again, as a new sequence (locks number 1 to 9) starting at the Maryland line, for a total of 28 lift locks in all. Construction began in 1837 and was completed in 1839 with much fanfare. Nicholas Biddle of Philadelphia was one of the speakers at the dedicatory celebration at Havre de Grace. A major flood almost immediately made the canal inoperative, however, and regular use did not begin until 1840.

Originally the new canal was to have followed the east bank of the river so that its Maryland terminus would have been at Perryville instead of Havre de Grace. This plan was conceived in order to take advantage of the already constructed eight-mile-long Susquehanna Canal, completed in 1803, that bypassed the lower rapids between Port Deposit and Love Island near the Pennsylvania line, above the site of the present Conowingo Dam. The

old east bank Susquehanna Canal accomplished this by means of 8 locks with a total lift of 59 feet. Being located entirely in Maryland, this canal did not need the approval of the Philadelphia dominated Pennsylvania legislature, but it was not an economic success, because of both the small size of its locks and the continued obstruction to navigation farther upstream in Pennsylvania. When approached about its sale, however, the owners of the Susquehanna Canal demanded a price too high for the new company, with the result that the Susquehanna and Tidewater Canal was routed along the west bank of the river instead. In 1840, after the new and much larger canal was opened to navigation, the Susquehanna Canal could be bought out at a lower price and it was thereafter essentially abandoned, although it remained intact until construction of the Columbia and Port Deposit Railroad in 1866.

The locks of the first improvements to the lower Susquehanna, construction of which started in the late 18th century, were of limited size; the Conewago Canal could handle from 9 to 15 tons and the Susquehanna Canal only up to 10. In contrast, the locks of the new canal, by handling two boats at a time, had up to 300 tons of capacity. Each Susquehanna and Tidewater lock was 170 by 17 feet in dimension, divided by an intermediate gate for handling either single boats of 135 tons or simultaneously two boats totalling up to 300 tons. Canal boats were typically 16 feet in width, making a tight fit within the locks. From Wrightsville (opposite Columbia) to Peach Bottom the locks were "composite locks," the chambers consisting of rough stone work lined with wooden planking. Vertical grooves are visible in the stone work of the composite locks, marking the location of the timbers to which the horizontal planking was attached. From the last Pennsylvania lock (#19) through Maryland the interior of the chambers were constructed of a smooth facing stone. The entrance portions of all locks were faced with smooth stone. The cost of 45 miles of canal, 50 feet wide and 6 feet deep with 28 large locks and other necessary structures, was $3,500,000.

During its heyday, between its opening in 1840 and the time when, following the Civil War, railroads had taken over all of the passenger and most of the freight traffic except for heavy bulk commodities such as lumber and coal, the Susquehanna and Tidewater transported large quantities of goods and many people. At its northern terminus in Wrightsville passage could be booked, not only to Baltimore and Philadelphia, but even all the way to Great Britain via those ports. The peak year for tolls collected ($211,141) before the Civil War was 1855 when nearly 8,000 boats passed through the canal (Livingood, 1947). Eventually, with the railroads carrying more and more freight, especially merchandise bound for the interior, downstream shipment of lumber, coal, and iron came to dominate canal traffic. Even in the peak pre-Civil War year of 1855, three-quarters of the toll revenue was generated by downstream traffic. Maintenance costs, because of floods and the lower Susquehanna's notorious ice jams, were always high and the canal's economic viability declined. In 1872 the canal, by then largely a coal carrier, was leased to the Philadelphia

and Reading Railroad Company which had constructed a branch line from the eastern anthracite fields to Columbia to feed Baltimore with coal via the canal. The canal was badly damaged by flooding in 1889, the same event which produced the Johnstown disaster, and following additional major flooding in 1894 it was abandoned by the railroad. Partial submergence by construction of the power dams was to follow.

SURVIVING CANAL REMAINS

Of the 45 miles of canal, only about 24 miles (53%) remain above present river level, with 13 of the 28 lift locks being now submerged. Figure (VI-1) is a profile of the canal showing the location of the locks and present lake impoundments.

Boats traveling downstream from above Columbia entered the canal at Wrightsville after having been towed across river from the canal basin at the southern terminus of the Eastern Division of the Pennsylvania Canal at Columbia. Between Columbia and Wrightsville a dam created a slackwater pool on the Susquehanna for that purpose, the towing mules walking a unique double-decked towpath built on the downstream side of the mile-long Columbia-Wrightsville covered bridge. The first eight miles of canal south from the entrance to Fishing Creek involve a drop of only ten feet. The flatness of this "Long Level" north of Lock 2 is the result of an area of outcropping Conestoga Marble on the west shore of the river. The remains of the Long Level stretch and of Lock 1 (the Wrightsville guard lock) and Lock 2 at the south end of Long Level, are above the waters of Lake Clarke which is impounded behind the Safe Harbor Dam.

The next series of locks, numbers 3 through 6, provided a total lift of 32.5 feet in the stretch of river north of the Safe Harbor Dam and now submerged. Near former Lock 6 (Lockport) an outlet lock provided access to the Susquehanna, where the river was impounded by another low dam, so that boats could cross over to the Conestoga Navigation Company waterway that followed the Conestoga River for 18 miles into Lancaster County.

The sites of Lock 7 (Shenk's Ferry) and Lock 8 (York Furnace Weigh Lock) are present between the Safe Harbor Dam and the head of Lake Aldred which is impounded above the Holtwood Dam. At Lock 8 boats were weighed, to determine toll charges, on a submerged oak balance scale 75 feet in length. Unlike other locks, the weigh lock was covered and enclosed because of its special use. Downstream, Locks 9 and 10 are now beneath Lake Aldred and Lock 11, located at the west abutment of the Holtwood Dam, was destroyed during the dam's construction.

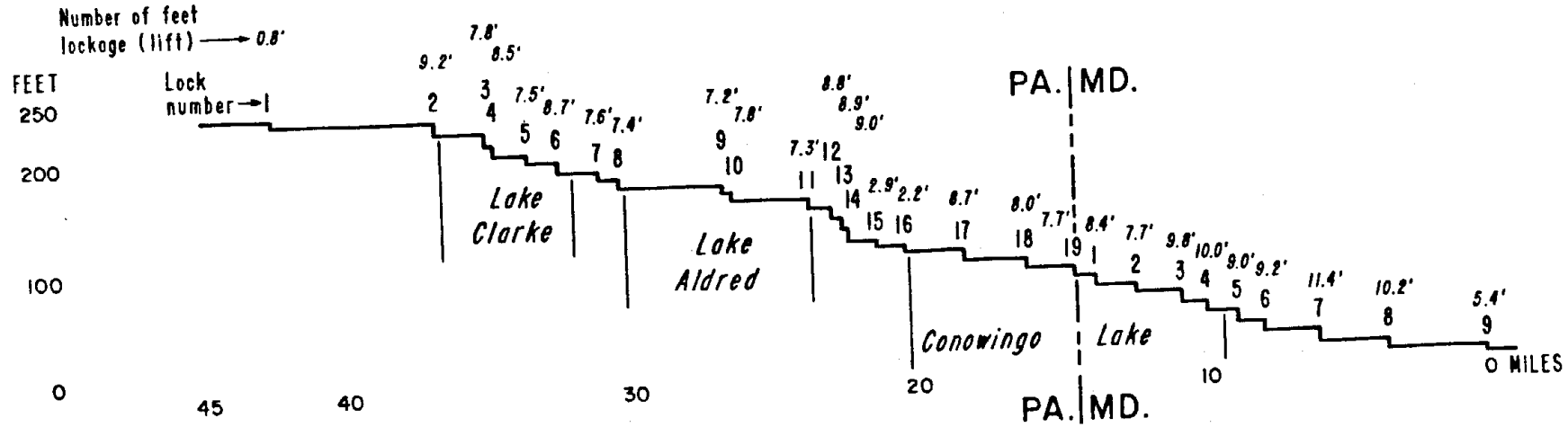

Figure VI-1. Longitudinal profile of the Susquehanna and Tidewater Canal.

From the Holtwood Dam southward the next sequence of locks (11 through 14) provided the greatest lift in the shortest distance: 34 feet in about a mile and a half. Locks 15 and 16 had considerably smaller lifts (about 5 feet total); these are located near the head of Lake Conowingo. Lock 17 is mostly inundated while numbers 18 and 19 (Pennsylvania) and 1 through 4 (Maryland) are now below water level. From Conowingo Dam south to the outlet lock to the Chesapeake (Lock 9) at Havre de Grace, the towpath (followed in large part by the Mason-Dixon Trail) and Locks 5 through 9 are generally in a good state of preservation. These last five locks provided a total lift of 46 feet. Lock 9, the Tidewater Lock with its restored lock tender's house adjacent, is preserved in the City of Havre de Grace's North Park as the Susquehanna Museum of Havre de Grace, Inc.

At the Lock 12 historic area, where we have lunch, Lock 12 has been stabilized and partially restored by the Pennsylvania Power and Light Company. South of it, beyond the highway and along the blue blazed Mason-Dixon Trail that goes under the Norman Wood Bridge, the abutments of a sixty-four foot long covered bridge that crossed the canal are found, as are the remains of Lock 13. Also visible between the covered bridge abutments and Lock 13 are foundation stones of a store and tavern that served the needs of canalers. A wall supporting the towpath rises above low ground along the river adjacent to Lock 13 which itself is in a good (but unrestored) state of preservation. Farther downstream is the site of McCall's Hotel and Lock 14 which has been mostly destroyed by ice jams and river floods. Just to the south of Anderson Run, which passes through the Lock 12 historic area, are restored lime kilns and the remains of a saw mill. Other ruins can be found farther north in the vicinity of the Holtwood Dam.

BRIDGES AND FLOODS

McCall's Ferry at the site of the present Holtwood Dam is particularly noted as the location, between 1815 and 1818, of the longest single-span wooden arch covered bridge in the world. The site, just upstream from the dam and now submerged, is where the river is particularly narrow. This afforded an opportunity for spanning the Susquehanna with a "permanent" bridge. At low water the original channel was 348 feet wide with a swift current running to a reported depth of over 100 feet (Shank, 1980).

Theodore Burr (1771 - 1822), perhaps the most famous of Pennsylvania's covered bridge builders, was selected by the McCalls Bridge Company as contractor. Burr spanned the gap with a 360 foot wooden arch that extended from the Lancaster shore to a pier near the York County side, using an additional 100-foot span from the pier to the western shore. The large arch was constructed in two sections on floats along the river bank. Because of river conditions it took two weeks, the assistance of hundreds of local farmers, and the aid of an early ice jam to move the

sections into place. Unfortunately, three years later another and unprecedented ice jam removed the entire structure and it was never rebuilt. Burr, who had been paid in company stock, therefore lost all compensation for his efforts, except for the reputation gained by the building his masterpiece.

Floods and ice jams were, and still are, common on the lower Susquehanna River. According to data assembled by Buchart-Horn Consulting Engineers (Shank, 1972), the Susquehanna experienced major floods of more than eight feet above bank-full stage in 1784, 1865, 1889, 1894, 1902, 1936, and 1972. The 1889 flood was general in Pennsylvania and is most commonly associated with the destruction of Johnstown. But it, and the 1894 flood, were also responsible for the final closing of the Susquehanna and Tidewater Canal. Ice jams, or "gorges," such as the extraordinary one that destroyed the McCall's Ferry Bridge can, in some intervals, be almost an annual occurrence. The 1936 flood was accompanied by a major ice jam that temporarily endangered the Safe Harbor Dam and washed out a deflection wall and four transformers at Holtwood, resulting in the shutdown of electric generation there. The 1972 "Agnes" flood, even though 3.5 feet above previous gauge heights, occurred at the beginning of summer and, without the effects of ice, caused less damage. On the lower Susquehanna the major damage was the washing out of the Shocks Mill railroad bridge above Columbia.

In contrast to the occasional destruction by ice jams and floods is the normal peace and tranquility of the original river. This was well described in an account of a river trip published in 1888 by Jacob Gossler, a resident of Columbia:

> If any of my readers desire a novel experience, an exhilarating ride, and a delightful excursion, let me suggest that ... some time in 'the pleasant month of June' during the 'June Fresh' ... charter a raft for 'Port' [Port Deposit, Md., at the head of Chesapeake Bay]. You may get one with a cabin, to which you can retreat in case of rain, or repair when you are hungry. At first you will float lazily along the broad, placid river, until you strike the 'chute' in the [Columbia -Wrightsville] dam, through which you rush with race-horse speed; then, subsiding to the natural current, you pass Little Washington (a nearly extinct town); and among the hundreds of islands that dot the again broadening river, noticing, as you glide by, the fisherman in their light, pointed canoes, rapidly propelled by a long iron-shod pole. Then through the cliffs, five hundred feet high, at Turkey Hill; then a rest in the shallows of the again wide and rocky stream, until, at McCalls Ferry, where you can throw a stone across the river, and where the water, two hundred feet deep, seems to stand on edge, and careers wickedly through the silent and sullen but swift current, the elastic raft throwing high the spray and bending and swaying like a veritable sea-serpent. Soon after, you glide quietly into 'Port,' whose glory has departed since lumber and lumbermen have become scarce (Gossler, 1886).

These sights would have been visible on the river and from the old Susquehanna and Tidewater Canal itself. While we are at Lock 12 hark back and, in imagination, enjoy the tranquility of yesteryear.

VII. PALEOZOIC CARBONATES OF THE CONTINENTAL MARGIN IN THE LANCASTER - YORK VALLEY, PENNSYLVANIA

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Pennsylvania Geological Survey

INTRODUCTION

Parts of this essay represent remnants of an introduction initially prepared for STOP 7. The intent was to set that stop in context with the other stops of this field conference which have significant relationships among them other than mere geographic proximity. The fact the leaders of this trip represent an unusually large number of organizations doing things for different reasons adds to its interest, but it appeared to be presenting the hazard that the forest might be lost for the trees, particularly for the many we hope will attend this conference who have had no occasion to concern themselves with the peculiarities of the Piedmont. As it developed, it became apparent that the introduction had grown out of proportion to a stop description, but contained elements that might be helpful to give a general perspective, especially to the stops of the second day. Hasty last minute revision to accommodate this broader objective may result in some unevenness of treatment. Of the various authors, I have worried about the Piedmont longest and looked at it least. I bring to it some firm opinions about what must have happened here, but I recognize that some must yet be bent to accommodate ground truth. I do not anticipate that all the complexities of the siliciclastic metamorphics of the Inner Piedmont will be resolved in my lifetime, but at least that task is fairly begun. I am more optimistic about the carbonates of the Outer Piedmont. Some of those opinions are set forth here. I do not ask you to simply believe them, but I do like a good argument. So have fun!

HISTORICAL CONTEXT

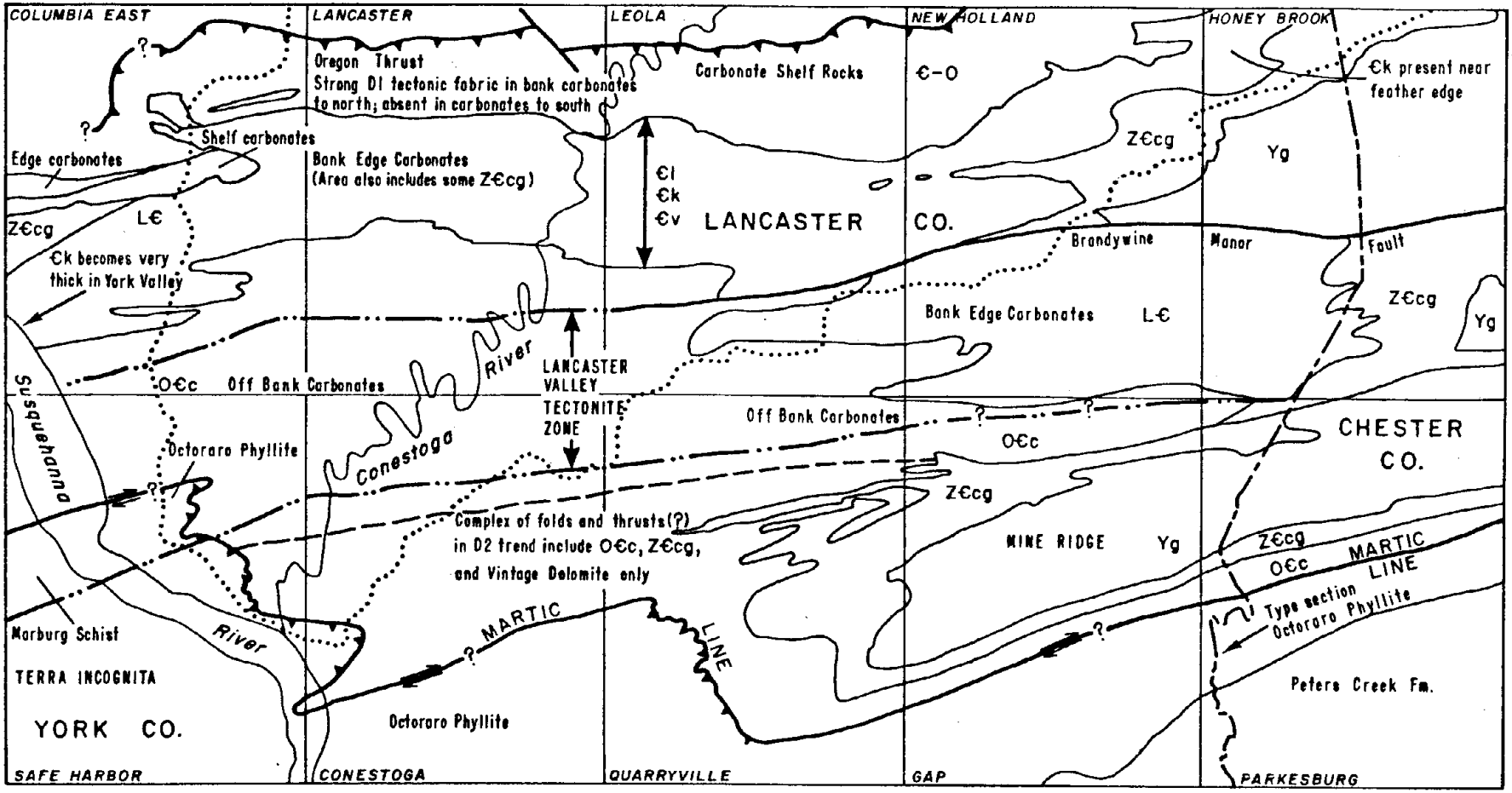
Near the beginning of long and fruitful collaboration, George Stose and Anna Jonas (1922) published on the lower Paleozoic section of Southeastern Pennsylvania. A major intent of that paper was to divide the thick carbonate section into mappable units. Consistent with the prevailing stratigraphic paradigm, they imported a number of unit names, largely from south central Pennsylvania and northern Maryland, where such units seemed to fit the general succession of strata. This usage survives only where no one has subsequently examined the rocks sufficiently to decide what else to call them. Where no such fit was possible they provided local names: in the carbonate section, the Vintage Dolomite, Kinzers Formation, Ledger Dolomite, and Conestoga Limestone. They subsequently mapped these units throughout almost their entire extent in a manner quite accurate enough for most purposes. With the exception of Conestoga (Stose and Jonas, 1923) they did not really attempt to explain this unique assemblage. In the last particular they were wide of the mark, but they are hardly to be faulted. Until enough became

known about ocean basins and the processes governing their dynamics such that meaningful and consistent interpretations of modern continental margins could be made, anything that could be said about ancient margins (if recognized) is largely irrelevant today. The term carbonate bank does not much appear in geologic literature before the 1950s, but it became recognized that this is the essential nature of most of the thick Lower Paleozoic carbonate section of the "miogeosyncline". John Rodgers's numerous significant contributions to Appalachian geology largely do not arise from direct observations on Pennsylvania rocks. He was, however, the man with the right experience at the right time to recognize a bank edge when he saw one. His observations on Pennsylvania (Rodgers, 1968) were almost in passing in a larger regional synthesis, but some enigmas of our Piedmont were resolved at a stroke. If one regards the Vintage, Kinzers, and Ledger as representing the inception and Lower to Lower Middle Cambrian evolution of a carbonate bank edge and the Conestoga as the more proximal off-bank facies, then much of what can be said about these formations at the various stops of the second day should be recognized as painting in the details of Rodgers's insight. The bank edge certainly persisted in some form longer than is represented by these rocks, except possibly the Conestoga; but that form shall ever remain conjectural, as rocks at that locus have been entirely removed by several episodes of subsequent erosion ranging from Upper Middle(?) Paleozoic to Holocene.

I do not mean to imply that many earlier workers in the Piedmont, including the Stoses, did not recognize that they were treating a complex of shallow and deep marine rocks, nor that they failed to grasp some of the implications. Rather, they lacked the conceptual apparatus to provide an explanation that was ever very satisfactory. When a passive margin (the bank is an unessential embellishment, but appropriate to the probable Lower Paleozoic paleolatitude of the Iapetan margin of Laurentia) implies the probable future existence of an active margin, we may make some significant observations about the tectonic features to be observed at the various stops, although the full story awaits the resolution of many problems, especially in the Inner Piedmont. (By way of illustration, the preceding statement was deliberately phrased to be gibberish to Stose. It should be fully intelligible to anyone who has taken a proper degree in geology in the last decade or so, although they might justly protest its grammatical complexity.) So there Mr. Stose! It is your punishment for abuse of the "Wissahickon."

SOME NAMES: NEW, FAMILIAR, FORGETABLE, AND ALMOST FORGOTTEN

The name "York Valley" has long been used in a geographically and geologically consistent sense to identify the relatively narrow carbonate valley bounded on the southeast by the Chillhowee Group and on the northwest by (somewhat) the same rocks or the Newark-Gettysburg basin. It contains the city of York and extends somewhat beyond the boundaries of that county. The name "Lancaster Valley" is newly used (in this guidebook and



EXPLANATION

..... Drainage divide, Conestoga River drainage

·· —·· —·· Boundary of tectonite zone other than Brandywine Manor fault

€-O Carbonate shelf rocks, undifferentiated

L€ Bank edge rocks
- €l Ledger Dolomite; reef facies
- €k Kinzers Formation; complex variation in lithology and thickness; restricted to vicinity of €l
- €v Vintage Dolomite; pre-bank blanket dolomite

O€c Conestoga Formation; off bank carbonates

Z€cg Chilhowee Group, undifferentiated; intra/post-lapetan rift clastics

Octoraro Phyllite—"Xwc" of State Geologic Map (Berg and others, 1980); most shoreward of deep marine pelites; probably includes Conestoga equivalents

Peters Creek Formation—Marine graywacke-pelite sequence; age of Octoraro and Peters Creek not properly established

Yg Middle Proterozoic gneisses of "Grenville" metamorphic age

Figure VII-1. Sketch map of the southern Lancaster Valley area.

by Valentino and MacLachlan, 1990) to identify the broad, predominantly carbonate-floored valley bounded on the north by the Newark-Gettysburg basin, on the east by the Honey Brook upland, on the south by the Chillhowee Group of Mine Ridge or the Octoraro Phyllite. The city of Lancaster is centrally located in this valley, which is almost entirely confined to Lancaster County and occupies a substantial portion of it. It is offered as a substitute for one of the less defensible usages of "Conestoga Valley." The Conestoga River neither drains all of the Lancaster Valley nor is that drainage confined to it. Indeed, the familiar Lebanon Valley located north of the Newark-Gettysburg basin is best defined by criteria related to the description of the Lancaster Valley offered here, and it also includes Conestoga drainage.

The Lancaster Valley Tectonite Zone is discussed by Valentino in this guidebook (Chapter III). Fairly precise limits have only recently been established in parts of central and eastern Lancaster County and are included on the sketch map of the southern Lancaster Valley area (Figure VII-1). The near coincidence of these boundaries with the chlorite grade retrograde metamorphic zone previously defined by Valentino and Faill (1989) from scattered thin-section studies would be remarkable if there were not substantial reason to believe they are, in fact, truly identical. Some structures most characteristic of the Lancaster Valley Tectonite Zone may be found at STOP 9 in the Conestoga Formation in the York Valley, with some notable differences. The "S1" of Lancaster County is absent and "D2" does not display the metamorphic imprint. These are some aspects of structural complexities addressed in part by Faill and MacLachlan (1989), but not fully resolved. The Lancaster-York Valley is a marriage of convenience to unite the major exposures of the carbonate bank edge. Owing to the afore-mentioned structural perplexities and other regional considerations, a divorce of this union may well be indicated in other contexts.

"Wissahickon" has been used elsewhere in this guidebook to identify metapelites south of the Lancaster Valley. Its place under this heading reflects its eminently forgetable status in this context, which was perpetrated by Stose to shove under one umbrella miscellaneous, poorly understood rocks of possibly the same general character. The convenience for mapping purposes under these circumstances is obvious, but it would have been better if he had not co-opted an otherwise usable name. Excusable, perhaps, but his failure to reflect that one *name* does not necessarily mean one *thing* has somewhat obfuscated subsequent Piedmont studies. The Wissahickon Formation (mica gneiss of Bascom, 1902) apparently remains a valid, possibly subdivisible unit in its type area east of the Rosemont Fault (Philadelphia and Delaware Counties).

The Octoraro Phyllite (schist of Bascom, 1902) was identified as a distinctive unit with a type section in the southeast corner of the south Lancaster Valley map. It has been revived by the USGS (Lyttle and Epstein, 1987) with the new descriptor but the same content. I have persuaded at least one colleague that,

in its refurbished dignity, it is preferable usage in this area. Provisionally, it has not been extended across the Susquehanna River for a variant of the same reason that Stose and Jonas (1939b) equivocated. The contact of the Marburg Schist and the "Wissahickon" appears to be a direct continuation of the south border of the Lancaster Valley Tectonite Zone, and some Marburg may be merely a retrograde aspect of the "Wissahickon." The Marburg, however, includes variants suggestive of some components of the Hamburg Klippe which have not been identified elsewhere in the Piedmont. Pending better integration of units in the Maryland western Piedmont and York County, this area should remain as on the map. In the remainder of this chapter, "Octoraro Phyllite" is used to refer to rocks traditionally called "Wissahickon."

A LITTLE STRATIGRAPHY

The Vintage Dolomite, Kinzers Formation and Ledger Dolomite are reasonably established with partial type sections in eastern Lancaster County. Several members of at least local significance will be demonstrated; but only the Kinzers, which varies considerably from the type in both thickness and composition, appears a possible candidate for more radical dissection. Stose and Jonas (1922) identified all three as Lower Cambrian equivalents of the Tomstown Dolomite, based on position and some Kinzers faunules. Additional finds, notably Campbell (1971) and Taylor (Chapter IX, this guidebook) somewhat extend the range, and Taylor will undertake to demonstrate (STOPS 6 and 10) that Kinzers and the reefy Ledger prograde eastward (??). The direction is surprising, but the fact is not, given our model.

The Conestoga Limestone is loosely defined by contemporary standards, but is easily mapped within the area of the regional sketch map. Structure and stratigraphic constraints limit the possible age as no older than Lower Cambrian and no younger than Chazyan. Possible Conestoga fossils are reported from the Susquehanna River region (near the Lower/Middle Cambrian boundary) and Schuylkill River region (Lower Ordovician "Beekmantown"), but the significance of these reports is enigmatic for various reasons. A suite of 10 large samples across the strike dissolved in acetic acid proved barren of conodonts and other determinable organic material (Anita Harris, personal communication, samples from MacLachlan and Root).

The USGS Stratigraphic Lexicon implies that Stose and Jonas (1923) is the definitive description to establish the formation; this paper is the best reference for the type, although the name was first published by the same authors a year before. The type area is the Conestoga Valley or, in some contexts, the valley of the Conestoga River. The latter, used with the most restricted definition possible applied to the reach from the southern side of the city of Lancaster south to the Octoraro Phyllite, provides the best reference as no type section is defined. This section provides maximum outcrop across the width of the crop belt, but exposure is discontinuous, the structure is complex, marker beds

are absent, and the thickness is indeterminable by stratigraphic measurements. All reported thicknesses are based on structural assumptions or models.

A LITTLE STRUCTURE

Comparative petrofabric studies by Valentino show the same metamorphic history and deformational fabrics in the Octoraro Phyllite on Turkey Hill (general area of STOP 1) and the Conestoga Formation immediately to the east with one significant exception: S1 of the Octoraro is a clearly transposed foliation while it is bed-parallel in the Conestoga. This area lies within the Lancaster Valley Tectonite Zone, and the Conestoga of the STOP 7 area is quite similar, though the small-scale plications are even more intense to the south while the larger scale D2 folding is indistinct there. The most attractive, and possibly only, interpretation of the fabric difference noted is that the Octoraro S1 developed prior to or during thrusting, apparently synchronous with the prograde metamorphism, which emplaced it over the Conestoga. This is, of course, the classic Martic or Taconic(!) Thrust.

That the Conestoga-Octoraro contact is broadly folded in D2 is obvious near STOP 1 and several other relatively short segments that are transverse to the regional trend of the Martic Line. The latter, including the much disputed exposure at Martic Forge, is clearly aligned with the strike of D2 and may be much impacted by D2 dextral shear, possibly of large magnitude, as is apparent in the fabrics of rocks in the Lancaster Valley Tectonite Zone. There are two phases of Taconic thrusting that may be closely dated stratigraphically in the Great Valley north of this area which well may represent distal movement on the Martic Thrust. Emplacement of the Hamburg Klippe(n) is post *Nemograptus gracillis* zone, between the upper and lower parts of Berry's graptolite zone XII of Upper Trentonian age at about 458 Ma. This is essentially identical with thrusting in the Taconic type area. The Klippe rocks of Pennsylvania are involved in subsequent overturned nappes and associated thrusting of the platform carbonates which have a Richmondian(?) neautochthonous cover and are in any case older than basal Silurian at about 435-440 Ma. The latter age best approximates the prograde metamorphic climax from radiometric determinations in Lancaster County (and in much of the Piedmont area). Continuous or episodic movement on the Martic Thrust during the whole interval is plausible.

Retrograde metamorphism associated with D2 is no younger than 330 Ma (Upper Mississippian) and possibly 360 ma (Uppermost Devonian) cooling ages. The D2 event differs in age and style from classic Alleghanian Orogeny; the latter is difficult to distinguish in much of the Piedmont area. Late Paleozoic (Alleghanian *sensu stricto*, about 270 Ma) deformation, however, is inferred to be the origin of the Oregon Thrust, which appears near the north margin of the Lancaster Valley map, among others and may produce the post D2 features of STOP 7. Current tectonic opinion is that it has the same fundamental cause, convergence of

the African (Gondwana) Plate; and it follows that D2 is best identified as an early phase of the Alleghanian (the superseded Appalachian Orogeny is perhaps a useful umbrella name in this case) rather than an extension of the Acadian, which is as close or closer in age (about 395 ma), but related to different elements of the Pangean assembly in its type area.

VIII. THE WEST YORK BLOCK: STRATIGRAPHIC AND STRUCTURAL SETTING

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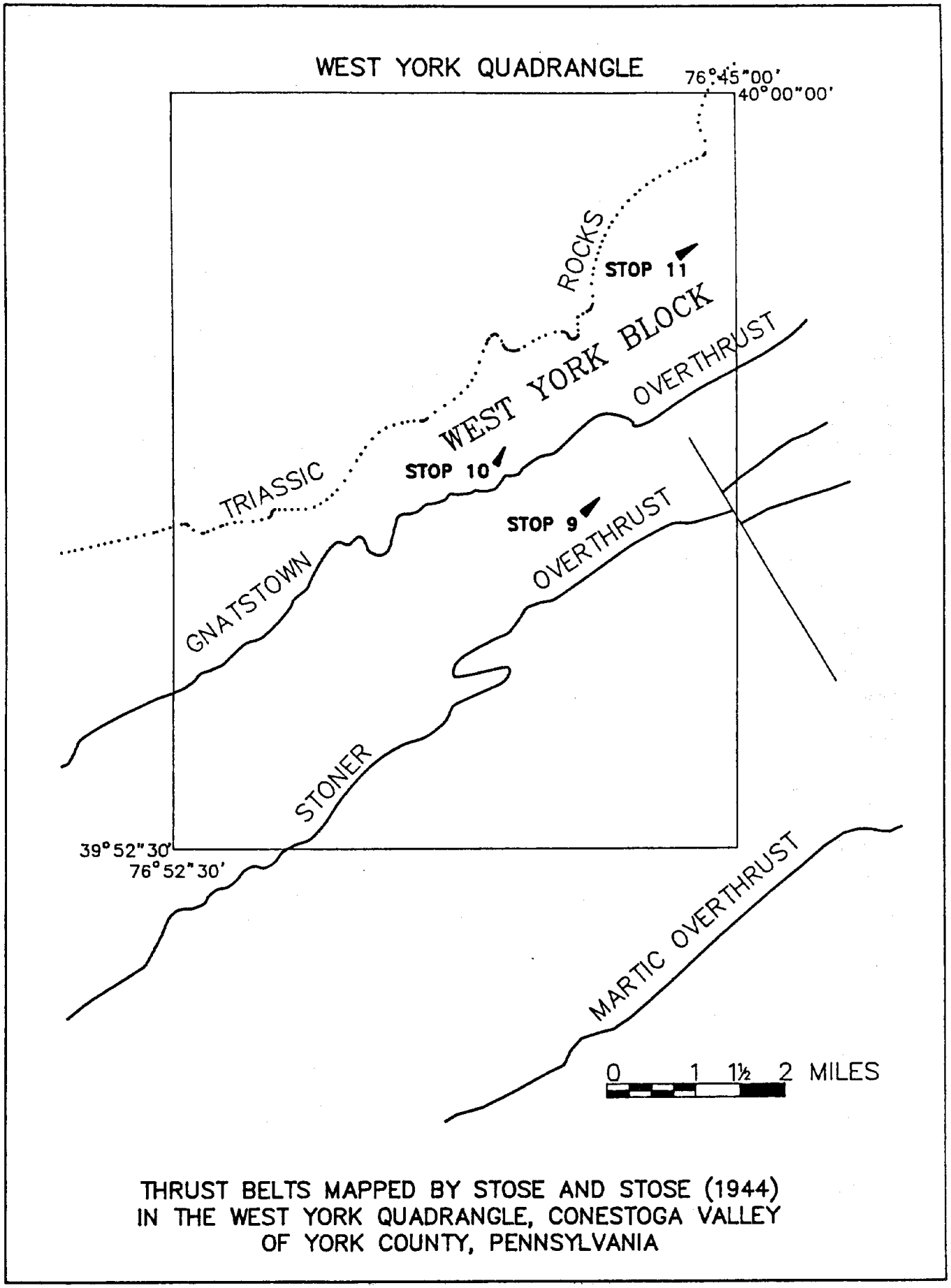
INTRODUCTION

The Conestoga Valley in York County is divided into discrete structural blocks by a series of subparallel northeast-trending thrust faults first mapped by Stose and Jonas (1933). The fault blocks bounded by these thrusts differ with regard to degree of deformation and stratigraphic content. The West York Block is the most forward (northwestward) of these fault blocks. It is this block, exposed in the West York Quadrangle (Figure VIII-1) situated north of the Gnatstown Fault and south of the Triassic overlap/fault boundary, that is the focus of this chapter.

The Conestoga Valley lies in a transitional position between the resistant crystalline rocks of the Piedmont uplands to the southeast and the Great Valley section of the Valley and Ridge Province to the northwest (Figure VIII-2). A few small areas of Precambrian exposure occur in the valley, but Lower Cambrian to Lower Ordovician strata dominate the bedrock geology. A diagram illustrating the various formations recognized in previous studies of the Conestoga Valley was constructed by Gohn (1976) and is here reproduced as Figure VIII-3. Stose and Jonas (1939) treated these units as time-stratigraphic packages with the Conestoga Formation lying unconformably atop the Ledger Dolomite. Gohn (1976) considered the base of the Conestoga to be an unconformity but recognized, as did Rodgers (1968) and Campbell (1969), that the Conestoga is a time-transgressive basinal facies. We consider most, perhaps all, of these formations to be time-transgressive lithostratigraphic units. Middle Cambrian and younger rocks may be much more extensively represented in the eastern portion of the valley (Lancaster and Chester Counties) than they are in York County. Evidence of this age difference is provided by new biostratigraphic discoveries within the West York Block (Taylor and Durika, Chapter IX, this guidebook). The West York Block represents only a small portion of the Conestoga Valley but is one of the least deformed blocks. It provides many valuable exposures of highly fossiliferous and only mildly deformed strata that afford an opportunity to reconstruct parts of the stratigraphic succession with a high level of confidence and with good biostratigraphic control.

The West York Block and other parts of the valley in York County are extensively mined. The resultant quarry and underground mine openings provide an unusual opportunity to map and study this limited area in considerably more detail than is normally possible in areas of non-resistant bedrock. These convenient exposures permit an examination of some previously held



THRUST BELTS MAPPED BY STOSE AND STOSE (1944)
IN THE WEST YORK QUADRANGLE, CONESTOGA VALLEY
OF YORK COUNTY, PENNSYLVANIA

Figure VIII-1.

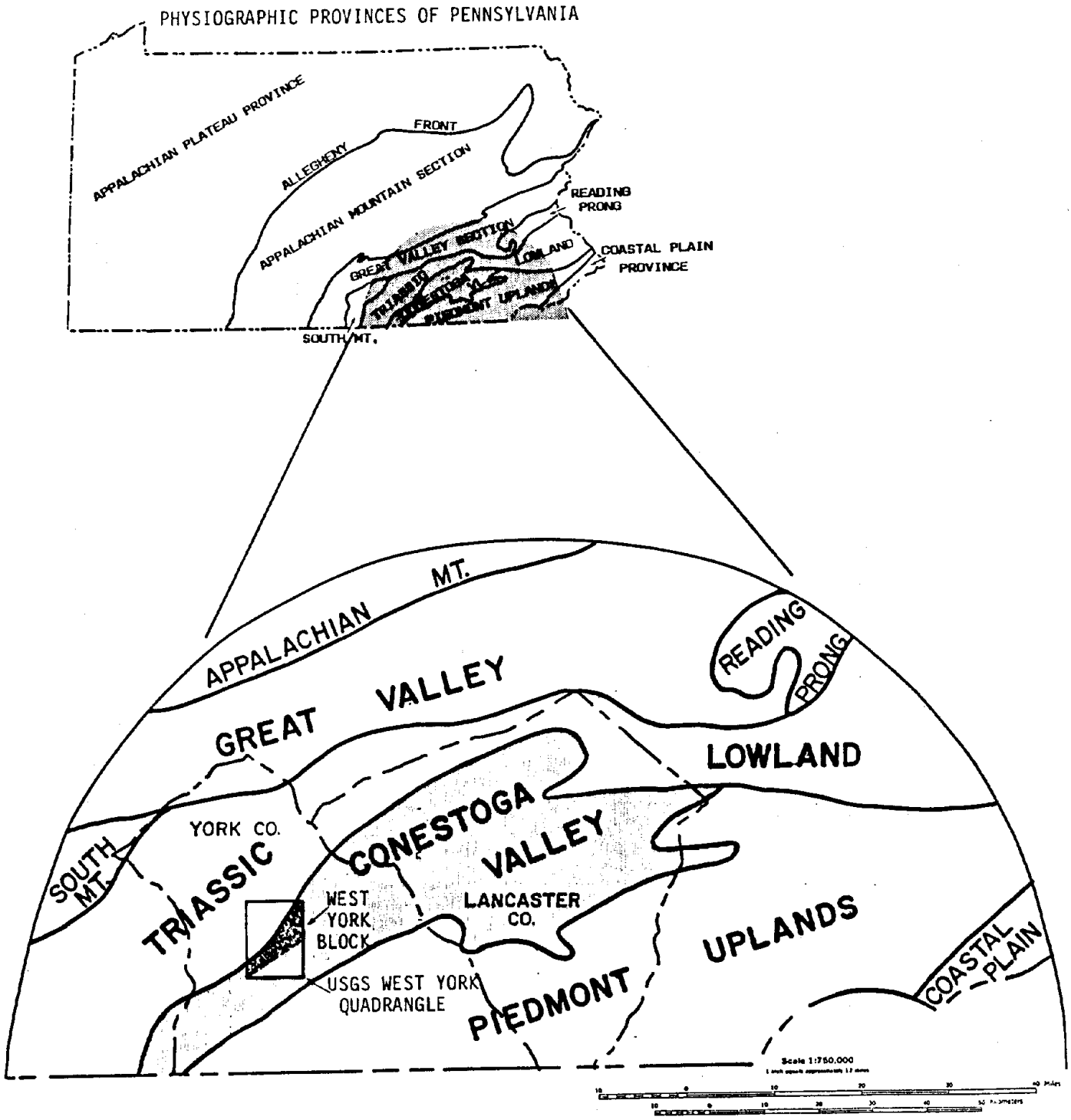
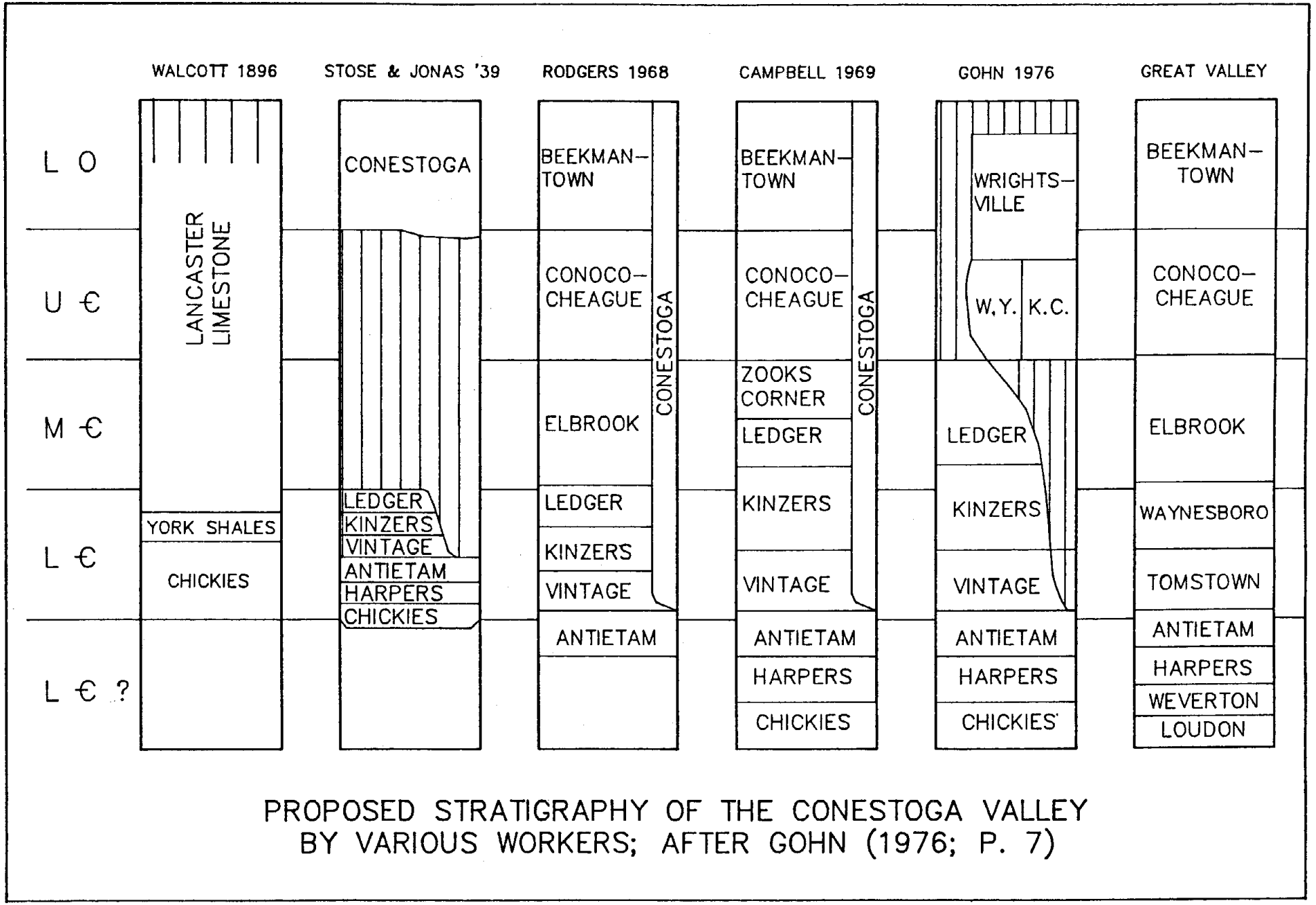


Figure VIII-2. Regional setting of the West York Block.



PROPOSED STRATIGRAPHY OF THE CONESTOGA VALLEY BY VARIOUS WORKERS; AFTER GOHN (1976; P. 7)

Figure VIII-3.

concepts regarding structural, stratigraphic, and age relationships.

STRATIGRAPHY

The stratigraphic sequence in the Conestoga Valley in York County has been, from its earliest descriptions, broadly subdivided into a lower portion dominated by clastic units comprising (in ascending order) the Chickies, Harpers and Antietam Formations above which lie dominantly carbonate units, specifically the Vintage, Kinzers, Ledger, and Conestoga Formations. A small area of Antietam and Vintage crop out in the western edge of the West York Block but, aside from that, the surface is underlain by the Kinzers, Ledger and Conestoga Formations. It is these three formations that we will be discussing in more detail.

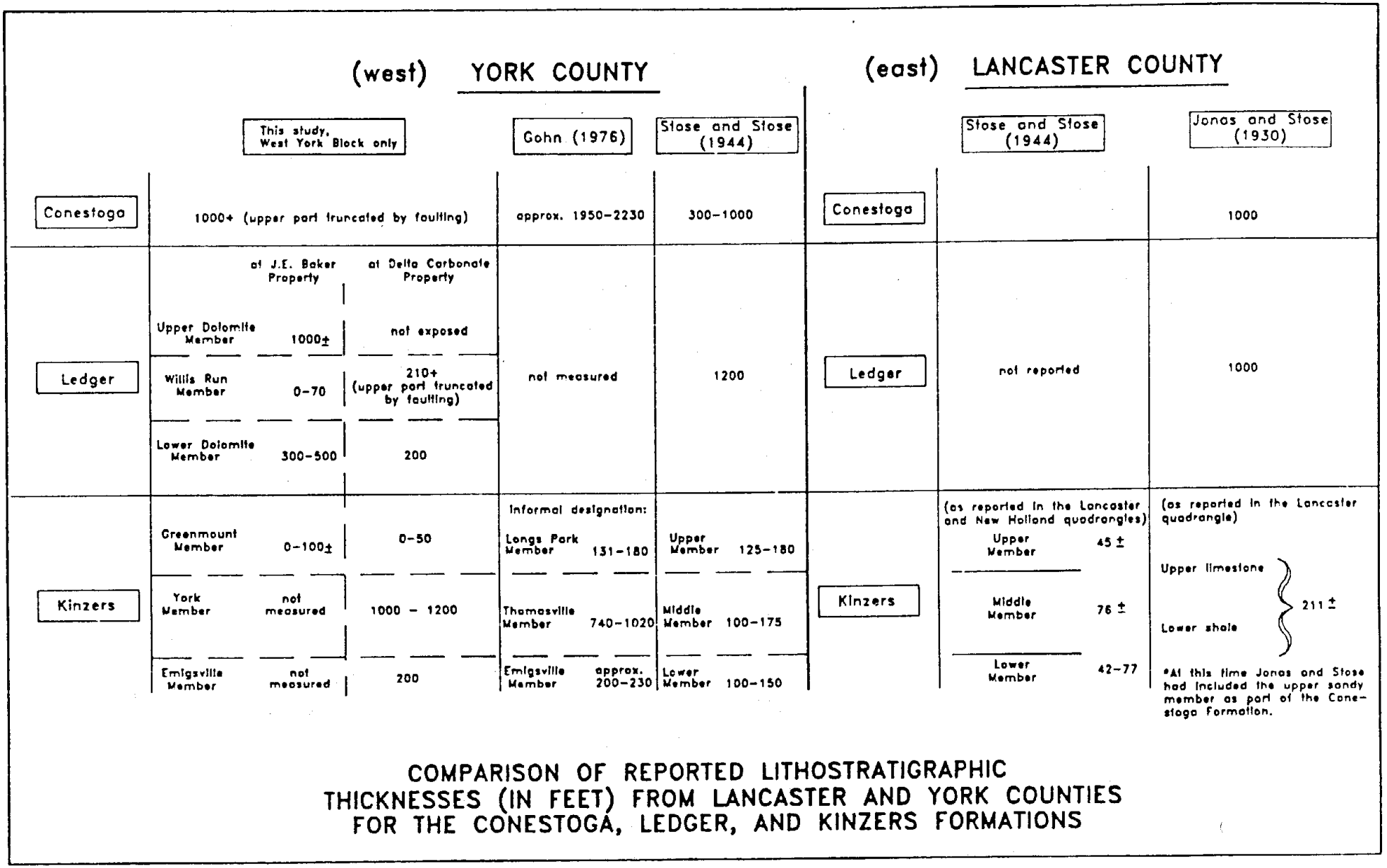
A comparison of reported lithostratigraphic thicknesses from Lancaster and York Counties for the Kinzers, Ledger and Conestoga Formations is provided in Figure VIII-4. A three-fold subdivision of the Kinzers Formation, similar to that established by Stose and Stose (1944) and Gohn (1976), is utilized for the West York Block. Some new member designations are proposed to account for locally specific characteristics that are mappable in the West York Block. A new three-part subdivision for the Ledger Formation in the West York Block is also proposed here.

The Emgisville Member of the Kinzers Formation

The Kinzers Formation will be examined at STOP 11 at the Delta Carbonate (formerly York Stone and Supply) Quarry. The basal Emigsville Member is not exposed in the quarry, but drilling shows it to be present and lithologically similar to that described throughout York County and the rest of the valley. The Emigsville Member is the most laterally persistent unit in the West York Block and is consistently encountered where expected in drilling and mapping activities. Gohn (1976, p. 74) provided the following environmental interpretation:

> The Emigsville Member is interpreted as a continuation of the basinal setting begun during the deposition of the Vintage sediments. The major difference is the gradual change over from carbonate mud accumulation, with occasional high-energy turbidite events, to siliciclastic mud accumulation due probably to changing conditions in the carbonate sediment production area, the carbonate platform.

The thickness of the Emigsville Member in the West York Block is tentatively given as approximately 100-200 feet based on outcrop width. The member is quite fossiliferous in Lancaster County, less so in York County, and has been uniformly ascribed to the Lower Cambrian. A more detailed treatment of the reported fossil content can be found in Gohn (1976) and Ryan (1986).



COMPARISON OF REPORTED LITHOSTRATIGRAPHIC THICKNESSES (IN FEET) FROM LANCASTER AND YORK COUNTIES FOR THE CONESTOGA, LEDGER, AND KINZERS FORMATIONS

Figure VIII-4

The York Member of the Kinzers Formation

The middle member of the Kinzers Formation is well exposed on both limbs of the broad syncline in Pit 1 at the Delta Carbonate Quarry (STOP 11, Location B.) This stratigraphic interval was informally designated the Thomasville Member by Gohn (1976) from exposures at the Thomasville Quarry, which is in the West York Block. Unfortunately, the sequence exposed at Thomasville, which was first described by Cloos, 1974), is unique because of extensive megabreccia development. Gohn (1976) himself noted that, "The detailed stratigraphy established in the Thomasville Quarry also cannot be demonstrated elsewhere." The section described by Cloos (1974) for Thomasville is as follows:

Stratigraphic column of Thomasville Member at Thomasville Quarry, abridged from Cloos (1974).

Top Not Exposed	Feet	Meters
upper dolomites and limestones	350+	106.8+
"top black" limestones	20-40	6.1-12.2
upper Thomasville breccia	50-300	15.3-91.5
"bottom black" limestones	20-30	6.1-9.2
lower Thomasville breccia	up to 300	up to 91.5
phyllite (Emigsville Mbr.)	740-1020'	225.8-311.2

A more typical sequence of the middle Kinzers in the West York Block, without extensively developed breccia beds, is exposed at the Delta Carbonate Quarry (STOP 11). For this reason, we are proposing the name York Member for the middle Kinzers with the type section at the Delta Carbonate Quarry. The generalized section is described in Figure S11-2 (in Stop description for STOP 11). The continuous outcrop of this member at the Delta Carbonate Quarry is a rare exposure. Less continuous partial outcrops elsewhere can be correlated to this type section in only a general way. Specific "marker beds" have not been identified.

The thickness of the York Member at Delta Carbonate is approximately 1000-1200 feet. However, the middle (limestone) member of the Kinzers in Lancaster County, as described by Stose and Stose (1944), is estimated at 75 feet. Stose and Stose (1944) described their Middle Member (= our York Member) as only 100-175 feet thick in York County. In defense of Stose and Stose, it should be noted that none of the large quarries that show the 1000 foot thickness of this interval were available to them when they did their mapping.

Primary sedimentary features are sparse in the York Member. Where present, they comprise oolites (not discernibly cross-bedded), burrows, indistinct reefy structure (the "leopard rock" of

Stose and Stose, 1944), occasional desiccation features, bioclastic lag deposits and megaconglomerate, all distributed within interbedded very pure (in excess of 99 percent carbonate content, most as $CaCO_3$) to moderately impure (insoluble content up to the high teens) carbonates. This suggests a variable environment of shelf margin to basin slope. Gohn (1976) interprets the megaclastic accumulations seen at the Thomasville Quarry as basin slope debris flows.

Another feature consistently observed in the pure white limestones of the York Member is a coarse crystalline texture, giving rise to the term "white marble." This is very curious as practically undeformed fossils occur with this lithology at Delta Carbonate.

The Greenmount Member of the Kinzers Formation

Above the York Member is an interval of very impure carbonate where the insoluble content can reach 50 percent. We propose here the name Greenmount Member for this upper member of the Kinzers Formation in the York Valley portion of the Conestoga Valley. In weathered outcrops it may appear as a sandstone/siltstone with very little carbonate remaining. This is Stose and Stose's (1944) Upper Member and might be the physical equivalent of the Longs Park Member recognized by Gohn (1976) in Lancaster County. Gohn (1976) describes this unit in contact with the overlying Ledger Formation. The Longs Park Member is fossiliferous and has been attributed to the Middle Cambrian (Campbell, 1969). This locality will be visited as STOP 6. The same apparent lithostratigraphic sequence occurs in Pit 2 at Delta Carbonate Quarry (STOP 11); however, fossils collected there indicate an Early Cambrian age. This is evidence for the time-transgressive aspect of this litho-stratigraphic unit. (The fossil assemblage collected at Delta Carbonate in this interval is described by Taylor and Durika in Chapter IX of this guidebook).

The spatial relationship of these members is uncertain. It is not at all clear that the Longs Park Member in Lancaster County is stratigraphically equivalent to the Greenmount Member in the York Valley. Certainly the age is different. For these reasons the upper part of the Kinzers in the West York Block is herein referred to as the Greenmount Member.

Like the basal Emigsville Member, the Greenmount Member is, for the most part, laterally continuous and consistently encountered where expected throughout the West York Block. The thickness of the Greenmount varies from 180 feet reported by Gohn (1976) to about 50 feet at Delta Carbonate. In some places it is missing, apparently having been removed by slumping (mass failure). On the southwest face of Pit 1 at Delta Carbonate, the Greenmount Member was lost to such a slump failure and is represented only by discontinuous and somewhat contorted slabs within the Ledger Formation (Figure S11-4, Stop description for STOP 11). During deposition, the Ledger probably was carbonate sand

that would have flowed quite easily, while the more argillaceous Greenmount sediment behaved more cohesively. South of the slump exposure, underlying the Greenmount Cemetery, the Greenmount Member is still in place. Such slump-related features are characteristic of shelfbreak deposits.

The long history of use of the name "Kinzers" dictates the need for its continued use. The rocks included within the Kinzers Formation, however, vary dramatically from the type locality in Lancaster County to the York Valley portion of the Conestoga Valley. The unit thickens from about 200 feet to 1450 feet (Figure VIII-4). Within the thickest portion, a great deal of stratigraphic detail is discernible. There may be some logic, therefore, in assigning group status to the name "Kinzers."

The Ledger Formation

Above the Kinzers Formation is the Ledger Formation, which was not subdivided by Stose and Stose (1944) or Gohn (1976). They both describe the Ledger as a pure dolomite of about 1000 feet thickness. In the West York Block, the Ledger is divisible into three parts: the Lower Dolomite Member, the Willis Run Member (limestone/some dolomite), and the Upper Dolomite Member.

The Lower and Upper Dolomite Members are quite similar and conform to the general description of the Ledger throughout the valley by prior mappers. Were it not for the intervening Willis Run (dominantly limestone) Member, the upper and lower units could not be practically subdivided. The Lower Dolomite Member is slightly purer than the Upper Dolomite Member with insoluble content rarely exceeding a few tenths of a percent. The Upper Dolomite Member may have insolubles up to two percent, but generally has values below one percent.

Problems of Correlation

The recognition of a thick (up to 210 feet), primarily limestone unit within the dolomite-dominated Ledger Formation in the West York Block is highly significant. The Willis Run limestone is fossiliferous at Delta Carbonate. The fossils, which are discussed in Chapter IX, are the first fossils reported from the Ledger Formation and conclusively demonstrate an Early Cambrian age for the Lower Dolomite and Willis Run Members.

As previously noted, the Longs Park Member (of Gohn, 1976), below the Ledger Formation in Lancaster County, has yielded Middle Cambrian fossils; therefore, the overlying Ledger Formation must be Middle Cambrian or younger. The Early Cambrian fossils recovered from the Willis Run Member in the West York Block demonstrate pronounced diachrony of the Lower Ledger if, in fact, the dolomite in the Lancaster area represents a lithologic continuation of the Ledger as it exists in the York area.

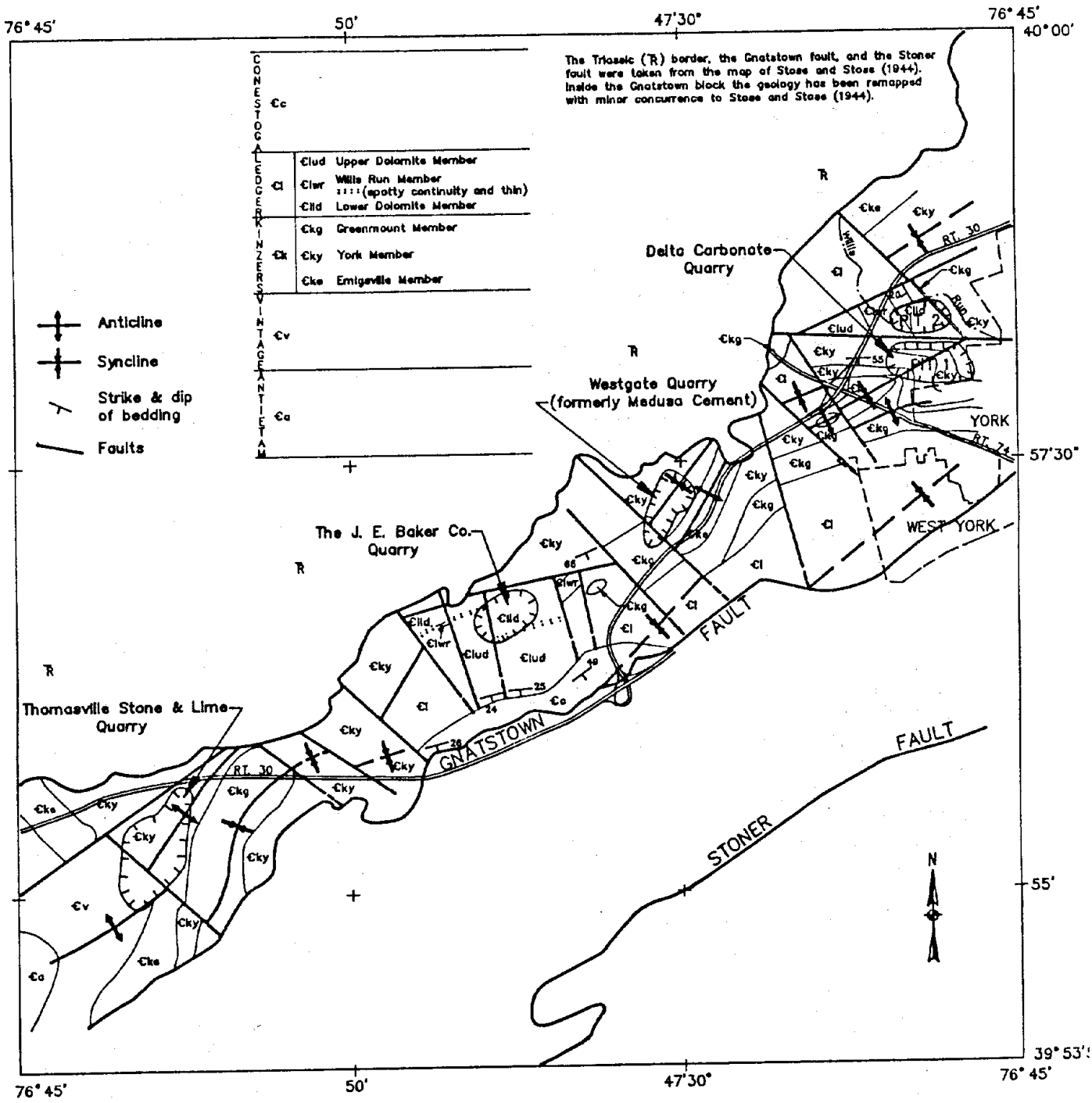
Depositional Environments and Paleogeography

The very pure and oolitic nature of the Lower and Upper Dolomite Members, in the West York Block and elsewhere in the Conestoga Valley, suggest a carbonate platform and/or platform margin environment. The Willis Run Member, which is finer grained, moderately to strongly bioturbate, and less pure (insolubles from 2 to 4 percent), represents a deeper shelf environment.

The Conestoga Formation overlies the Ledger Formation in the vicinity south of the J.E. Baker Company quarry. It will be examined in railroad cuts at STOPS 9 and 10. A variety of lithologies are present, including breccias with white limestone clasts set in a dolomite matrix; grey, platey to thin-bedded lime grainstone/packstone; calcareous shale; and light gray, thickly to massively bedded lime grainstones. We propose a slope setting for this sequence, not too far from the platform margin, such that debris flows composed of carbonate sand, and sometimes coarser
materials were occasionally contributed to the fine-grained siliciclastic sediments of deeper-water basin. The Conestoga Formation is truncated by the Gnatstown Fault in the West York Block; what is present has a thickness of about 1000 feet.

The relationship of the Conestoga Formation to the other units of the Conestoga Valley has been a point of much debate. Jonas and Stose (1930) and Stose and Stose (1944) considered the Conestoga Formation to be an Ordovician unit lying above an unconformity cutting across the Cambrian units. Gohn (1976) also considered the Conestoga Formation to be lying above older units in an unconformable relationship. Rodgers (1968) proposed that the Conestoga represented a deep-water basinal equivalent to all units between the Vintage Formation and Conococheague Group. Rodgers interpreted this relationship as representing the eastern edge of the North American continent during the Cambrian and Early Ordovician time.

Resolution of the problem has been difficult owing to the scarcity of fossils in the Conestoga. Gohn (1976) summarized the few problematical Conestoga fossil occurrences and concluded, "Biostratigraphic data for the Conestoga, therefore, remains sparse and inconclusive." In many areas where the Conestoga crops out, the rocks are highly sheared and metamorphosed to the point where fossils would not be expected. In the West York Block, however, where the rocks are not so deformed as the rest of the valley, some fossils have been recovered. Although the material is limited and not of very good quality, it suggests a Middle Cambrian age for the Conestoga Formation in the West York Block. The fauna recovered is described by Taylor and Durika (Chap. IX, this guidebook).



GEOLOGY OF THE WEST YORK BLOCK

G. R. Gonis and D. A. Hopkins, 1987

Figure VIII-5

STRUCTURE

Figure VIII-5 is a geologic map of the West York Block. The structural style of the West York Block is somewhat enigmatic within the broader structural framework of the region. The upper limit of the West York Block is the Gnatstown "Overthrust" of Stose and Stose (1944) and its probable continuation as their Highmount "Overthrust." Stose and Jonas (1944) delineated, in their mapping, a series of major sub-parallel faults, which they called overthrust faults, that effectively divide up the York County portion of the Conestoga Valley into a series of imbricate blocks (Figure VIII-1). They did not, however, elaborate on stratigraphic or structural contrasts between blocks.

The rocks of the West York Block are far less deformed than most rocks elsewhere in the valley. The West York Block exhibits only one clear episode of folding, which produced broad, open folds with vertical axial planes. High angle, transverse faults, however, are abundant and pose the most serious problems in mapping the West York Block. Also, small scale, wedge-like thrust faulting can be quite well developed near fold axes as a result of space adjustments.

Complex deformational features, i.e., small scale and/or tight folding, overturning, crenulations, flat overthrusting, penetrative cleavage, boudinage, complicated shear zones, deformed fossils and oolites, are scarce to absent in the West York Block. The color index of phosphatic fossils recovered suggests a thermal exposure of most likely 300°C or higher (J. E. Repetski, personal communication). The Emigsville Member of the Kinzers Formation is a phyllite within the West York Block. The rocks of the West York Block, therefore, display an unusual post-burial history of considerable heating but minimal shear.

In central Lancaster County, the rocks of the Conestoga Valley have been affected by complex nappe formation (Wise, Freedman, and Henderson, 1968). Even in Lancaster County, however, a boundary separating areas containing nappe structure from areas of less deformed conditions was identified. (See also Valentino, Chap. III and MacLachlan, Chap. VII, this guidebook). The Conestoga Valley is situated geographically in a position to be possibly affected by both Taconic and Alleghanian orogenic events; however, the effects of these events are not expected to be geographically uniform, nor are the rocks uniform in their responses to orogenic forces. These conditional variables have produced disharmonic local responses to regional deformations. For reasons that are not understood at present, the West York Block has been, overall, deformed to a lesser degree than most areas of the Conestoga Valley.

At STOPS 10 and 11 we will visit outcrops within the West York Block where the comparatively less deformed rocks will be seen. A graphic example of this slighter degree of deformation is

the undistorted fossils collected at these locations and illustrated in Chapter IX.)

At STOP 9, rocks southeast of the West York Block, across the Gnatstown Fault, will be visited. Here the deformational character of the Conestoga Formation more closely resembles that seen at the H. R. Miller Quarry (STOP 7), although the rock at STOP 9 appears to lack the Taconian(?) cleavage seen as S1 at STOP 7. We also know that, in general, the folds within the fault block between the Gnatstown Fault and the Stoner Fault are much tighter than those in the West York Block, and the Emigsville Phyllite has well-developed penetrative cleavage south of the Gnatstown Fault. Overall, deformation seems to increase progressively to the southeast in a stepwise fashion across the series of faults in the Conestoga Valley in York County.

ACKNOWLEDGEMENTS

The writers wish to express their sincere appreciation to the following individuals who contributed valuable insight and advice regarding the geologic interpretations contained in this chapter: John Repetski, Tom Dutro, Avery Drake, John Taylor, Rodger Faill, and Dave MacLachlan.

IX. LITHOFACIES, TRILOBITE FAUNAS, AND CORRELATION OF THE KINZERS, LEDGER AND CONESTOGA FORMATIONS IN THE CONESTOGA VALLEY

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INTRODUCTION

It is the purpose of this paper to summarize what is presently known about the paleogeography, depositional environments, and trilobite faunas of the Lower and Middle Cambrian carbonate platform and proximal off-platform deposits that compose the Kinzers, Ledger, and Conestoga Formations in the Conestoga Valley. In this paper, "Conestoga Valley" refers to the Conestoga Valley Section of the Piedmont Physiographic Province, a broad valley carved primarily into the carbonate strata of several Cambrian formations southeast of the Triassic Lowlands Section and northwest of the Piedmont Uplands Section. A location map is provided as Figure IX-1. The importance of integrating the lithostratigraphy and biostratigraphy lies in the indispensable nature of the faunal data for establishing age relationships of various rock units mapped in this area and for allowing recognition of coeval strata in other areas of the Appalachians and elsewhere in North America. Conversely, a clear understanding of the depositional setting obtained from sedimentological studies of these strata is of great value in evaluating the environmental and temporal significance of the trilobite faunas recovered from the Kinzers, Ledger and Conestoga Formations.

Our knowledge of these rocks and included fossils has increased dramatically over the last two decades, but much remains to be unravelled. Several factors have operated (and continue to operate), in combination, to limit the rate of progress in development of well-constrained stratigraphic models for the Cambrian carbonates of the Conestoga Valley: 1) a scarcity of continuous exposures, making it difficult to impossible to determine the relative stratigraphic positions of distinct faunas and lithofacies, 2) severe physical deformation of much of the Conestoga Valley sequence (with the exception of the West York Block: see discussion below and Chapter VIII), 3) a strong diagenetic, in places even metamorphic, overprint that has erased primary textures and made impossible the recovery of fossils in many areas and stratigraphic intervals, 4) a complex facies mosaic with dramatic changes in thickness and lithology across short lateral distances, a pattern characteristic of shelf-marginal deposits, and 5) an imprecise and outdated biostratigraphic data base in serious need of taxonomic reevaluation and refinement through additional precise and systematic sampling.

RECENT DEVELOPMENTS

Nonetheless, considerable progress has been made over the last few years in more closely constraining stratigraphic and

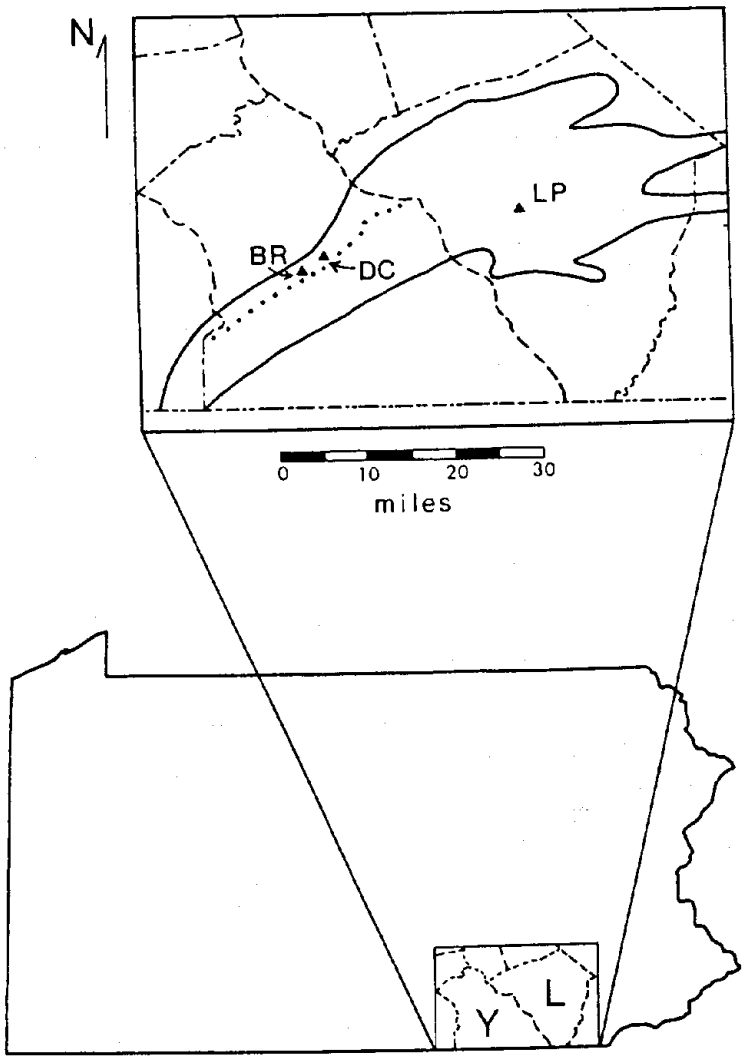


Figure IX-1. Map showing location of the Conestoga Valley (outlined with dark, solid line in insert) in York (Y) and Lancaster (L) Counties, southeastern Pennsylvania, and the field conference stops discussed in the text. LP=Longs Park (STOP 6), BR=Baker Co. RR cut (STOP 10), DC=Delta Carbonate quarry (STOP 11). Gnatstown fault is shown as a dotted line in York County.

structural models for the Conestoga Valley. All five factors listed above have, at least to some extent, been overcome through recent developments (most of them directly attributable to the initiative of Bob Ganis of Tethys Consultants and Dave Hopkins of The J. E. Baker Company). A brief explanation of these developments is provided below.

Lack of Exposure and Complex Structure

Although scarcity of natural exposures and severe physical deformation remain a problem in many areas of the Conestoga Valley, access to several large, active quarries and extensive drill core data from surrounding properties has facilitated recognition and detailed mapping of several members within the Kinzers and Ledger Formations (see Ganis and Hopkins, Chapter VIII, this guidebook) in and around York, Pennsylvania. This mapping has shown that strata in the West York Block (that area of the Conestoga Valley north of the Gnatstown Fault and south of the Triassic overlap) are much less strongly sheared than those within other portions of the Conestoga Valley; complicated structures depicted on previously published geologic maps of this area (e.g. the klippen shown by Stose and Stose, 1944) apparently are artifacts of limited bedrock exposure and extreme lithologic heterogeneity.

Quarry exposures and drill core data reveal that several formations contain isolated pockets or lenses of lithologies that characterize higher or lower rock units. For example, the York Member of the Kinzers Formation locally contains pockets of dolomite indistinguishable from varieties within the overlying Ledger Formation. Conversely, the pure "white marble" so characteristic of the York Member is not restricted to that unit but occurs in some places within the dolomites of the overlying Ledger. In isolated pasture or shallow quarry exposures, such isolated occurrences of characteristic lithologies might easily be misinterpreted as structurally displaced. In brief, it is not the structure that is very complicated, it is the stratigraphy.

Strong Diagenetic/Metamorphic Overprint

The quarries and drill core data in the West York Block also reveal a highly irregular spatial distribution of recrystallization and dolomitization. Because of the non-uniform distribution of these diagenetic/metamorphic processes, pockets and intervals of strata with preserved primary textures and recoverable fossils are found at various levels within the West York sequence; some of the larger and more stratiform of these intervals are mappable and have been identified as members within the Kinzers and Ledger Formations. The depositional characteristics and faunas recently recovered from two such intervals (Willis Run Member of the Ledger Formation and the Greenmount Member of the Kinzers) are discussed in detail in later sections of this chapter.

Complex Shelf-Marginal Lithofacies Mosaic

A detailed depositional model recently developed for the Shady Dolomite in southwestern Virginia (Barnaby and Read, 1990) provides valuable insights into the anatomy of the Appalachian platform margin in Early Cambrian time. This model, developed from extensive drill core data in less deformed strata of identical age and similar depositional setting, significantly reduces the difficulty of recognizing and accurately interpreting the

origin and relationships of the numerous platform-margin and proximal off-platform lithofacies present in the Conestoga Valley.

Outdated and Imprecise Biostratigraphic Data Base

In the relatively undeformed rocks of the West York Block, it is possible (with some caution because of numerous high-angle transverse faults) to measure considerable thicknesses of strata and establish with confidence the relative stratigraphic position of fossiliferous horizons within each measured section. We began systematic sampling in the West York Block in the fall of 1989 and have continued intermittently through 1990. The preliminary results are discussed below and shown in Figure IX-2 and IX-3. We optimistically view these results as a first step toward a more current and precise biostratigraphic framework for correlation of these units within and beyond the Conestoga Valley.

REGIONAL DEPOSITIONAL SETTING

Faunal provincialism was well developed in the Early Cambrian. The existence in the Iapetus Ocean of oceanographic barriers to dispersal and interaction caused the development of distinct faunas that now are found in the Lower Paleozoic strata of the North Atlantic region. These faunas allow those rocks deposited on or near Laurentia (North America) to be identified and distinguished from sedimentary sequences that originated on or near other continents and island areas within the Iapetus Ocean (Conway-Morris and Rushton, 1988; Theokritoff, 1979, 1985). The taxonomic composition of their trilobite faunas demonstrates conclusively that the Lower Cambrian strata of the Conestoga Valley were deposited on or immediately adjacent to the margin of the Laurentian continent. The presence of such endemic, characteristic Laurentian trilobite genera as *Olenellus* and *Protypus* rule out the possibility that these strata constitute part of an accreted terrane. There is, in fact, no evidence anywhere in the central Appalachians of "exotic" or accreted Early Paleozoic terranes (Avalon and Meguma) like those sutured to North America in the northern and southern Appalachians (Williams and Hatcher, 1982).

Comprehensive treatments of Cambrian faunas and sedimentary units in the central Appalachians have been provided by Palmer (1971) and Read (1989); the former includes more detailed information on Cambrian faunas in the Appalachian region while the latter represents a recent analysis of the sequence stratigraphy of Lower Paleozoic passive-margin carbonates in the central and southern Appalachians. From these regional syntheses it is clear that the Conestoga Valley is one of only two places in the entire length of the Appalachians where deposits of the Early Cambrian carbonate platform margin are preserved intact. (As previously mentioned, Early Cambrian platform margin deposits are also preserved intact in the Shady Dolomite in southwestern Virginia.) In other areas, Lower Cambrian shelfbreak deposits either were

completely destroyed during closure of the Iapetus Ocean, or they are preserved only as olistoliths within toe-of-slope limestone conglomerates (olistostromes) now incorporated in major allochthons in the northern Appalachians. The well-known, highly fossiliferous boulders of the Levis Conglomerate in Quebec (Rasetti, 1943, 1944, 1946, 1948) are a good example of this style of preservation.

The Kinzers, Ledger and Conestoga Formations fall within sequence 2 and sequence 3 of Read (1989). In the West York Block, the Conestoga Limestone lies within sequence 3 while the Kinzers and Ledger represent sequence 2 (primarily 2B, the upper half of that sequence). During deposition of sequence 2B, the Appalachian platform was a high-relief, rimmed shelf with a very narrow (1 to 1.5 kilometer) belt of cemented algal bioherms and well-winnowed carbonate sands at the seaward edge of a broad, shale-dominated shelf (Barnaby and Read, 1990). Platform-to-basin relief was several hundred to more than a thousand meters (Read, 1989). Polymictic periplatform breccias and lime grainstones (foreslope carbonate sands) formed at the base of the slope and gave way seaward to rhythmites consisting of dark basinal shales interbedded with thin beds of limestone deposited as turbidites. This facies mosaic is best preserved and thoroughly documented in the Shady Dolomite in southwestern Virginia (Barnaby and Read, 1990). It is a rather unusual depositional pattern for the Lower Paleozoic passive margin in that carbonate deposition was restricted to a narrow belt at the platform margin. Except for sequence 1, which represents an early post-rift phase of clastic shelf deposition, the Appalachian passive margin normally was a broad carbonate platform with shale deposition occurring only in intrashelf basins and off-platform areas (Read, 1989).

The depositional model developed for the Shady Dolomite is of value in study of Conestoga Valley carbonates in at least two respects:

1) The Upper Shady Dolomite provides solid evidence of a very narrow, constructional carbonate rim whose upward growth produced a very thick sequence of carbonate strata that are replaced only a few kilometers to the east by shale-dominated basinal deposits. This pattern of rapid lateral facies change can be used to explain the dramatic increase in thickness and carbonate content displayed by the Kinzers Formation from the Lancaster area to the West York Block. Without that independent evidence of strong facies contrast, the dramatic differences in thickness and composition between the structural blocks of the Conestoga Valley might appear explainable only by substantial structural telescoping of the area with significant structural transport along the faults bounding those blocks.

2) Many lithofacies well-preserved within the Upper Shady Dolomite are also recognizable in the carbonates of the Conestoga Valley. Specific examples are provided below in the section on Lithofacies and Depositional Environments. The origin and significance of these lithofacies often are more easily established

in Virginia where the structural and diagenetic/metamorphic overprints are not as strong.

LITHOFACIES AND DEPOSITIONAL ENVIRONMENTS

Rodgers (1968) was the first to recognize the dramatic changes in thickness and lithofacies of Cambro-Ordovician strata across the Conestoga Valley as signatures of the transition from platform to off-platform deposits. The most recent and thorough study on the sedimentology of Conestoga Valley carbonates, however, is that of Gohn (1976). The depositional environments described for the formations and members below are based in large part on that study with additional insights gained from our own field observations and from similarities noted with lithofacies recognized in the Shady Dolomite. The Conestoga Valley provides ample opportunity to follow transects from basinal lithofacies, through toe-of-slope sediment accumulations, into carbonate platform deposits. This can be accomplished both 1) laterally, by travelling northwestward across the valley from one fault block to the next, and 2) vertically, by moving up-section within the stratigraphic sequence provided in an individual fault block. We will do both during the field conference. The optimal approach is to climb through the stratigraphy of the West York Block; this provides a view of the lithofacies where they are least deformed and most easily related, spatially and temporally, in a relatively uninterrupted superpositional sequence.

The Lower Cambrian carbonates of the West York Block compose a shoaling-upward sequence. The off-platform facies of the Vintage and Kinzers Formations give way upward to the shallow platform deposits of the overlying Ledger Dolomite, recording the progradation of the carbonate platform through the Early Cambrian. Three thin packages of dark, deep shelf or off-shelf lithofacies (Emigsville, Greenmount and Willis Run Members) record brief interruptions in this general pattern of shallowing and progradation. Subsequent retreat of the platform, probably in the Middle Cambrian, resulted in deposition of toe-of-slope and basinal facies (the Conestoga Limestone) atop the Ledger Dolomite throughout the Conestoga Valley. A revised stratigraphic model, developed to accommodate new biostratigraphic data from the West York block, is presented below in the section entitled Coarse Biostratigraphy and Depositional History.

Kinzers Formation

The three members of the Kinzers Formation differ sufficiently to warrant separate discussion of their features and implied environments of deposition. All three members are assigned to off-shelf environments, their differences attributed to relative proximity of the platform margin.

Emigsville Member

This basal member of the Kinzers is a laterally persistent basinal facies with two components. The basal third of the member is argillaceous dolomite and dolomitic limestone. This carbonate interval is overlain by a shale (or phyllite, depending on structural position) interval. The member is well known for its rich Lower Cambrian fauna (Resser and Howell, 1938; Campbell, 1969; Ryan, 1987). The basal carbonates are interpreted as turbidites derived from the carbonate platform, an origin similar to that of the underlying Vintage Dolomite. The laterally persistent shale/phyllite records a period of reduced carbonate sediment influx into basinal environments. The reason for this is not well established. We have observed thin greenish mudstones at two other horizons higher in the West York Block sequence that may record similar but less prolonged interruptions in carbonate production/influx. These are noted below.

York Member

Although post-burial processes have obliterated the primary textures in many parts of this member, enough features remain to allow recognition of this middle member of the Kinzers as primarily a foreslope facies of considerable lithologic heterogeneity. It includes well-bedded lime mudstones, oolitic and bioclastic lime grainstones, and minor occurrences of periplatform breccia. Gohn (1976) called this member the Thomasville Member after exposures (unfortunately, atypical exposures: see Chapter VIII) with megabreccias in the Thomasville Quarry. Oolitic intervals display tabular bedding with well- developed normal grading and are interpreted as down-slope accumulations of shelf-derived sediment. This member, which thickens from less than 100 feet in the Lancaster area to more than 1000 feet in the West York Block, is equated (at least in part) with the proximal periplatform deposits that accumulated to an approximate thickness of 600 meters (nearly 2000 feet) immediately seaward of the constructional rim facies in the Shady Dolomite (Barnaby and Read, 1990). However, it is possible that some (perhaps much) of the York Member would be more appropriately correlated physically with the Patterson Member, the basal member of the Shady Dolomite. The Patterson Member was deposited in a deep, subtidal carbonate ramp setting prior to development of the high relief shelf margin. Reports of archaeocyathid reefs in the middle member of the Kinzers in York County (Stose and Jonas, 1939; Stose and Stose, 1944) invite this comparison, in that mud mounds constructed in part by archaeocyathids are abundant in the Patterson Member. Some of the well-bedded, darker lime mudstone intervals in the York Member might then be interpreted as off-mound ramp deposits similar to those that compose much of the Patterson Member. Additional scrutiny of the York Member's components is needed to resolve this matter of platform profile and evolution.

Greenmount Member

The highest member of the Kinzers Formation is a distinctive, impure carbonate whose features suggest a second deepening event (or decline in carbonate production/influx) toward the end of Kinzers deposition. Fine grained siliciclastic sediment is abundant in this dark, pyritic, laminated to somewhat nodular limestone, occurring as black shaley interbeds and dispersed silt-sized quartz and feldspar grains. Because of the high insoluble content, the unit forms a prominent ridge (this is the upper "sandstone" member of Stose and Stose, 1944). Many beds yield fossils. The trilobite faunas of this member are described in detail in a later section. The occurrence of the eodiscid trilobite genus *Pagetides* in this member provides additional evidence of off-platform deposition. This genus has been reported only from deep marine shales and limestone boulders in toe-of-slope conglomerates (Rasetti, 1948; Shaw, 1955; Theokritoff, 1979; among others). In addition to trilobites and brachiopods, the limestone beds contain clusters of dark, hollow, elongate quartz crystals suggestive of sponge spicules but displaying euhedral growth lamellae throughout. These crystals are clearly primarily authigenic but their hollow interiors and concentration in pockets of limestone along with other bioclasts would still seem to call for interpretation as "reconstituted" siliceous spicules.

Ledger Formation

As noted by Gohn (1976), the dolomites of the Ledger Formation contain few preserved primary features for environmental analysis. The few that remain reflect much shallower deposition than that interpreted for the underlying Kinzers and Vintage formations. The recent discovery of a limestone member near the middle of the Ledger (Ganis and Hopkins, Chapter VIII, this guidebook) and analysis of identical facies in the Shady Dolomite confirm that this formation is primarily a platform facies. It is possible, however, that some parts of this formation originated in a periplatform setting.

Upper and Lower Dolomite Members

These massive light-colored dolomites, virtually devoid of fine-grained siliciclastic impurities, contrast markedly with the dark-colored, shaley basinal facies that immediately underlie (Greenmount Member) and overlie (Conestoga Formation) them. Relict oolitic textures are abundant and cross-stratification is preserved in some places. Virtually identical lithologies in the Austinville Member of the Shady Dolomite characterize the zone just landward from the organic buildups that formed the platform rim; these massive dolomites are logically interpreted as dolomitized back-reef and shelf-margin sands (Barnaby and Read, 1990). Whether the Ledger dolomites also include the algal framestone facies, i.e., the well-cemented reef itself, is difficult to establish owing to the scarcity of primary textures. The reef lithofacies has been documented in the Willis Run Member (the

newly defined middle member of the Ledger) in one surface exposure.

In any case, the evidence is strong that the sharp contact between the Lower Dolomite Member and the underlying Greenmount Member of the Kinzers Formation represents the establishment of shallow platform conditions in areas previously occupied by deep-water, off-platform environments. In one core that we examined, a greenish mudstone occurred at this formational contact, suggesting that the transition involved a period during which carbonate deposition slowed or ceased completely.

Willis Run Member

This recently discovered interval of limestone within the Ledger Formation in the West York Block has provided valuable insights to the age and environmental conditions of Ledger deposition. The fossils recovered from the Willis Run Member and their significance are discussed in the following section. Here we wish to describe the lithofacies and discuss their significance. The features of the member have been documented and fossils recovered from exposures in the upper part of the Delta Carbonate Quarry (STOP 11) and from small pasture exposures and drill cores through the unit just north of the J. E. Baker Quarry.

In the quarry exposures the member displays remarkable uniformity, in stark contrast with the overwhelming lithologic variability of the other carbonate units in the quarry (specifically the York Member of the Kinzers and the Lower Dolomite Member of the Ledger). The Willis Run here is essentially monofacial, consisting of moderately bioturbate, thinly bedded, somewhat nodular lime mudstone to wackestone with abundant thin dolomitic laminae that are disrupted to varying degrees by burrowing and possibly by compaction.

The generally fine-grained character, fairly extensive bioturbation, and the absence of shallow water features (desiccation cracks, well-winnowed lime grainstones, stromatolites and thrombolites, etc.) or even tempestites (coarse-grained storm beds) indicates deposition in a fairly deep subtidal environment below storm wave base. The three- dimensional pattern of burrowing (ichnofabric) and the absence of characteristic turbidite features (normal grading, parallel laminations, etc.) in the limestones confirms a deeper shelf, rather than off-shelf, environment for these strata. A single, distinctive horizon within the member, however, suggests a period of even deeper-water deposition. This horizon, which appears as a thin but prominent *reentrant with limonitic staining on the quarry wall, is a highly siliceous, pyritic bed immediately overlain by very dark (organic-rich) limestone that grades upward into the normal burrow-mottled subtidal lithology previously described. Also noteworthy is a thin (1-2 foot) interval, near the base of the member, of greenish mudstone similar to that observed in core material at the base of the Ledger Formation, and similarly inter-

preted as representing arrested carbonate production and/or influx.

The possibility that the Willis Run Member is not a depositional package, but merely non-dolomitized portions of the sedimentary facies that (where dolomitized) produced the Ledger dolomites, has been considered but rejected for the exposures in the Delta Carbonate Quarry. Neither the relict primary textures nor the overall appearance of the Ledger dolomites are consistent with derivation from a fine-grained, subtidal lithofacies like that which dominates the Willis Run Member at that location. We interpret these strata as the result of deep subtidal deposition in a small intrashelf basin.

Limestones assigned to the Willis Run Member in drill cores and small pasture exposures near the J. E. Baker Company Quarry, however, are quite different; shallow platform lithologies are at least as common as the bioturbate fine-grained subtidal lithofacies. Cross-stratified oolitic lime grainstones are common and, in some intervals, show hematitic staining suggestive of exposure. Light-colored fenestral lime mudstone (algal boundstone?) was seen in drill core. One small field exposure has revealed a reef facies comprising stromatolitic lime boundstone with marine, cemented shelter cavities and associated bioclastic, lithoclastic, and pisolitic and/or oncolitic lime grainstones. The drill core data show the member to be lenticular in this area, pinching laterally and actually disappearing in some places. In the J. E. Baker Quarry, for example, there is no limestone separating the Upper and Lower Dolomite Members. There is, however, an unusually well-bedded interval at the base of the Upper Dolomite Member, suggesting that the Willis Run Member may be present but dolomitized at that location. Drill core data have confirmed an analogous situation in the Greenmount Member in one area where, although completely dolomitized, the unit is still recognizable on the basis of its laminated, pyritic character.

Conestoga Limestone

The lithofacies of this formation leave little doubt as to the depositional setting. The most spectacular lithofacies is that of polymictic megaconglomerate or megabreccia. These massive, unsorted toe-of-slope debris flow deposits include clasts ranging upward to 30 feet or more (although the vast majority are much smaller boulder-to cobble-sized clasts). Other lithologies within the formation include lithoclastic lime grainstone consisting largely of platy clasts of dark laminated limestone, massive to thickly bedded peloidal or oolitic lime grainstone, thinly bedded limestone-shale rhythmite, black graphitic limestone, and black phyllite. A member stratigraphy proposed by Gohn (1976) for this formation reflects systematic variation in relative abundance of these lithologies both up-section in the West York Block and laterally from southeast (distal) to northwest (proximal) across the Conestoga Valley. The West York Member, characterized by an abundance of megaconglomerates and minor proportion of shaley lithofacies, occurs at the base of the

formation in the northwestern part of the valley. This member is a proximal toe-of-slope facies consisting primarily of coarse sediments transported by gravity-flow mechanisms from the adjacent shelf margin.

Excellent analogs are available farther to the north and the south in the Appalachians. The periplatform breccias of the Upper Shady Dolomite in Virginia (Barnaby and Read, 1990) differ only in the degree of recrystallization. Large, light-colored clasts in the Conestoga megabreccias are all recrystallized to marble while those within Shady Dolomite breccias retain primary textures including algal boundstone with well-preserved isopachous fibrous marine cements, oolitic lime grainstone, and other obviously shelf-derived reef and rim facies lithologies. Similar proximal toe-of-slope debris flow deposits with included algal framestone clasts are also found in abundance and pristine condition in the somewhat younger Cow Head Group in the northern Appalachians (James, 1981; James and Coniglio, 1985; James and Stevens, 1986; James and others, 1989).

The West York Member of the Conestoga Formation is replaced to the southeast by the Kreutz Creek Member which consists primarily of dark lime mudstones and phyllite. The lithoclastic limestones are fewer in the Kreutz Creek Member and are finer grained, lacking the megaclasts so prevalent in the West York Member. These differences within the Conestoga Limestone clearly document a proximal to distal trend toward the southeast across the Conestoga Valley. Gohn (1976) also documented proximal-distal trends in the Wrightsville Member, which overlies both the West York and Kreutz Creek Members. The trends involve 1) a general increase in percentage of argillaceous lithologies relative to peloidal and lithoclastic limestones toward the southeast, and 2) a change from lithologic associations indicating slope or proximal submarine fan deposition in the northwest to associations characteristic of mid-fan environments to the southeast.

TRILOBITE FAUNAS AND CORRELATION

The exceptional Lower Cambrian fossils of the Kinzers Formation are some of the most famous fossils in the central Appalachians (Resser and Howell, 1938; Campbell, 1969; Ryan, 1987). The shales of the Emigsville Member at the base of the formation have yielded not only an impressive array of large, complete olenellid trilobites but also a variety of remarkable "soft-bodied" forms reminiscent of (although somewhat older than) the Burgess Shale in British Columbia. A list of the taxa reported includes such familiar genera as *Anomalocaris* and *Sidneyia*. A wide variety of non-mineralized algae, annelids, sponges, and cnidarians have also been recovered. The focus of this paper, however, is the somewhat less spectacular but very useful trilobite faunas of the overlying carbonate units.

Trilobite faunas from the carbonate units above the Emigsville have been discussed by Walcott (1896), Resser (1938),

Stose and Jonas (1939), Stose and Stose (1944), and Gohn (1976, 1977). However, the most comprehensive treatment of the Kinzers faunas is that of Campbell (1969) who described one trilobite-bearing fauna from the basal strata of the middle member of the Kinzers and four other faunas that include trilobites from the upper member. The two highest faunas, recovered by Campbell from the highest beds of the Kinzers in the Longs Park section near Lancaster (STOP 9), include definitive Middle Cambrian genera (*Ogygopsis* and *Peronopsis*). This established a Middle Cambrian (or younger) age for the Kinzers-Ledger contact in that section. The lowest fauna from the upper member includes the definitive Lower Cambrian genus *Bonnia*, indicating that the boundary between the Lower and Middle Cambrian lies somewhere within the upper member of the Kinzers in the Lancaster area. The complications that this age determination has created for correlation to the York area are discussed below.

Unlike the Upper and Middle Cambrian, the Lower Cambrian has not been extensively subdivided into numerous zones and subzones. A single biozone, the *Bonnia-Olenellus* Zone, encompasses almost all of the Lower Cambrian in the Appalachian region. In the West York Block, for example, this zone is at least 2000 (and probably more than 3000) feet thick, including at least part of the Antietam Formation, all of the Vintage and Kinzers Formations, and the lower two members of the Ledger Formation. Earlier attempts to refine Lower Cambrian biostratigraphy in the Appalachians through identification of zones established for inter-regional correlation throughout North America (Resser, 1938; Howell and others, 1944; Lochman Balk and Wilson, 1958) were not successful, primarily because of regional differences in age-equivalent faunas and the use of genera (rather than species) for definition of the zones. What clearly is needed is greater taxonomic precision and regional focus--an Appalachian standard zonal sequence based on thorough documentation of the vertical distribution of trilobite species. Other fossil groups are also common in the Lower Cambrian carbonates of the Appalachians and many have considerable potential for contributing to this refinement. Small phosphatic fossils such as inarticulate brachiopods and discinellids are of particular interest given their demonstrated biostratigraphic utility in Lower Cambrian sequences outside the Appalachians and their potential for recovery from dolomitized and recrystallized carbonates.

It is widely known that greater resolution is possible. Campbell (1969), for example, documented the existence of at least 6 distinct trilobite-based "faunules" within the *Bonnia-Olenellus* Zone in the Conestoga Valley and was able to establish relative ages for most, despite the structural complexity and poor exposure in the area. We have recovered at least two other trilobite assemblages that characterize specific intervals within that zone (see Figure IX-3 and the accompanying discussion in the following section). The work of Willoughby (1977) in the southern Appalachians is also noteworthy in the context of improving Lower Cambrian biostratigraphy in this region.

The primary obstacle to establishment of a workable biostratigraphic framework within the Lower Cambrian in this region has been the lack, or imprecision, of stratigraphic context for faunal collections because of structural complications and lack of continuous exposure. The relatively simple structure and extensive quarry exposures in the West York Block provide an opportunity to overcome this obstacle and work toward a biostratigraphic standard for Lower Cambrian carbonates in the Appalachian region.

Coarse Biostratigraphy and Depositional History

Figure IX-2 shows the stratigraphic levels from which identifiable trilobites were recovered in the West York Block. Only the highest collection, obtained from lime grainstone beds in the Conestoga Limestone at STOP 10, includes Middle Cambrian material; all other collections are unquestionably Lower Cambrian. We recovered fragmentary trilobite material from the Conestoga locality and assign one cranidium to the Middle Cambrian genus *Modocia*. Inarticulate brachiopods recovered from samples from the same locality acidized by John E. Repetski (U. S. Geological Survey) were assigned by A. R. Palmer (Geological Society of America) and A. J. Rowell (University of Kansas) to *Prototreta*, a genus restricted to the Middle Cambrian.

The highest beds exposed in the Willis Run Member of the Ledger Formation in Pit 2 of the Delta Carbonate quarry (STOP 11) are highly fossiliferous. The weathered surfaces of these beds have provided numerous trilobite fragments, brachiopod valves, and other fossils. The skeletal material, which is very well preserved, is found primarily (perhaps exclusively) in dolomitic laminae. J. T. Dutro (U. S. Geological Survey) and A. R. Palmer identified olenellid trilobites from these beds, establishing an Early Cambrian age for most or all of the Willis Run Member. We have since recovered *Zacanthopsis virginica*, a species characteristic of the upper part of the Lower Cambrian.

The Early Cambrian age established for the upper member of the Kinzers and lower members of the Ledger Formation contrasts with the Middle Cambrian age established for the uppermost Kinzers and overlying Ledger Dolomite in the Lancaster area. Earlier notions (Campbell, 1969; Gohn, 1976) that the shaley upper member of the Kinzers represents an essentially synchronous transgressive tongue of basinal clastics into more proximal periplatform deposits apparently were incorrect. A revised stratigraphic model, consistent with the new biostratigraphic data, is provided in Figure IX-2. The details of this model and its implications for depositional history and platform margin evolution in the Conestoga Valley are described in the following paragraphs.

In the absence of evidence that the unit is diachronous, it is assumed that the Emigsville Member records a period of basinal hemipelagic deposition that interrupted accumulation of carbonate turbidites (Vintage Formation) in periplatform environments, affecting the entire Conestoga Valley simultaneously. Emigsville

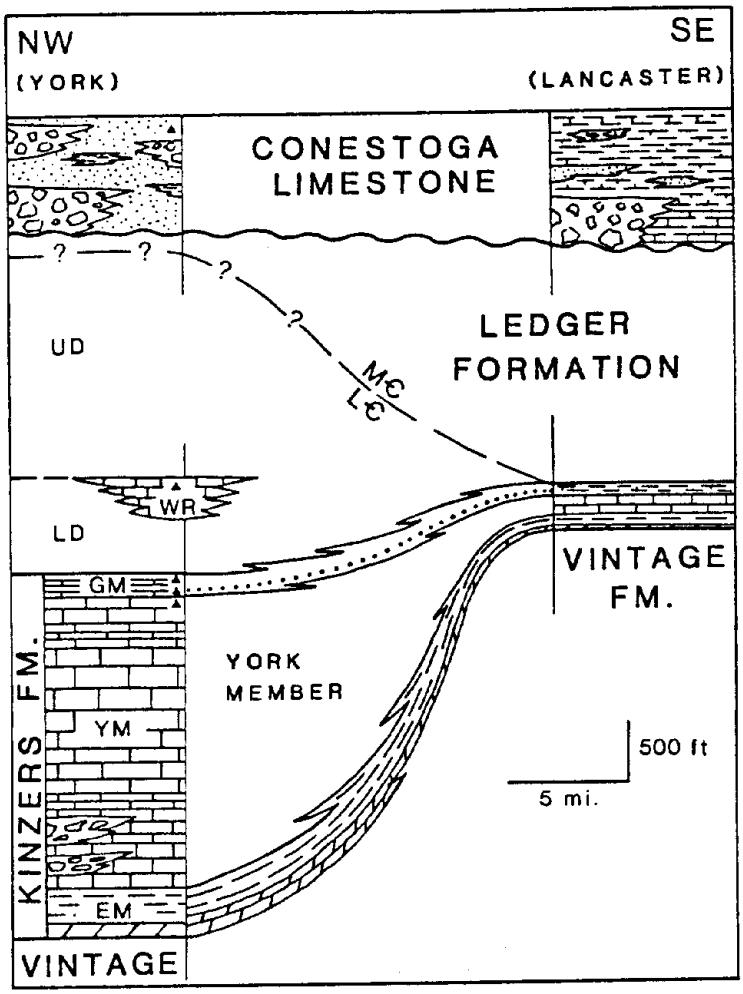


Figure IX-2. Revised stratigraphic model for Lower Cambrian carbonates in the Conestoga Valley which record seaward progradation of platform carbonates (dolomites of the Ledger Formation) over periplatform facies (Kinzers Formation) through Early Cambrian and earliest Middle Cambrian time. Letter pairs in the York column are initials of members established for the Kinzers and Ledger Formations in the West York Block: EM, YM, and GM mark the Emigsville, York, and Greenmount Members of the Kinzers Formation; LD, WR, and UD denote Lower Dolomite, Willis Run, and Upper Dolomite Members of the Ledger Formation Black triangles on the right side fo the York column mark horizons from which identifiable trilobites have been recovered. Question marks on the Lower-Middle Cambrian boundary in the York area reflect uncertain placement of this horizon (within Upper Dolomite Member vs. at the Ledger-Conestoga contact). A late Early Cambrian time line (dotted line) within the upper member of the Kinzers is somewhat speculative, based on the assumption that the Greenmount deepening event is recorded in a black shale interval witin the Langs Park Member near Lancaster. Clastic lithologic symbols are used in the Conestoga Formation to denote carbonate breccias and lime sands; the calcareous shale symbol denotes limestone-shale rhythmite.

deposition was followed by reestablishment of dominantly carbonate sedimentation in a deep ramp or periplatform setting in the West York area that resulted in accumulation of more than 1000 feet of carbonate sediment (York Member). At the same time, distal off-platform environments in the Lancaster area received considerably less shelf-derived sediment, forming a much thinner middle Kinzers unit in that area. A brief(?) interruption in the ramp/periplatform deposition in the York area caused accumulation of dark, shaley carbonates now identified as the Greenmount Member. This event may be recorded in the Lancaster area as a shaley interval (possibly in the middle to lower part of the Longs Park Member) but additional biostratigraphic data from both areas are needed to evaluate this hypothesis.

Greenmount deposition ended as progradation of the carbonate platform brought shoal-water carbonate environments into the York area. Through the late Early Cambrian, well-winnowed carbonate sands (and some algal boundstones) accumulated to form the Ledger Dolomite in the northwestern areas of the Conestoga Valley while basinal deposition continued to the southeast in the Lancaster area, forming the shales and shaley lime mudstones of the upper Kinzers. At the same time, deeper shelf deposition occurring in small, isolated intrashelf basins in the York area, created the burrow-mottled lime mudstone facies of the Willis Run Member of the Ledger Formation. With continuing progradation of the platform, the shoal-water carbonates eventually reached the Lancaster area sometime (early?) in the Middle Cambrian.

The biostratigraphic data from the West York Block are too imprecise to establish the lithostratigraphic position of the Lower-Middle Cambrian boundary; they indicate only that it lies 1) within the Upper Dolomite Member of the Ledger Formation, 2) within lower part of the Conestoga Limestone, or 3) at the contact between the Ledger and Conestoga Formations. The presence of Middle Cambrian platform carbonate facies (Ledger) in the Lancaster area, along with well established proximal-distal trends in the Conestoga indicating that the shelf margin lay to the west when it was deposited, makes it highly unlikely that the base of the Middle Cambrian lies within the Conestoga. A more reasonable interpretation is that deposits of similar age to the Ledger Dolomite of the Lancaster area lie within the Upper Dolomite Member of the Ledger in the West York Block. Alternatively, strata of that age in the West York Block may have been lost to erosion and are now represented only by an unconformity at the base of the Conestoga Formation. This contact (Ledger-Conestoga) has not yet been examined either in quarries or drill cores in the West York Block. The profound facies contrast (toe-of-slope debris flows immediately overlying shoal-water platform lithofacies) and the sharp, erosional nature of this formational contact documented in the Lancaster area (Stose and Jonas, 1939, Plate 18) suggest that it is a disconformity. The Middle Cambrian fossils recovered from the Conestoga in the West York area, however, show that the magnitude of that unconformity is considerably less than that envisioned in earlier studies that assigned an Ordovician age to the Conestoga Limestone.

Regardless of its nature (conformable or not), the base of the Conestoga records a retreat of the platform margin to the west, presumably sometime in the Middle Cambrian. The cause of this retreat is not clear. Although the position of the stratigraphic break in the West York Block (between Lower and Middle Cambrian) invites comparison with regressive signatures elsewhere in North America that are attributed to the Hawke Bay Event of Palmer and James (1979), the presence of Middle Cambrian faunas in the upper Kinzers in the Lancaster area indicates that the backstep of the margin postdates that eustatic event. It is hoped that additional faunal data and scrutiny of the Conestoga-Ledger contact will provide useful information regarding the nature and mechanism of platform withdrawal.

Detailed Biostratigraphy in the West York Block

Bed-by-bed sampling for trilobites in some of the more fossiliferous intervals of the West York sequence has provided results that represent a small but encouraging step toward development of a refined Lower Cambrian biostratigraphy for this region. Some of the trilobite species recovered are illustrated in Figure IX-4. Figure IX-3 is a range chart showing the vertical distribution of trilobite species documented for the Greenmount Member of the Kinzers Formation in Pit 2 of the Delta Carbonate Quarry (Stop 11). It was this kind of thorough, systematic sampling of continuous sections that produced a precise biostratigraphic framework of thin trilobite zones and subzones in North American Middle and Upper Cambrian strata (Palmer, 1954, 1965; Robison, 1964; Winston and Nicholls, 1967; and Stitt, 1971, 1977, to cite just a few).

As shown in Figure IX-3, beds at the top of the Greenmount Member yield a slightly different trilobite fauna from that of the strata near the base of the unit. The lower fauna is characterized by *Pagetides leiopygus* and *Periomella yorkensis*. The higher fauna lacks those species but includes *Bonnia occipitalis* and *Protypus marginatus*, two species not found in the lower fauna. All four of these species also occur in the well known limestone conglomerates of the lower St. Lawrence Valley, Quebec (Rasetti, 1948). These conglomerates are Lower Ordovician olistostromes with Lower, Middle, and Upper Cambrian limestone boulders enclosed (often all together in the same bed) in a matrix of black shale. Different boulders often contain very different trilobite assemblages. Individual boulders from three localities (Levis, Orleans, and Bic) in Quebec contain the four trilobite species that define two faunas in the Greenmount Member. The species associations in those boulders are identical to those documented in the Conestoga Valley. The single boulder from Bic yielded *Bonnia occipitalis* and *Protypus marginatus* but did not contain either *Pagetides leiopygus* or *Periomella yorkensis*. Conversely, the boulders from Levis and Orleans contained *Pagetides leiopygus* and *Periomella yorkensis* but did not yield *Bonnia occipitalis* or *Protypus marginatus*.

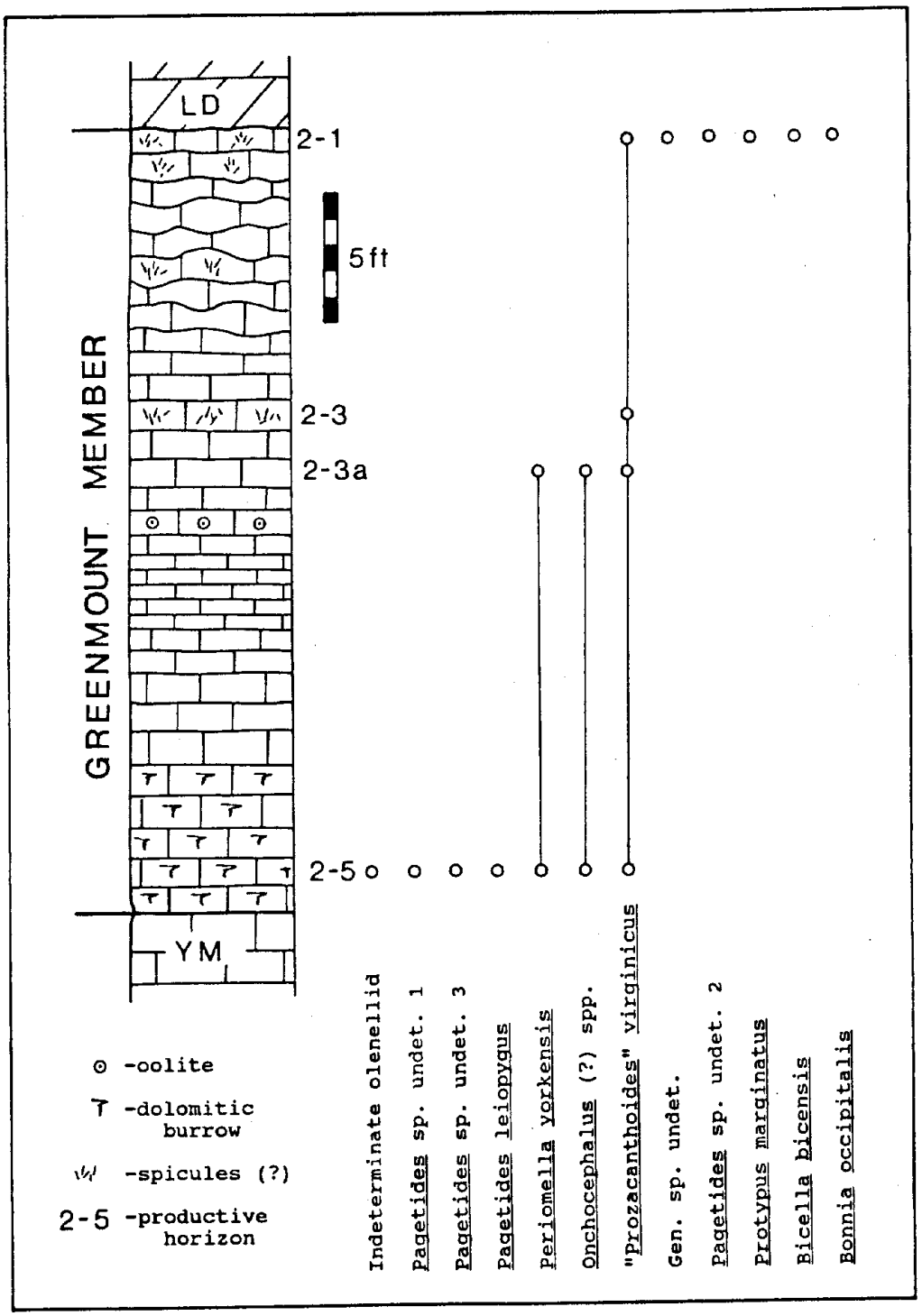


Figure IX-3. Detailed section and range chart showing stratigraphic distribution of trilobite species recovered from the Greenmount Member of the Kinzers Formation in Pit #2 of the Delta Carbonate quarry (STOP 11). Overlying and underlying rock units are the Lower Delomite member (LD) of the Ledger Formation and the York Member (YM) of the Kinzers Formation. Numbers to the right of the column (e.g. 2-5) denote productive sample horizons.

2. position within the Lower Cambrian - Although still rather loosely constrained, these faunas apparently occur somewhere within the middle to upper part of the *Bonnia-Olenellus* Zone in the central Appalachians. This conclusion is based on the recovery of *Olenellus* from strata 1500 to 2000 feet below the base of the Greenmount Member and also from levels at least 3-400 feet above the top of the member.

The recovery of these distinct faunas in stratigraphic context in the Conestoga Valley significantly enhances their biostratigraphic potential. The range data from the Greenmount Member establish the following relationships which had remained uncertain owing to the nature of occurrence (in olistostrome boulders) in Quebec:

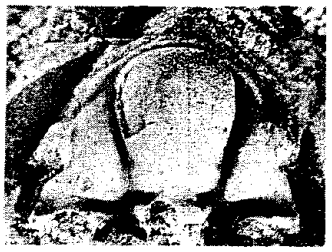
1. relative age - The relative stratigraphic position of the faunas indicates that the *Bonnia occipitalis* fauna is slightly younger than the *Pagetides leiopygus* fauna.

3. temporal vs. spatial (environmental) contrast - The occurrence of the two faunas at different levels within essentially the same lithofacies (the dark, shaley limestones of the Greenmount Member) strongly suggests a temporal, rather than environmental, contrast. In other words, they appear to be faunas that inhabited similar environments at different times rather than species associations of similar age but different environmental settings (biofacies).

The preceding discussion of preliminary results from high-resolution sampling of the Greenmount Member is provided as an example of the potential that exists in the fairly fossiliferous sequence of the West York Block for recovery of biostratigraphic data that may be brought to bear on local and regional geologic problems. Additional sampling is underway in exposures of the Willis Run Member of the Ledger Formation and still more is planned for what has been reported (Gohn, 1976) to be an unusually fossiliferous section of the York Member in the Thomasville Quarry. The data recovered in that sampling will likely answer many questions regarding the stratigraphy and depositional history of the Conestoga Valley and will pose, we hope, even more.

ACKNOWLEDGMENTS

All science is, to some extent, a collective endeavor. For no discipline is this more true than it is for stratigraphy. We have drawn heavily from the monumental works of others in constructing the stratigraphic model presented herein. Special thanks are due G. R. Ganis and D. A. Hopkins, in particular, for introducing us to the stratigraphy of the Conestoga Valley and sharing their insights on the complexity of the rock units involved, and the patterns that they have discerned despite that variability. The work of G. S. Gohn and L. D. Campbell on the sedimentology and paleontology, respectively, of the area created a framework for correlation and comprehension to which we have really added only minor adjustments. The access provided to quarry exposures and core materials by the J. E. Baker Company



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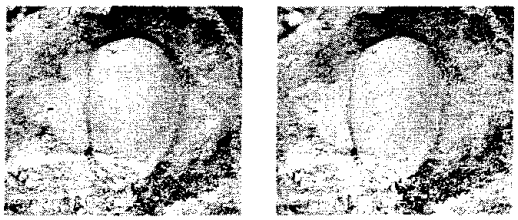

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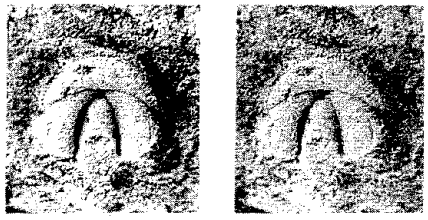

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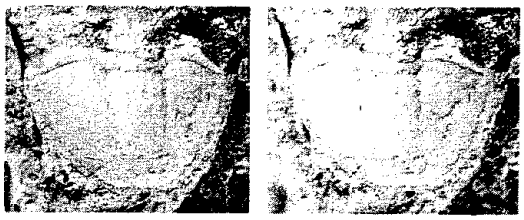

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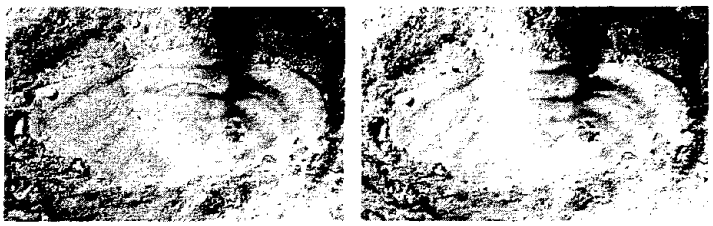

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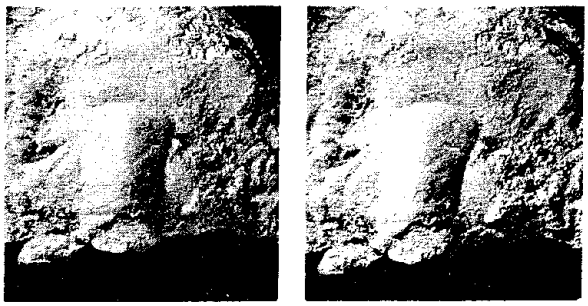

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and Delta Carbonate, Inc., is greatly appreciated. Our thanks to J. E. Repetski and J. T. Dutro of the U.S.G.S. for help in locating suitable sections, processing samples for phosphatic fossils, and assisting in a variety of other ways as we worked to become familiar with the faunas and rock units of the area. A. R. Palmer and A. J. Rowell shared freely of their knowledge of Lower and Middle Cambrian inarticulate brachiopods and trilobites. Expenses were defrayed by a grant from the Pennsylvania Geological Survey. We acknowledge the donors of the Petroleum Research Fund, administered by the American Chemical Society, for partial support of this research. We also thank J. A. Dembosky, P. G. Martini, and K. D. Martini for assistance in preparation of the manuscript.

Figure IX-4 (facing page). Stereographs of trilobites form the Greenmount Member of the Kinzers Formation in the Delta Carbonate quarry (STOP11) near York, Pa. Species characteristic of the "upper fauna" (from sample horizon 2-1, the uppermost 1 to 2 feet of the member) are shown in stereographs 1 to 3, 9, and 10. Characteristic species of the "lower fauna," recovered from the lower two-thirds of the member (sample horizons 2-3, 2-3a, and 2-5), are illustrated in stereographs 4 to 6. "*Prozacanthoides*" *vigginicus* (stereographs 7 and 8), the most abundant species in the member, was recovered from all productive horizons. The generic name is place in quaotation marks because of taxonomic problems. Although originally assigned to *Prozacanthoides*, it clearly is not congeneric with the type species of that genus. Additional study is needed before it can be appropriately reassigned. All specimens are presently reposited in the paleontological collections at Indiana University of Pennsylvania (IUP). Magnification, IUP collectin number, and sample horizons are provided for each figured specimen. All photographs are dorsal views. Some specimens (e.g., 6 and 10) are very slightly deformed.

1 - 2: *Bonnia occipitalis* Rasetti, 1948. 1: testate cranidium, X3.5, IUP 1001, Hor. 2-1. 2: large, partially exfoliated pygidium, X1.5, IUP 1002, Hor. 2-1.

3: *Protypus marginatus* Rasetti, 1948. partially exfoliated cranidium, X3, IUP 1003, Hor. 2-1.

4: *Periomella yorkensis* Resser, 1938. partially exfoliated cranidium, X1.8, IUP 1004, Hor. 2-5.

5 - 6: *Pagetides leiopygus* Rasetti, 1945. 5: fragmentary cranidium, X7.5, IUP 1005, Hor. 2-5. 6: testate pygidium, X11, IUP 1006, Hor. 2-5.

7 - 8: "*Prozacanthoides*" *virginicus* Resser, 1938. 7: exfoliated cranidium, X4, IUP 1007, Hor. 2-3a. 8: exfoliated pygidium, X4, IUP 1008, Hor. 2-5.

9: Genus and species undetermined, exfoliated fragmentary cranidium, X1.8, IUP 1009, Hor. 2-1.

10: *Bicella bicensis* (Resser, 1938). testate cranidium, X4.4, IUP 1010, Hor. 2-1.

X. MINERAL RESOURCES OF YORK COUNTY, PENNSYLVANIA

David Hopkins
The J. E. Baker Company

G. Robert Ganis
Tethys Consultants, Inc.

HISTORY

From the earliest days of European settlement until the present time, the geology and mineral resources of York County have played a critical role in the development of the region. The rumored mineral deposits of the area west of the Susquehanna River spurred the first survey of part of the area that was to later become York County. This survey was made on April 10th and 11th, 1722, by the authority of Governor Keith. The area surveyed was at first called Kieth's Mine Tract. Apparently, Governor Keith was trying to find the area of a rumored copper mine (Prowell, 1907).

Many early homes were built of locally derived stone. Also, clays weathered from carbonate rocks were used to manufacture building bricks. Many farms underlain by carbonate rocks had small quarries and kilns where framers would burn lime for agricultural uses and whitewash. Beginning in the mid 1880s, both limonitic and magnetic ore deposits were developed to supply regional iron furnaces (Prowell, 1907). Many other quarries, both large and small, were opened for various products, e.g., slate, building stone, sand and gravel (both residual and alluvial), aggregate, and lime.

MINERAL PRODUCTS

From these humble beginnings, the greater York area has become one of the largest and most diversified mineral production centers in Pennsylvania. This expansion in the mineral industry has occurred despite rapidly expanding residential and industrial encroachments, adverse zoning regulations, highly restrictive mining laws, and an increasingly hostile public opinion concerning the industry.

The mineral industries of York County currently produce a wide variety of products. Some of these products have common yet important applications, such as construction aggregate and agricultural stone. Other minerals are used to produce cement. York County is the only location in the United States where a refractory grain suitable for dolomite bricks is manufactured. These bricks find widespread use in the steel and cement industries. York County is one of two sites in the United States where white cement is produced, and the only area in Pennsylvania where a whiting material is produced. Whiting is a white form of calcium carbonate used a filler material. Other products manufactured

from the mineral resources of the York area include: face brick, fluxstone, glass stone, poultry grit, mineral fillers (other than whiting), acid neutralization stone, flue gas desulfurization stone, and landfill clay (Berkheiser and others, 1985; DER, 1989).

GEOLOGY

By far, the largest mineral production in the area comes from the Lower Cambrian carbonate formations. The largest volume is produced from the middle limestone member of the Kinzers Formation. The Vintage and Ledger Formations rank second and third in volume production (DER, 1989). Other formations which are currently used for mineral production are: the Chickies Quartzite, Antietam Phyllite/Quartzite, and Harpers Phyllite, all Lower Cambrian, and the Triassic New Oxford Formation. In the past, extensive quarries were developed in the Peach Bottom Slate (age uncertain) in the extreme southeastern corner of the county.

ECONOMIC IMPORTANCE

A total of 11 companies produce an estimated 5.4 million tons per year of material from 13 separate open pits and one underground mine (Berkheiser and others, 1985). These companies employ approximately 820 workers and have estimated combined total finished-product sales of $130 million per year (DER, 1989). This excludes sales of white cement which also is manufactured mainly from locally-produced materials. If white cement sales were included, the total value would be substantially higher.

The full economic impact of the local mining industry typically goes unnoticed and unappreciated by nearly everyone, despite the fact that everyone uses these mineral products directly or indirectly every day. The figure given above clearly indicates the importance that a local mining industry can have on a local and state economy.

LIST OF MINERAL PRODUCERS IN YORK COUNTY

The locations of the 13 quarries listed here are shown in Figure X-1.

1. Name: The J. E. Baker Company
   Rock type: Dolomite, limestone, clay
   Formation: Ledger
   Uses: Refractory dolomite, agricultural stone, mineral fillers, fluxstone, fluidized bed stone, construction aggregate, clay
   Comments: The J. E. Baker Company is the only manufacturer in the United States of refractory grain suitable for dolomite bricks. Worldwide exports account for approximately 25% of their sales.

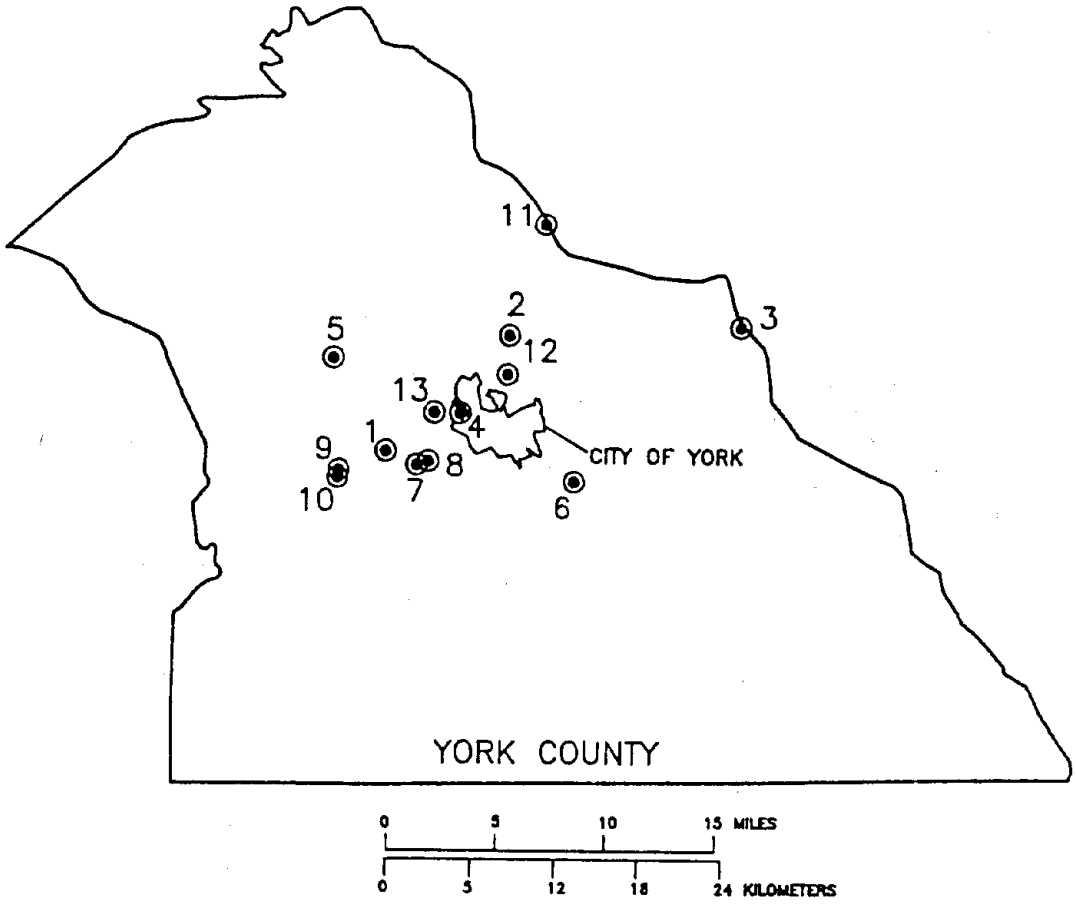

Figure X-1. Locations of quarries in York County, Pennsylvania.

If time permits, the Field Conference will drive through this quarry after STOP 10.

2. Name: Codorus Stone and Supply Co., Inc.
   Rock type: Limestone/dolomite
   Formation: Vintage
   Uses: Various types of construction aggregate
   Comments: Quarry operations are subcontracted to the General Crushed Stone Company.

3. Name: County Line Quarry, Inc.
   Rock type: Dolomite, quartzite, phyllite
   Formations: Vintage, Antietam, Harpers
   Uses: Various types of construction aggregates
   Comments: This quarry recorded the largest production of any quarry in York County in 1989: over 1.25 million tons. County Line Quarry recently purchased Neuman's Quarry (Chickies Quartzite); this is operated as York Silica Sand, Inc.

4. Name: Delta Carbonate, Inc.
   Rock type: Limestone/dolomite
   Formations: Kinzers and Ledger

Uses: Various types of construction aggregate, agricultural stone, whiting and other fillers

Comments: This operation has undergone several ownership changes in the last few years. Names that previously have referred to this site include: Bestone, Inc. and York Stone and Supply Company. Delta Carbonate operates two pits at this location. This site also has an underground mine for whiting material that currently is inactive. Construction aggregate is produced from this site by York Building Products Company, Inc. under a long-term agreement. Delta Carbonate is a subsidiary of Millington Quarry, Inc., of New Jersey. This quarry will be visited as STOP 11 of the Field Conference.

5. Name: Glen-Gery Corporation
Rock type: Shale
Formation: New Oxford
Use: Brick manufacture
Comments: Glen-Gery operates two pits in York County: one near Dover in the Triassic New Oxford Fm., and one south of York in the Cambrian Harpers Phyllite. Glen-Gery also utilizes some local subsoils and clays.

6. Name: Glen-Gery Corporation
Rock type: Phyllite
Formation: Harpers
Use: Brick manufacture

7. Name: Omya, Inc.
Rock type: Limestone
Formation: Kinzers
Uses: Whiting, other fillers
Comments: Formerly known as White Pigment Corp., this operation currently is under option to a group in which the principals of Millington Quarry, Inc. have an interest.

8. Name: Penroc, Inc.
Rock type: Limestone
Formation: Kinzers
Uses: Whiting, other fillers, agricultural stone, construction aggregate
Comments: Penroc is the former Gold Bond Building Products division of National Gypsum. This operation recently was purchased by a group in which the principals of Millington Quarry, Inc. also have an interest. Penroc operates two pits: the Consolidated Quarry, located at the plant site, and the Ensminger Quarry, located approximately 1.25 miles west of the plant, adjacent to the Omya Quarry.

9. Name: Thomasville Stone and Lime Co. (division of Medusa Cement Co.)
   Rock type: Limestone/dolomite
   Formation: Kinzers
   Uses: Whiting, other fillers, agricultural stone, agricultural aggregate, cement stone, fluxstone, glass stone, poultry grit, acid neutralization stone
   Comments: Thomasville Stone and Lime currently operates the only underground mine in York County. This operation supplies white limestone to the Lehigh Portland Cement Co. in West York for the manufacture of white cement.

10. Name: York Building Products Co.
   Rock type: Limestone/dolomite
   Formation: Vintage
   Uses: Various types of construction aggregate, clay
   Comments: This is the same company that has a long-term agreement to produce aggregate from the Delta Carbonate Quarry. This operation is located adjacent to Thomasville Stone and Lime Co.

11. Name: Waste Management, Inc.
   Rock type: Residual clay
   Formation: Ledger
   Use: Landfill clay
   Comments: Waste Management has recently begun development of a clay pit just north of Saginaw near the Susquehanna River. Waste Management is extracting a residual clay that has developed from weathering of the Ledger Dolomite. This area previously was quarried for dolomite by the J. E. Baker Co. in the 1940s.

12. Name: York Silica Sand, Inc.
   Rock type: Quartzite
   Formation: Chickies
   Uses: Construction aggregate, brick facing
   Comments: This operation recently was purchased and reactivated by County Line Quarry, Inc. It formerly was owned by York Stone and Supply Co.

13. Name: West Gate Quarry
   Rock type: Limestone/dolomite
   Formation: Kinzers (within present pit); Ledger (not yet developed
   Uses: Construction aggregate
   Comments: This operation was formerly operated by Medusa Cement and produced material for their white cement plant. The operation is currently being held as officially active, although the pit is water-filled and its production is minimal. The quarry is owned by Millington Quarry, Inc.

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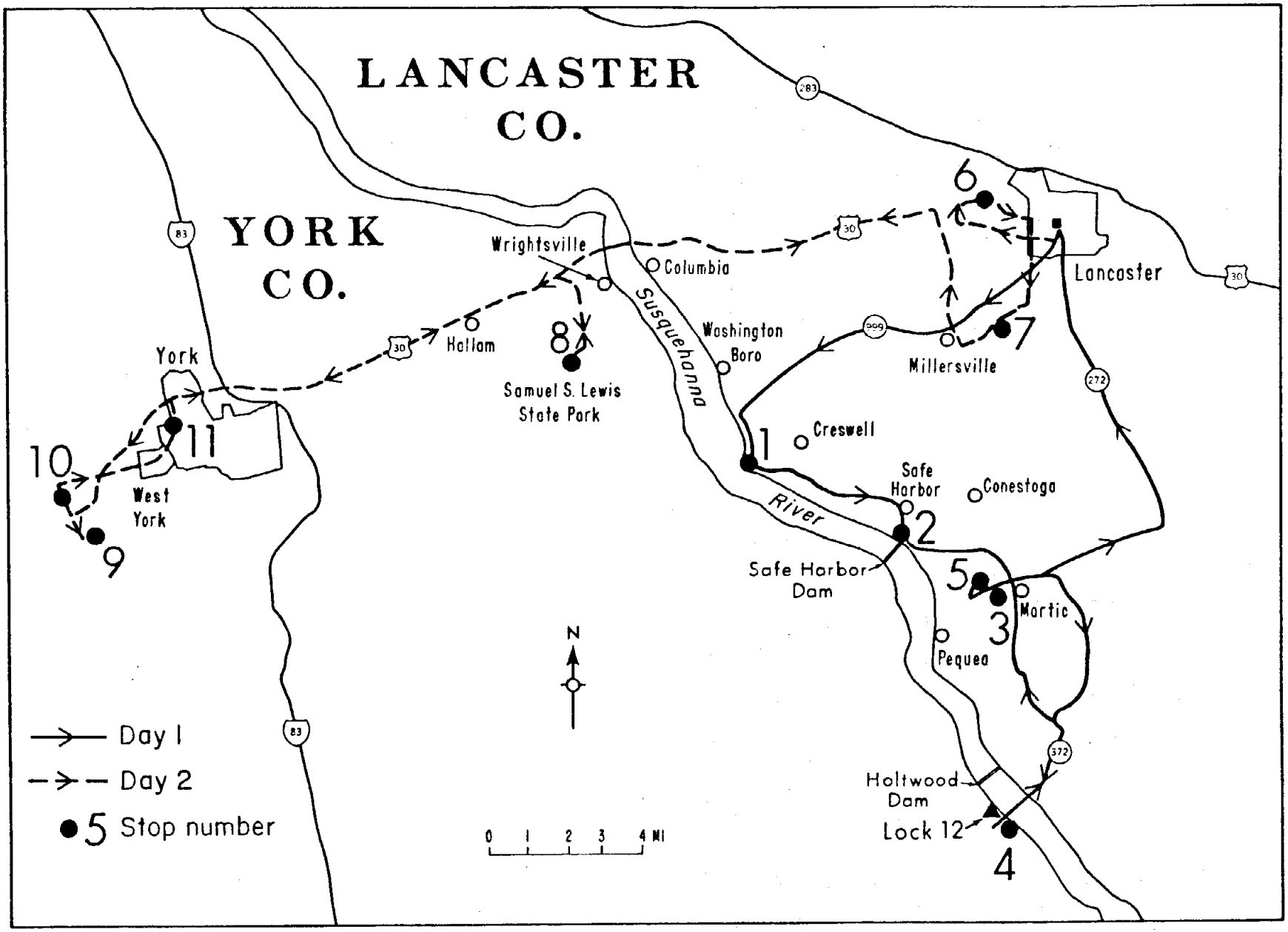


Route of the 55th Annual Field Conference of Pa. Geologists

ROAD LOG--DAY 1

Incidental Notes by
William M. Jordan
Millersville University

Mileage Inc	Cum	
0.0	0.0	START. Leave from Chestnut Street entrance of Brunswick Hotel.
0.1	0.1	TURN RIGHT and procede one block south on Duke Street. Lancaster City Hall on right. Lancaster was laid out and established as the seat of Lancaster County in 1730. On September 27, 1777 the Continental Congress, in retreat westward from Philadelphia, met in Lancaster for one session. From 1799 until 1812 Lancaster served as the capital of Pennsylvania. It was incorporated as Lancaster City in 1818. The extension of many Lancaster streets beyond the relatively small 18th century "core" did not occur until the period 1870-1900. St. James Episcopal Church, founded in 1744, on left. Buried in the churchyard are James Ross, signer of the Declaration of Independence, and Edward Hand, friend and Adjutant General to George Washington.
0.1	0.2	TURN RIGHT onto Orange Street. Procede west; the new Lancaster County Court House is on the left. The older court house building behind it, built in 1852 in the Roman Revival style, faces East King Street.
0.2	0.4	After crossing North Queen Street, the Central Market (Lancaster Farmers' Market) built in the Romanesque Revival style in 1889, is one-half block to the left on North Market Street. CONTINUE west on Orange Street.
0.3	0.7	TURN LEFT onto Charlotte Street.
0.1	0.8	BEAR RIGHT onto Manor Street, procede toward southwest.
0.7	1.5	Cross intersection of Manor Street with West End Avenue (on right) and Hershey Avenue (on left). PROCEDE STRAIGHT AHEAD toward southwest on continuation of Manor Street which is now called the Millersville Pike.
1.2	2.7	Cross intersection of Millersville Pike with Millersville Road (PA 741). Continue straight ahead, Millersville Pike is now called Manor Avenue.
0.3	3.0	Entering "downtown" Millersville.
0.4	3.4	BEAR RIGHT onto Blue Rock Road (PA 999) at the Getty convenience store at the intersection with George Street. The campus of Millersville University is located one mile to the left (south) on George Street. Millersville University, founded in 1855 as the Lancaster County Normal School, is the oldest component of the 14-unit State System of Higher Education.
0.9	4.3	Cross the Little Conestoga Creek.
0.2	4.5	Now crossing outcrop of the Safe Harbor (Rock Hill) Triassic-Jurassic diabase dike.
2.6	7.1	Cross Central Manor Road at intersection known as "Central Manor;" continue straight ahead on Blue Rock Road (PA 999). Central Manor was to have been the middle of a city, as large as Philadelphia, planned by

William Penn. On his orders, an area of 16,000 acres was surveyed in 1717-18 and designated as "Conestoga Manor," to be used by William Penn and his heirs and assigns forever. Eventually Penn's Manor was divided and sold as farmland to mainly Quaker and Mennonite settlers before the Revolution.

0.3 7.4 BEAR LEFT, at Central Manor Church, continuing on Blue Rock Road (note that PA 999 bears off to right as Washington Boro Road).

1.9 9.3 Passing Lancaster Area Sewer Authority sewage treatment plant on right.

0.4 9.7 End of Blue Rock Road at the Susquehanna River. The "Blue Rock" (an outcrop of Conestoga Formation, now hidden by the Conrail tracks) was a landmark at the terminus of a ferry that crossed the Susquehanna to the York County shore. TURN LEFT onto River Road, procede toward the south, crossing Witmer Run. This location is the western end of the former "Great Minqua Path" used by the Minqua (or Suquehannock) Indians in the mid-17th century to carry beaver skins east to the white settlements on the Delaware. The Susquehannocks were considered to be the most warlike of the Indians living along the Susquehanna. Major archeological excavations of Susquehannock settlements have been made in this area. A large village (the Schultz Site) was located on a knoll just south and east of where River Road crosses Witmer Run. This village covered 5 acres and housed a population of 800 to 1,000. After moving into the area from the north about 1575 (displacing the earlier "Shenk's Ferry Indians" who occupied the area from about 1250 to 1550), the Susquehannocks occupied several sites on both sides of the river, relocating their village every 20 to 25 years. Captain John Smith made contact with the Susquehannocks in 1608. Their power was broken in 1675 as the result of a long war with the Iroquois. A remnant population called the Conestoga Indians persisted in the area, although decimated by European diseases, until exterminated by the "Paxtang Boys" from the Harrisburg area in massacres that occurred on December 14 and 27, 1763. The last indians killed, mainly old men, women, and children, had been housed in the old Lancaster Jail in a futile attempt to offer them protection.

0.8 10.5 View of wooded, north-facing slope of Turkey Hill ahead. The base of this slope, along Wissler's Run, is the contact of the Wissahickon Formation (albite-chlorite schist) with the impure carbonates (phyllitic marble) of the Conestoga Formation. This contact is the famous "Martic Line."

0.4 10.9 CAREFULLY PULL OFF THE ROAD TO THE LEFT beyond Wissler's Run. Park on the approximate position of the Martic Line along the edge of field at the base of Turkey Hill. CAREFULLY cross road.

WHEN CROSSING ROAD, BEWARE OF RAPIDLY MOVING TRUCKS THAT DESCEND THE HILL! NORTHBOUND DESCENDING TRAFFIC ENTERS A BLIND CURVE AT THIS SPOT. USE EXTRA CAUTION! Procede on foot west to the Conrail railroad tracks that parallel Susquehanna River. Walk south along tracks to STOP 1.

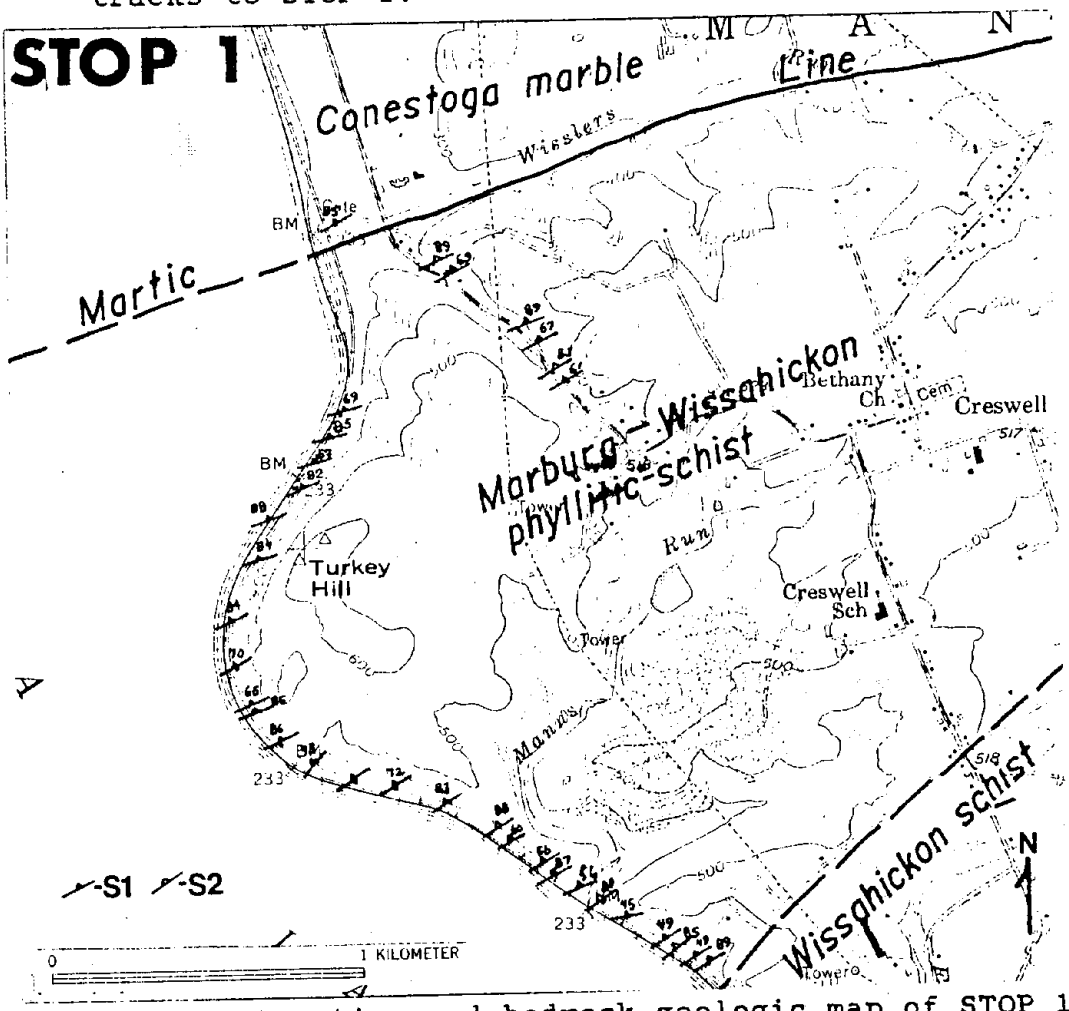


Figure S1-1. Location and bedrock geologic map of STOP 1

STOP 1. THE WISSAHICKON SCHIST AT TURKEY HILL

Leader: Dave Valentino

THE MARTIC LINE

The contact between marble of the Conestoga Formation and Marburg-Wissahickon phyllitic schist projects through the area on the north side of Turkey Hill (Figure S1-1). Although the actual lithologic contact, traditionally known as the Martic Line, is not exposed at this locality, marble (to the north) and phyllitic schist (to the south) crops out over a distance of 150 meters.

The southernmost occurrence of the Conestoga Formation in this area consists of small exposures of gray phyllitic marble on

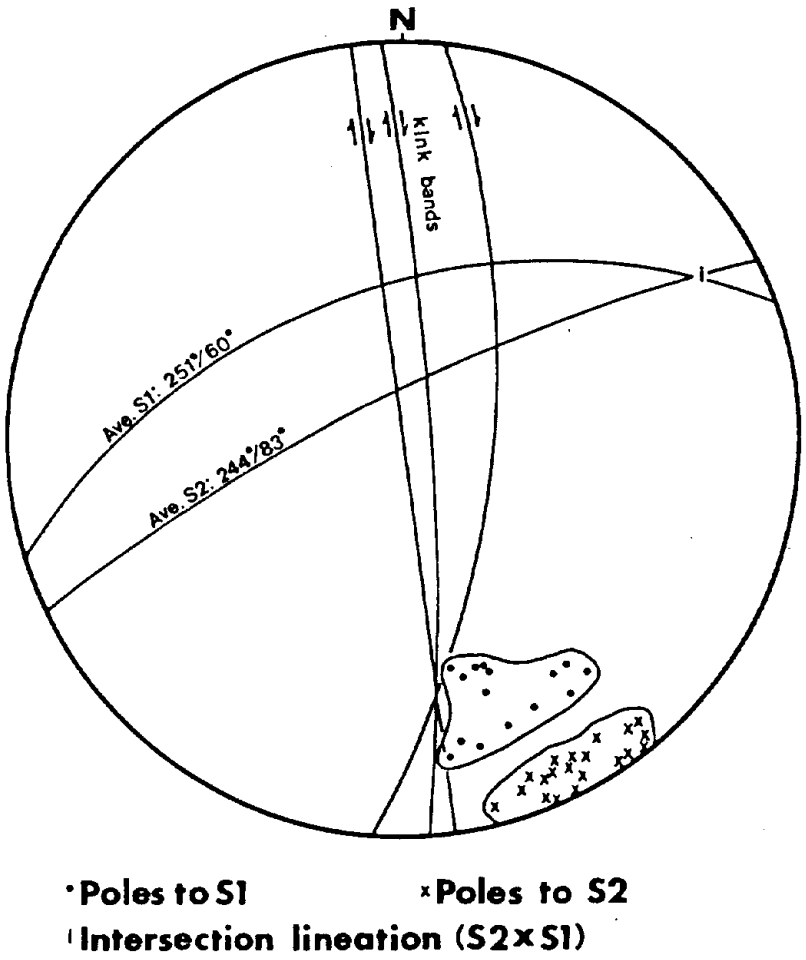


Figure S1-2. Schmidt net stereographic projection for various structural features observed at STOP 1.

the south side and in the bed of the dirt road that traverses along the north edge of Turkey Hill. Although the marble is deeply weathered and locally slumped, steeply northwest dipping internal structures can be observed easily.

The Marburg-Wissahickon phyllitic schist crops out as exposures scattered through the wooded area parallel to the dirt road, but is best exposed along the railroad tracks about 600 meters to the south. Where the phyllitic schist is well exposed it bears fine grained muscovite, chlorite, quartz and less plagioclase.

STRUCTURES

The rocks at Turkey Hill have been subjected to two phases of deformation: 1) phase one, characterized by coarsely crystalline, moderately to steeply dipping schistosity defined by parallel metamorphic minerals in both the Conestoga Formation and the rocks of the Wissahickon Group, and 2) phase two, characterized by steeply dipping crenulation cleavage and new

penetrative schistosity of variable strength, but most strongly developed in the Turkey Hill area.

Phase One

The schistosity (S1) is penetrative on the outcrop, hand sample and thin-section scale. This penetrative planar fabric is defined by parallel alignment of phyllosilicates, planar aggregates of plagioclase and quartz (millimeters thick) and discontinuous layers of vein quartz (centimeters thick). When viewed in thin-section the S1 schistosity is defined by elongation or parallel alignment of nearly every crystal in the rock. The primary schistosity (S1) has the general orientation of 060° to 080° strike and 65°-75° dip to the northwest (Figure S1-2).

F1 Isoclinal Folds

Interfolial isoclinal flow-folds, irregular folds associated with vein quartz masses, and microscopic structures all show evidence of ductility during formation. The axial planes of isoclinally-folded vein quartz layers (centimeters thick) are parallel to the schistosity. These isoclinal flow-folds are usually between 2 and 10 centimeters in amplitude, although many folds smaller and larger in size can be observed. The hinge areas of the folds are thickened and the limbs have been attenuated. The limbs of these folds are always parallel to the S1 schistosity, suggesting ductile flow parallel to S1 schistosity, and commonly are rootless. The orientation of the hinge axes are very difficult to determine because the steep faces of the rock exposures usually allow for only a two-dimensional view of the folds. However, when measurable, the hinge axes generally are subhorizontally oriented or parallel to the strike of S1 (Figure S1-2). Freedman and others (1964) proposed that the isoclinal folds and regional S1 developed during the emplacement of an a nappe structure with the transport direction to the northwest, perpendicular to the hinge axes of the folds and parallel to the S1 schistosity. The magnitude of displacement is unknown.

Phase Two

The second deformation phase (D2) is characterized by a penetrative cross cutting schistosity in the northernmost Marburg-Wissahickon lithology of the Turkey Hill area. The S2 in the Marburg-Wissahickon strikes between 070° and 080°, and dips steeply between 75° and 90° to the northwest (Figure S1-2). This S2 schistosity is defined by the parallel alignment of second-generation muscovite and chlorite and planar aggregates of fine grained quartz. South of Turkey Hill (about 2 kilometers) the S2 appears as moderately to weakly developed crenulation cleavage with new growth of retrograde chlorite in the hinge and muscovite on the limbs of the crenulations. Near Safe Harbor Dam (STOP 2) the S2 schistosity is not present. The orientations of the crenulation cleavage and the penetrative S2 foliation are

identical and retrograde metamorphic minerals defining the S-surfaces are the same (second-generation muscovite and chlorite), suggesting synchronous development. Associated mineral extension lineations (L2) are defined by weathered elongate pyrite crystals, and by quartz fiber pressure shadows. These lineations plunge between 0° and 10° SW and trend approximately 250° (Figure S1-2).

Relative Timing

The relative timing between S1 foliation and S2 can be determined easily in the field. The intersection of the S1 and S2 schistosities defines a lineation which is diamond shaped in profile view (Figure S1-3). Truncation of the S1 schistosity at the S2 schistosity surface clearly shows the relative timing of deformation.

EVIDENCE FOR NNW-SSE DIRECTED COMPRESSION

Micro-slip-folded S1, defining F2 folds, has been observed where S2 is weakly developed (Figure S1-4). The sense of motion across the weak S2, determined by the curviture of the trace of S1 schistosity near the S2 boundary, appears to be inconsistently up and down dip, as best seen in vertical rock surfaces perpendicular to S2. The apparent up and down offset across the S2 schistosity surfaces is the result of crenulation of the pre-existing moderately to shallowly dipping fabric (S1). Often the S2 schistosity appears to have recrystallized over the S1 schistosity with little or no disturbance of microlithons defining S1. This texture and the upright F2 folds are indicative of a strong component of compression perpendicular to the S2 schistosity. Since the S2 schistosity is steeply dipping to the northwest, a NNW-SSE subhorizontally oriented compressive stress seems to be indicated.

EVIDENCE FOR DEXTRAL STRIKE-SLIP DEFORMATION

Near Turkey Hill abundant asymmetric quartz pressure fringes occur on pyrite porphyroclasts (see Figure III-11b) that indicate consistent strike-slip dextral motion (pyrite type: Ramsay, 1983). On a large block of rock located along the railroad track (about 800 meters from the north end of the outcrop) these microstructures can be observed in the field. While the shear sense cannot be determined from the structures in this block because it has fallen from the outcrop, numerous oriented samples collected from the outcrop along the railroad have provided excellent microstructures for kinematic analysis.

LEAVE parking area, procede up hill (to south) on River Road.

0.8 11.7 Turkey Hill Dairy complex on right at the top of the hill. Operations of the Lancaster County Solid Waste Management Authority landfill are visible, behind the dairy, to the right.

0.5 12.2 TURN RIGHT at intersection, continuing on River Road. The community of Creswell is to the left.

0.1 12.3 Entrance to Lancaster County Solid Waste Management Authority landfill on the right.

0.6 12.9 BEAR LEFT, continue on River Road and pass through community of Highville. The high ground followed by River Road is on the Wissahickon Formation, while the low ground visible north and east (to left) is underlain by Conestoga Formation on the far side of the Martic Line which follows the northeastern base of the high ground.

2.0 14.9 Beyond Pittsburgh Hill Road, River Road crosses the Martic Line while descending into the valley of another Witmer Run, this one a tributary of the Conestoga River. Continue on River Road.

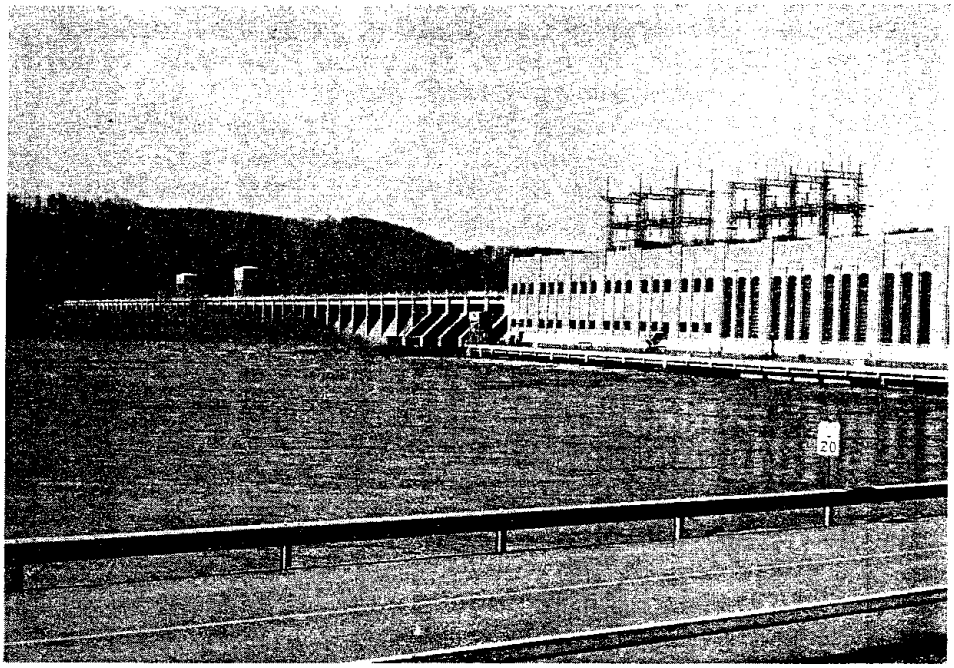
0.9 15.8 Cross the Conestoga River. Visible upstream, to the left, at the far end of the bridge, are the partially preserved remains of Lock 8 of the Conestoga Navigation Company canal. This waterway extended from the Susquehanna River to the city of Lancaster as a 18-mile long slackwater canal system. It consisted of nine dams created to impound navigation pools on the Conestoga River, with adjacent locks to bypass the dams. Lock 8 had a 100 foot length, 22 foot width, and a 6 foot lift. Lock 9, at the Susquehanna River but now gone, had a lift of 8 feet. The canal operated only from 1828 until 1837 due to recurrent ice and flood water damage.

0.1 15.9 TURN RIGHT at end of the Conestoga River bridge. Continue south on River Road. Pennsylvania Power and Light Company Conestoga River (Safe Harbor) Park on right.

0.3 16.2 Main Street, leading from the village of Conestoga, enters from left. Continue straight ahead (south) on River Road.

0.3 16.5 Outcrop of Vintage Dolomite on the left. In the 19th century this area was the site of an extensive iron industry that included, starting in 1848, the rolling of iron railroad rails. This mill used the output of Safe Harbor Anthracite Furnace which had a capacity of 12,000 tons per year. Limonite ore was obtained from a deposit located on the Martic Line at the intersection of Pittsburgh Valley and Pittsburgh Hill Roads in the valley of Witmer Run. Later, in 1881, magnetite orebecame available from the Antietem Formation outcropping on the property of the Pequea Magnetic Iron Mining Company along Pequea Creek several miles to the east. Under various owners (Standard Iron Mining and

S1
S2



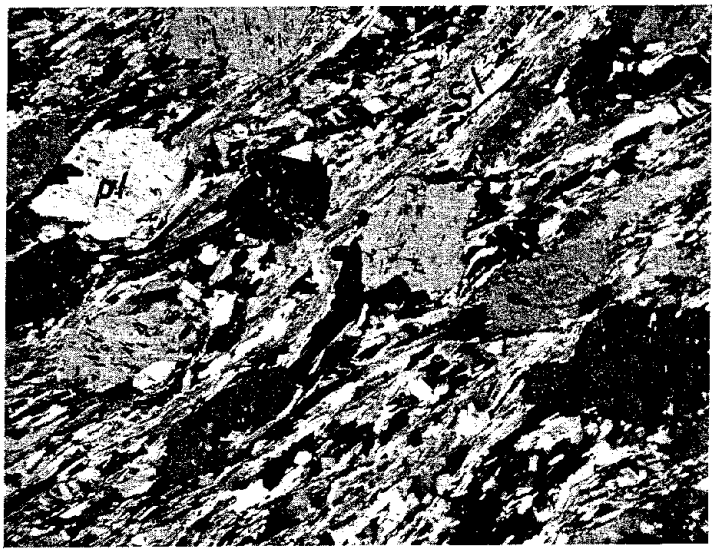

S1
P1

Figure S1-3 (top left, facing page). Photomicrograph of S2 schistosity cross-cutting the S1 penetrative schistosity in the Marburg-Wissahickon phyllitic schist; field of view is 2.5 mm.

Figure S1-4 (top right, facing page). Photomicrograph of F2 folds from the Marburg-Wissahickon phyllitic schist; field of view is 2.5 mm.

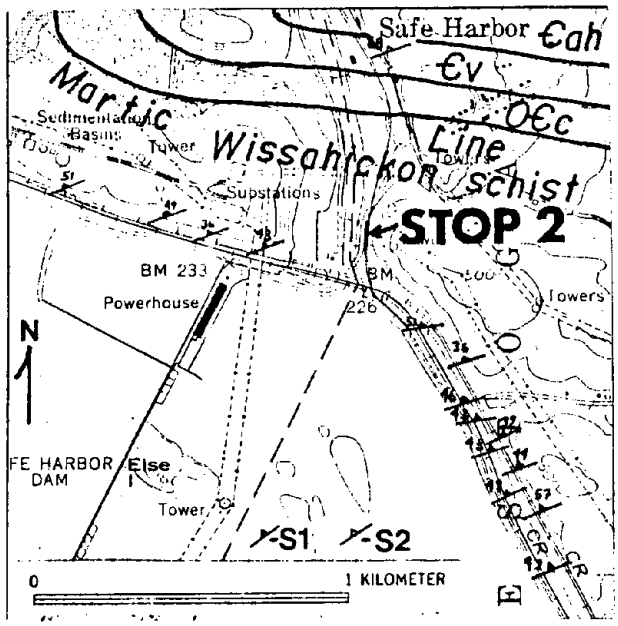
Figure S2-2 (center, facing page). Photograph of the Safe Harbor hydroelectric plant as viewed from the east shore of the Susquehanna River.

Figure S2-3 (bottom, facing page). Photomicrograph of coarse crystalline Wissahickon muscovite-chlorite-biotite-plagioclase schist from the Safe Harbor area; field of view is 6.0 mm.

Furnace Company, and later the Safe Harbor Iron and Steel Company) that mine operated sporatically until 1913.0.3 16.8 BEAR RIGHT onto access road leading to the Safe Harbor Hydroelectric Plant. Outcrop of Wissahickon Formation (STOP 2) on left.

0.3 17.1 Pass beneath overhead Conrail railroad tracks while crossing road bridge over the Conestoga River. The lower of the two railroad track levels is that of the former Columbia and Port Deposit Railroad (later Pennsylvania Railroad), while the upper high viaduct carries the "low grade line" of the Pennsylvania Railroad completed in 1906. PULL INTO PARKING LOT ON RIGHT. The Safe Harbor Dam, built by the Safe Harbor Water Power Corporation in 1931, impounds the Susquehanna River as 11.5 square mile Lake Clarke, with a normal pool elevation of 228 feet. Lake Clarke extends upriver beyond Turkey Hill. Below the dam, downstream, Lake Aldred is impounded behind the Holtwood Dam and has a pool elevation of 169 feet. At this location, carved on rock islands in the Susquehanna, were numerous prehistoric indian petroglyphs. The majority of these rock carvings are now submerged beneath the waters of Lake Clarke, although some remain below the dam on Big and Little Indian Rocks offshore from this parking area. Little Indian Rock is usually submerged, but Big Indian Rock remains visible at all seasons. The petroglyphs on the bedrock islands, although marred by modern carvings dating from 1780 to the 1980s, are now listed on the National Register of Historic Places. Persifor Frazer, Jr. of the Second Geological Survey of Pennsylvania, writing in the late 1800s concerning the destruction of the petroglyphs, stated that, "...in addition to the natural causes of obliteration it is a pity to have to record the vandalism of some of the visitors to the locality who have thought it an excellent practical joke to cut spurious figures alongside and sometimes

over the top of those made by the Indians." Efforts at preservation of the original glyphs include plaster casts made in 1863-1864 by the Linnaean Society of Lancaster County and now at the North Museum on the Franklin and Marshall College campus in Lancaster. At that time more than 80 distinct figures were visible. In 1889 sketches of the petroglyph-covered rocks were made by W. J. Hoffman and in 1930, as the Safe Harbor Dam was being constructed, Donald A. Cadzow of the Pennsylvania Historical Commission recorded the surviving carvings, making casts that are now at the William Penn State Museum in Harrisburg. Cadzow's report on this salvage project was published in 1934.



Firgure S2-1. Location and bedrock geologic map of STOP 2.

STOP 2. WISSAHICKON SCHIST-GNEISS AT SAFE HARBOR

Leader: Dave Valentino

This outcrop is at the entrance to the Safe Harbor Hydroelectric Dam at the mouth of the Conestoga River (Figure S2-1). The Safe Harbor Dam (Figure S2-2) is owned and operated by the Pennsylvania Power and Light Company (PP&L) and the outcrop is in part of the Conestoga Park, maintained for public recreation by PP&L.

LITHOLOGY

The lithology at this locality is muscovite-chlorite-biotite-plagioclase schisto-gneiss. Although this rock is mostly comprised of phyllosilicates, the interlocking of grains of abundant plagioclase and quartz gives it a somewhat gneissic

texture. In addition to the abundant plagioclase porphyroblasts (1-5 mm in diameter), biotite porphyroblasts (2-5 mm in diameter) are common at this locality (Figure S2-3).

STRUCTURES

The primary schistosity is defined by parallel alignment of muscovite and chlorite, and planar aggregates of quartz. Thin veins of quartz are parallel to the primary schistosity as is compositional layering defined by quartz and plagioclase-rich layers alternating with phyllosilicate-rich layers. The compositional layering is folded into isoclines with thickened hinge areas and atenuated limbs that usually are discontinuous (F1 folds). The axial planes are parallel to the primary schistosity described above. The schistosity in the Safe Harbor Dam area strikes 070° and dips 40° to the northwest (Figure S2-4). This locality is situated on the northern limb of the Tucquan Antiform.

The hinge axes of crenulations define a lineation on the schistosity surfaces. These crenulations are associated with the second phase of regional deformation. Cleavage associated with the crenulations is non-existent in most of the outcrop. However, where the cleavage has been observed it is very weakly developed, strikes 080° and is steeply dipping 70°-90° to the northwest (Figure S2-4).

The eastern end of the outcrop contains numerous kink bands (Figure S2-5). Kink bands are zones transecting earlier planar fabric, rotating or reorienting the fabric about some axis that is perpendicular to the direction of motion. The boundaries or walls of the kink band are the boundary planes between un-kinked and kinked portions of the rock. A kink band is in effect a zone of offset or shearing between two non-deforming bodies. Shear sense is determined by the sense of rotation of the fabric within the kink band.

The kink bands range from narrow zones less than a centimeter wide to broad zones up to 15 centimeters wide, with offset generally less than half the width of the kink band. These kink bands generally strike 030°-070° with 60°-80° southeastern dip and normal offset (Figure S2-4). The kink bands are younger than the S2 fabric because both S2 and S1 are deformed by them, in places even by the same kink band.

Earlier work by Freedman et al. (1964) documented all of these structures. The primary schistosity was referred to as S1 and related to large-scale nappe emplacement to the northwest (Freedman et al., 1964) during the Taconian Orogeny (Lapham and Bassett, 1964). The S2 or weakly developed cleavage at this locality was related to the uplift of the Tucquan Antiform (Freedman et al., 1964 and Wise, 1970).

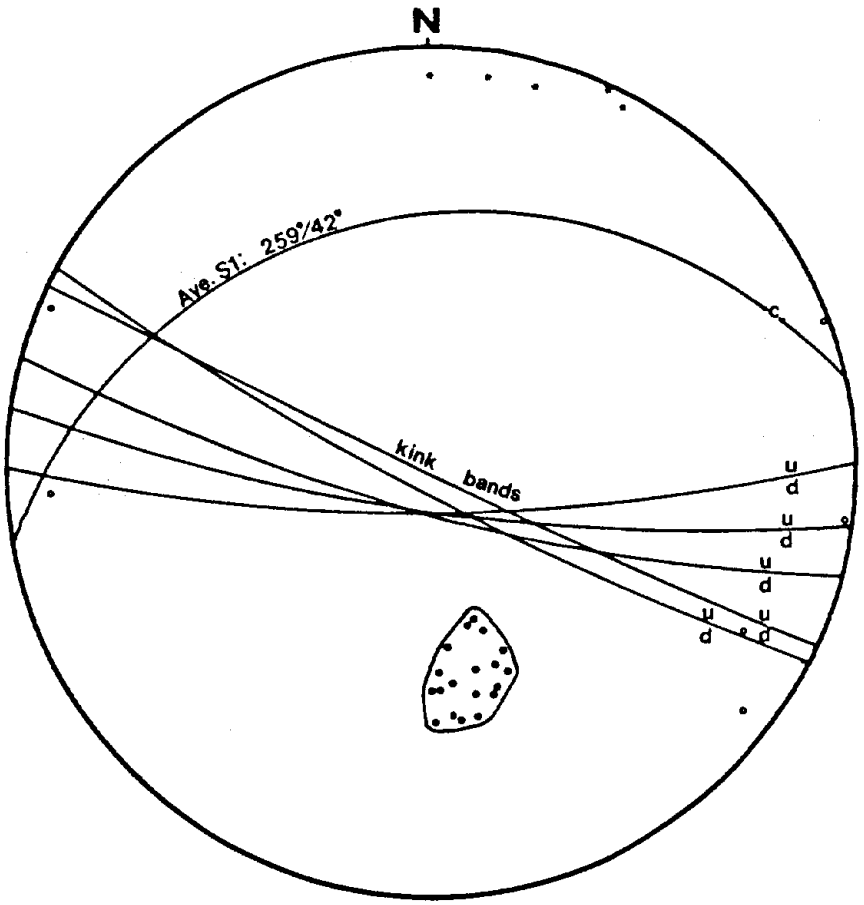


Figure S2-4. Schmidt net stereographic projection of S1 schistosity and kink band data collected at STOP 2. Although brittle faults are not discussed, the poles to possible fault surfaces are represented by the open circles.

METAMORPHISM

The metamorphic mineral assemblage muscovite-chlorite-biotite is indicative of upper greenschist facies. Biotite porphyroblasts (2-5 mm in diameter) can be observed on fresh exposure surfaces (Figure S2-6). Parallel alignment of muscovite and chlorite define the S1 schistosity while the biotite crystals have grown with basal surfaces both parallel and at a high angle to the schistostiy. There do not appear to be any differences among the variously oriented biotite crystals, other than orientation. Biotite is generally pleochroic shades of brown with thin (0.01-0.04 mm) green-brown lamallae within the crystals. Biotite and chlorite also contain abundant minute zircons (0.04 mm diameter) with well-developed pleochroic halos (Figure S2-7).

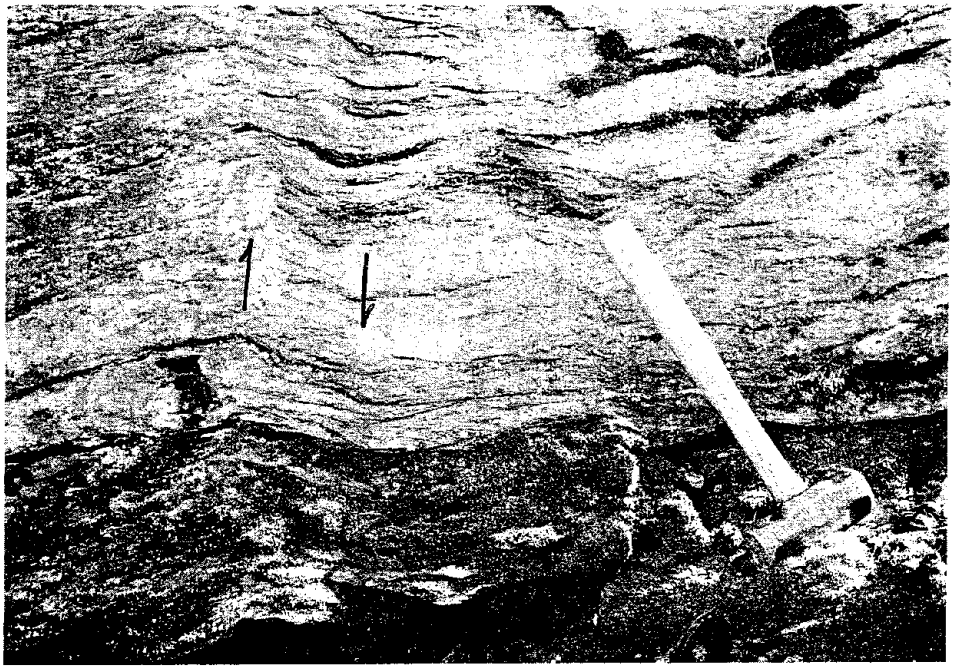
The mapped biotite-garnet isograd is located about 5 km south along the Susquehanna River (Figure S2-8), in the area of Pequea (Faill and Valentino, 1989; Valentino and Faill, 1990). The presence of well-developed biotite indicates that this exposure is within the biotite zone. The chlorite-biotite isograd is located approximately 5.5 kilometers north along the Susquehanna River (see Figure III-1).

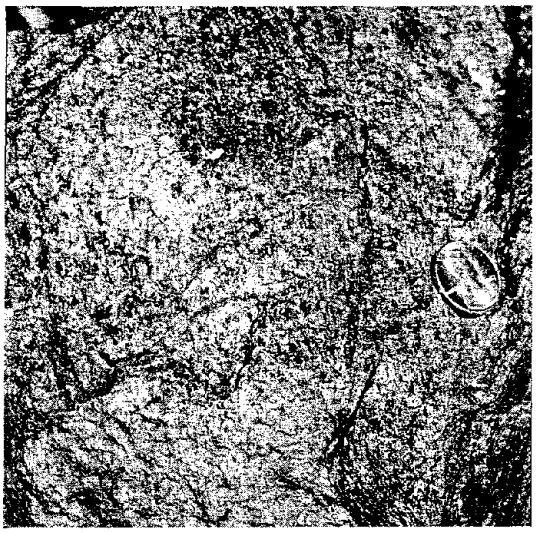
Thin-section analysis of rocks from this locality has revealed a second episode of metamorphism. Muscovite laths (0.5 to 2.0 mm long) have overgrown the S1 schistosity at a high angle. New growth of chlorite at the expense of the primary biotite also has been observed (see Figure III-15c). The assemblage muscovite-chlorite is indicative of lower greenschist facies. This second metamorphic episode is comparatively much less penetrative than the metamorphism that formed the primary minerals in the rock (see Chapter III for a more extensive discussion).

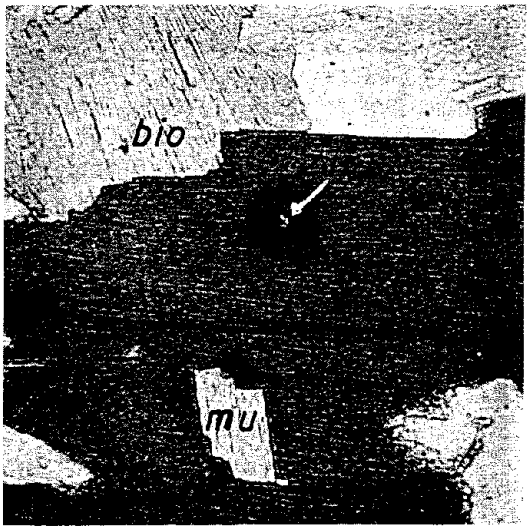
EVIDENCE FOR PRE-D1 DEFORMATION AND METAMORPHISM

Figure S2-8 is a photomicrograph of the schist from the Safe Harbor outcrop. Plagioclase crystals contain abundant inclusions that are aligned and define a micro-foliation within the crystal. The inclusions are zircon, ilmenite, magnetite, sphene, muscovite, chlorite, and (rarely) epidote. The most abundant inclusion mineral is ilmentite. In Figure S2-8 the dominant schistosity, oriented horizontal in this view, is the S1 regional schistosity which is the primary schistosity at the outcrop. Between two bands of S1 schsitosity there is an earlier schistosity preserved which is oriented parallel to the inclusion patterns in the plagioclase. These textures suggest that: 1) plagioclase overgrew an earlier schistosity, now preserved within the crystals by residual ilmenite alignment, 2) this schistosity, present in the rock prior to the penetrative development of S1, contained muscovite, chlorite, magnetite, ilmenite, sphene, zircon and epidote.

The metamorphic mineral assemblage muscovite-chlorite-(epidote) which defines this pre-S1 schistosity is indicative of greenschist facies metamorphism. This assemblage of minerals is lower grade than the primary M1 assemblage that now dominates the rock (muscovite-chlorite-biotite). The primary schistosity in the rock (S1) has been related to the Taconian Orogeny (Freedman et al., 1964; Wise, 1970). Sedimentary layering probably can be ruled out as an origin for the pre-S1 fabric because sedimentary layering usually is widely spaced and defined by compositional variations much thicker than the pre-S1 fabric. This pre-S1 schistosity probably reflects an earlier stage of deformation and metamorphism also associated with the Taconian event.






bio
mu

S1
pre-S1

Figure S2-5 (top, facing page). Photograph of a kink band.

Figure S2-6 (left center, facing page). Photograph of biotite porphyroblasts on the S1 schistosity surface.

Figure S2-7 (right center, facing page). Photomicrograph of biotite with zircon inclusion. Notice the dark halo around the zircon; field of view is 2.5 mm.

Figure S2-8 (bottom, facing page). Photomicrograph of pre-S1 schistosity preserved in a plagioclase crystal and between moderately developed S1 schistosity zones, from the Wissahickon lithology at STOP 2; field of view is 0.8 mm.

		Return to River Road.
0.3	17.4	TURN RIGHT and continue southeast.
1.2	18.6	BEAR RIGHT, continue southeast on River Road.
1.7	20.3	Intersection with River Hill Road (on right) and Pequea Creek Road (on left). Continue southeast on River Road. This route was known as the "Raftman's Path" during the era of downstream-only river navigation, prior to canalization of the lower Susquehanna in the mid-19th century.
0.7	21.0	Intersection with Colemanville Church Road on right. Continue on River Road.
0.2	21.2	Pass under high viaduct of the now abandoned Pennsylvania Railroad "low grade line" as it crosses Pequea Creek.
0.2	21.4	TURN RIGHT at intersection with Pequea Boulevard (PA 324) at Martic Forge. The name "Martic" is derived from the town of Martock in Somerset in the west of England where early settlers, from Hesse-Darmstadt in Germany, assembled before leaving for Pennsylvania in the late 17th century. Martic Forge was the site of the Martic Ironworks, built in 1751, which operated until the end of the Revolutionary War.
0.9	22.3	Intersection with Fox Hollow Road on left. DISEMBARK FOR STOP 3. Walk south on Fox Hollow Road to the Colemanville Covered Bridge. To the right is the site of the "Lower Forge" at Colemanville, consisting of a rolling mill and forge built in 1828. Charcoal iron blooms for boiler plate and the manufacture of nails were produced until 1872. CROSS COVERED BRIDGE, TURN LEFT AND PROCEDE UPSTREAM on orange-blazed Conestoga Trail through State Game Land No. 288 along old trolley bed following east bank of Pequea Creek to buses waiting at Martic Forge.

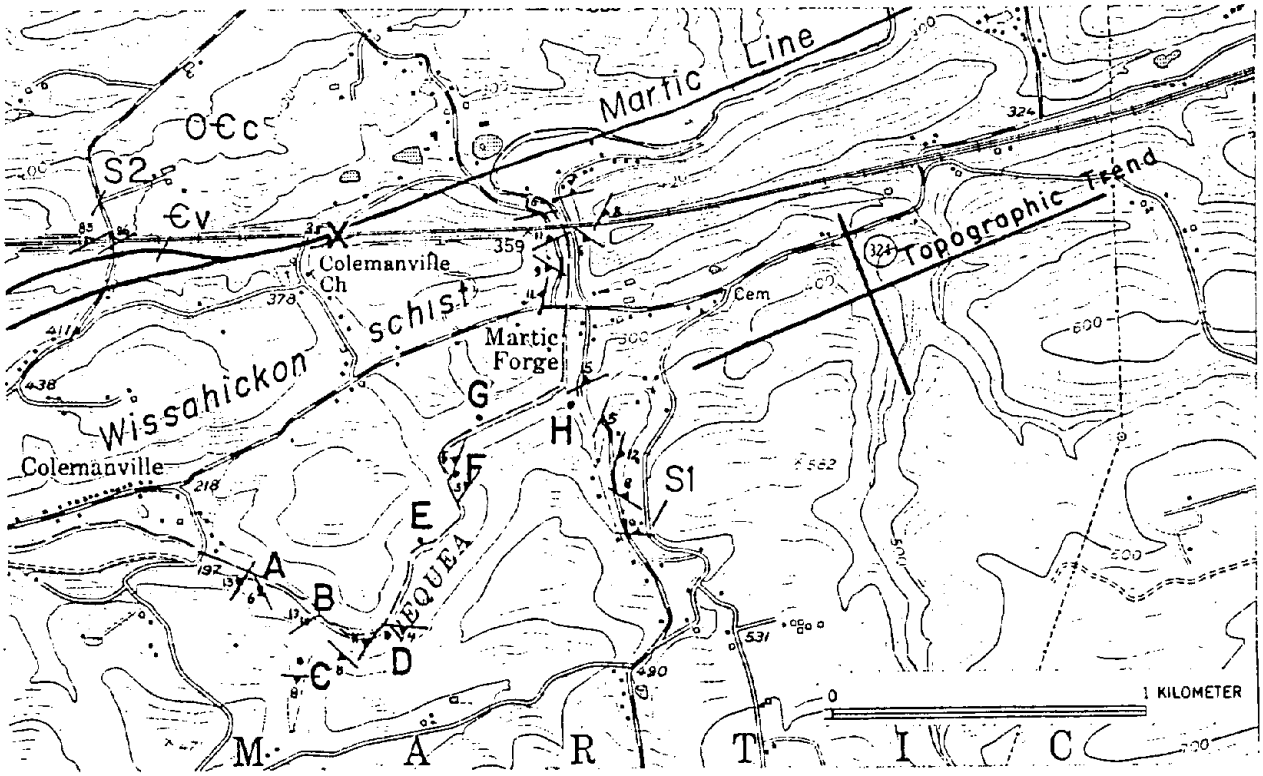


Figure S3-1. Location map for STOP 3, traverse along Pequea Creek. Letters A-H denote specific sites discussed in the stop description. Letter X is the location of STOP 5. S1 and S2 are orientations of schistosity. Oec=Conestoga Fm., €v=Vintage Fm. Map is part of Conestoga 1:24,000 scale topographic map. Heavy line normal to topographic trend line is drainageway trend.

STOP 3. GEOMORPHIC AND STRUCTURAL FEATURES ALONG THE PEQUEA CREEK NEAR SAFE HARBOR

Leaders: Bill Sevon and Dave Valentino

At this stop the group will walk a distance of about 1 mile (Figure S3-1). Most of the trail is on private property. Please treat the landscape with respect: leave only footprints and take only photographs.

GENERAL DESCRIPTION

The trail follows the bed of a former trolley line which ran from Pequea to Martic Forge. The trolley was operational until about 1931. A few hundred feet upstream from the covered bridge are some stone foundations on both sides of the creek. This was the site of a dam built for a small hydroelectric operation. It appears that the dam existed around 1900, but details about this facility were not researched.

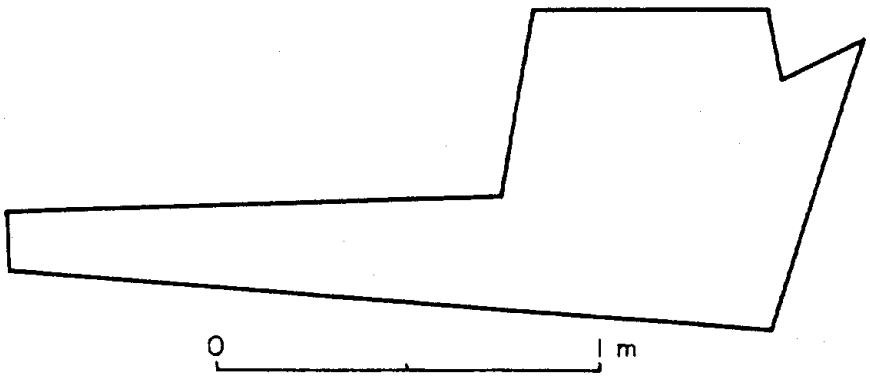


Figure S3-2. Outline of shape of block of Wissahickon schist which is being squeezed out of the outcrop at SITE A.

Pequea Creek is one of the several tributaries to the Susquehanna River in the Holtwood Gorge area which has a convex profile in its lower reaches. Thompson (1988; Chapter V, this guidebook) has argued that these profiles result from erosional disequilibrium which developed during Pleistocene deglaciations when the Susquehanna River carried larger than normal volumes of water and debris and rapidly incised its bed. Tributary streams in the Holtwood area were not capable of eroding their beds at the same rate because they did not have increased water volumes or debris loads, thus creating the disequilibrium condition and the resultant convex stream profile.

The result of this disequilibrium is that Pequea Creek is eroding its bed headward and this is well shown in the traverse area. At Site C there is a rapids at a very narrow constriction of the stream where an erosional knickpoint occurs. The effect of the incision of Pequea Creek on its tributaries is discussed in detail at Site B.

At sites C and D, details of geologic structure can be observed and further insight gained into the tectonic history of the Tucquan Antiform.

SITE A

Site A is an outcrop of Wissahickon Schist that was once quarried for local use. Of interest here is a block of schist which is being squeezed out of the outcrop face. The main smooth face of the outcrop is approximately vertical and has an strike of about 315° (N45°W). The irregularly shaped block (Figure S3-2) has moved as much as 30 cm out from its original position of alignment with the free face. The movement is occurring along planes provided by schistosity and fractures. Two factors are presumed responsible for the movement: pressure and freeze-thaw action. The orientations of the several planes suggest that at least some of them intersect at depth within the rock, thus

creating a wedge-shaped piece. Pressure from the weight of the overlying rock, aided by lubrication of the planes by water andfreeze-thaw action during the winter, presumably is forcing the wedge of rock outwards from the free face.

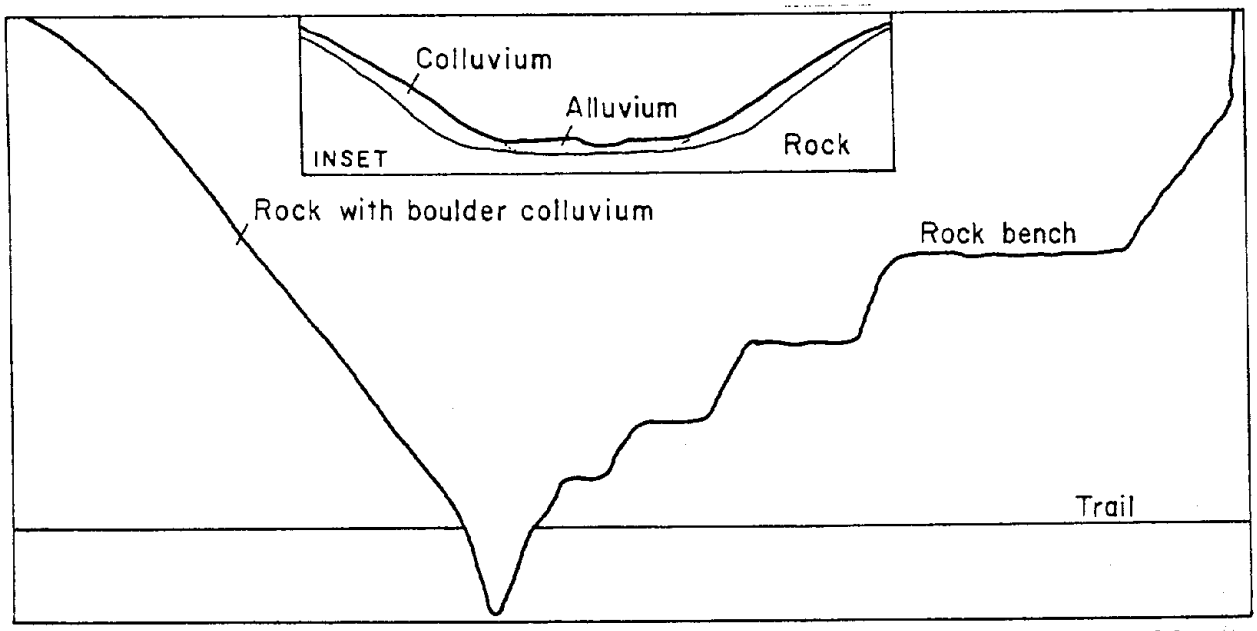


Figure S3-3. Cross section of the shape of the tributary valley at SITE B, the mouth of the tributary. The inset shows the cross section shape of the valley above the major knickpoint at SITE C. The cross sections are not drawn to scale.

SITE B

Site B is at the mouth of a small tributary to Pequea Creek which shows the effect of lowering of the base level of the larger stream. The tributary has a narrow, steep-sided valley in its lower part, a waterfall at the major knickpoint, and a floodplain and low slopes on the valley sides in the upper part where the character of the pre-incision valley is preserved.

Rock on the northwest side of the tributary at its mouth is stepped (Figure S3-3) and this is interpreted to indicate that there were several phases of renewed erosion which cut the valley. The heights of the benches above the stream bed at the trail are 18, 25, 36, and 47 feet. The upper bench is the broadest and is presumed to be the downstream correlative of the valley floor above the waterfall knickpoint.

As you walk up the tributary valley to the waterfall, note the character of the valley. The stream gradient is steep. The valley floor and the valley sides are covered with abundant blocks of schist (coarse-textured colluvium) which broke off, presumably through freeze-thaw action, from rock exposed during the lowering of the stream bed. Most of the blocks appear to be too large for the stream to move except during extreme flood events. The valley slopes have the appearance of stability, but some very slow downslope movement probably occurs. Note that

Figure S3-5. Photograph of the plunge pool pothole outcrop at SITE C. Notice the subhorizontal orientation of the primary schistosity.

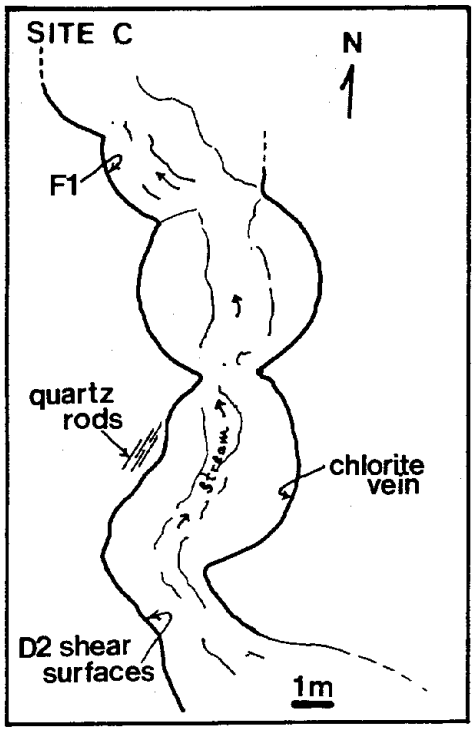


Figure S3-8. Detailed map of SITE C, the plunge pool exposure.

the valley narrows upstream with more rock outcrop closer to the stream.

SITE C

Geomorphic Features

Site C (Figure S3-5) is the waterfall which marks the main knickpoint of the tributary. The stream currently is cutting laterally, following a dipping fracture. As the stream undercuts the rock, pieces eventually collapse into the streambed. Before the stream started to migrate along the fracture, it poured over the lip of the rock and eroded the valley headward by potholing. The remnants of three potholes are readily discernable. It is possible that potholing played a major role in the headward erosion of this stream, but no evidence remains except these potholes. Note that the rock at the waterfall forms a barrier all the way across the valley. This barrier is higher than the valley floor immediately upstream. Upslope from the waterfall the barrier gradually loses definition and disappears into the slope. This barrier presumably approximates the shape of the rock floor of the upstream valley.

The valley upstream from the waterfall is totally different from that below the falls. Here is the pre-incision valley with a flat, well-developed floodplain and gently sloping valley walls with no rock outcrops (Inset, Figure S3-3). The slopes are covered with fine-textured colluvium which is exposed in several meander cutbanks upstream from the falls. The colluvium is derived from weathered bedrock and saprolite. The floodplain material is thin and rock is exposed in the stream bed in places. Erosion of this part of the valley is occurring immediately upstream from the rock barrier where the slopes are steeper and the stream is flowing across bedrock.

Geomorphic History

The history of this small valley can be summarized as follows. Prior to incision the whole valley had the form now preserved in the upper part. As Pequea Creek lowered its bed, the tributary started incision of its bed. As the bed was lowered, erosion stripped the valley sides of any loose, easily eroded material such as fine-textured colluvium, deeply weathered schist, and saprolite. This stripping exposed fresh rock which was then subjected to breakup into large blocks by the rigors of Pleistocene climates. Incision of the barrier at the major knickpont has caused the start of lowering of the valley above the waterfall.

When did all this occur? Presumably during the Pleistocene if the landscape has undergone as much erosion as the offshore record indicates. The form of the upper part of the valley represents the pre-incision condition, although the floor of the valley was probably a few meters higher. Each bench at the mouth of the tributary may represent a period of stability during which

the junction of the tributary with Pequea Creek was matched and erosion was at a minimum, at least at the mouth of the tributary.

Structural Features

The rock at Site C is a muscovite-chlorite-biotite schist with plagioclase porphyroblasts (Figure S3-4). In the uppermost plunge pool a thin (centimeter to decimeter) vein comprised almost entirely of chlorite with a trace amount of muscovite (Figure S3-6) cross-cuts compositional layering in the schist at a low angle. Alignment of chlorite crystals in the vein defines a fabric parallel to the outcrop schistosity, indicating that the vein was emplaced either during or prior to the development of the regional S1 schistosity.

The foliation is defined by the parallel arrangement of muscovite, chlorite and biotite. Quartz rods (Figure S3-7) define a lineation that trends 030° to 050°. This lineation will be seen again at Site D. The foliation has an average orientation of 230° strike and 08° north dip. The minute hinge axes of D2 crenulations define a lineation on the foliation surfaces that trends 030° to 050°. The axial planes of the crenulations are not easily recognized at this locality due to the very small wavelength and amplitude of the crenulations (sub-millimeter). In other areas the axial planes of crenulations are steeply dipping to the northwest and generally strike 230° to 240°.

An exposure upstream about 3 meters from the uppermost plunge pool contains discrete shear surfaces that cross-cut the coarsely crystalline schistosity (Figure S3-8). One of these surfaces is best exposed under the small overhang on the outcrop surface that faces eastward. These surfaces, defined by recrystallized and reoriented muscovite and new growth of biotite, are associated with northeast-directed thrusting, as explained in Chapter III of this guidebook.

SITE D

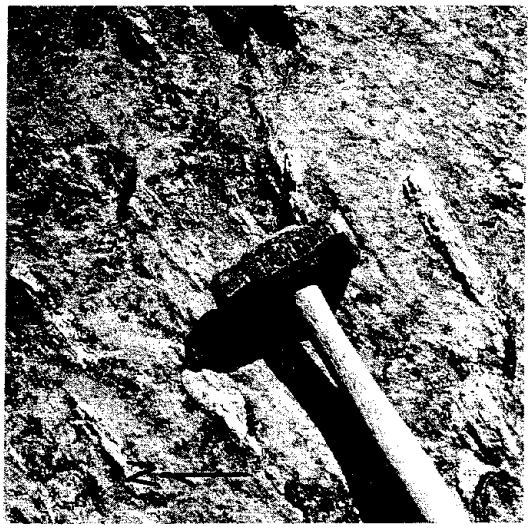
Geomorphic Features

This is a major knickpoint on Pequea Creek. The stream valley is very constricted and a rapids exists in the streambed. It is not known if the stream flows on in situ rock here, but it seems likely. There are a few small potholes near water level on the side of the stream opposite the trail.

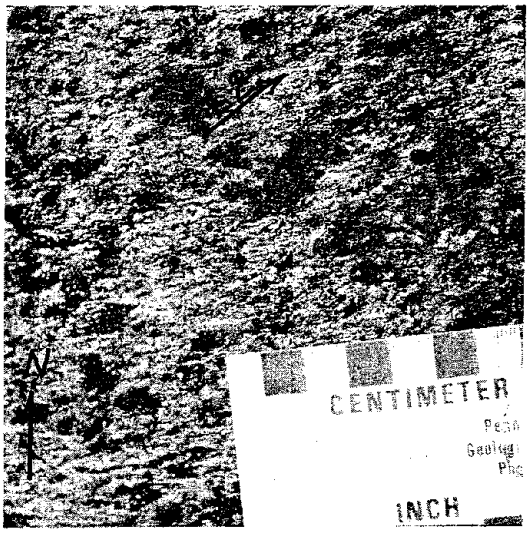
Structural Features

The exposure just above creek level provides an excellent opportunity to examine the metamorphic features of the rock, particularly microstructures and mineral assemblages related to late (post-Taconian) northeast-directed thrusting. See Chapter III for a detailed discussion. The rock at this site is a muscovite-chlorite-biotite garnet schist with abundant

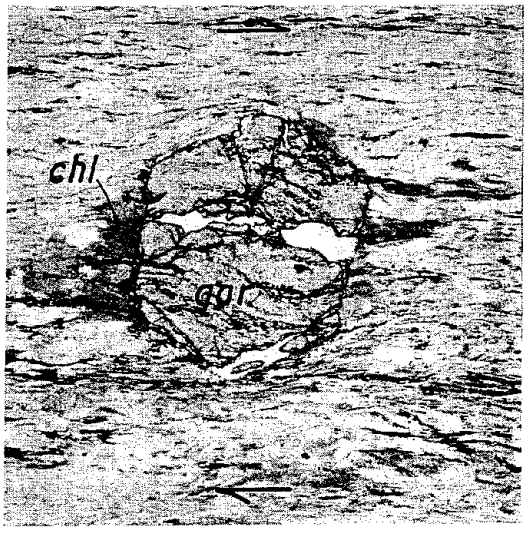
S3-4



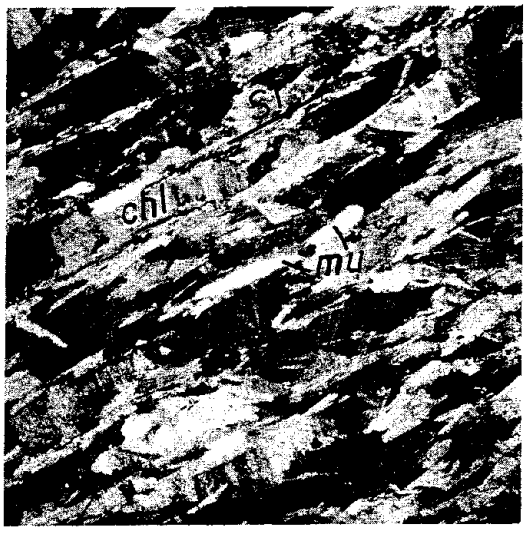
S3-7



S3-9



S3-11



S3-6

Figure S3-4 (center top, facing page). Photomicrograph of coarse crystalline muscovite-chlorite-plagioclase schist from the area of STOP 3; field of view is 6.0 mm.

Figure S3-6 (bottom right, facing page). Photograph of a vein rich in chlorite with trace muscovite; field of view is 2.5 mm.

Figure S3-7 (center left, facing page). Photograph of quartz-rod lineations on subhorizontally oriented S1 schistosity.

Figure S3-9 (center right, facing page). Lineations defined by (M2) chlorite pressure fringes on (M1) garnet porphyroblasts. The view is looking down on a flat-lying exposure surface (SITE D).

Figure S3-11 (bottom left, facing page). Photomicrograph of garnet with chlorite pressure fringes; field of view is 2.5 mm. View looking northwest at a subvertical face cut perpendicular to the schistosity. Shear sense is top-to-the-northeast.

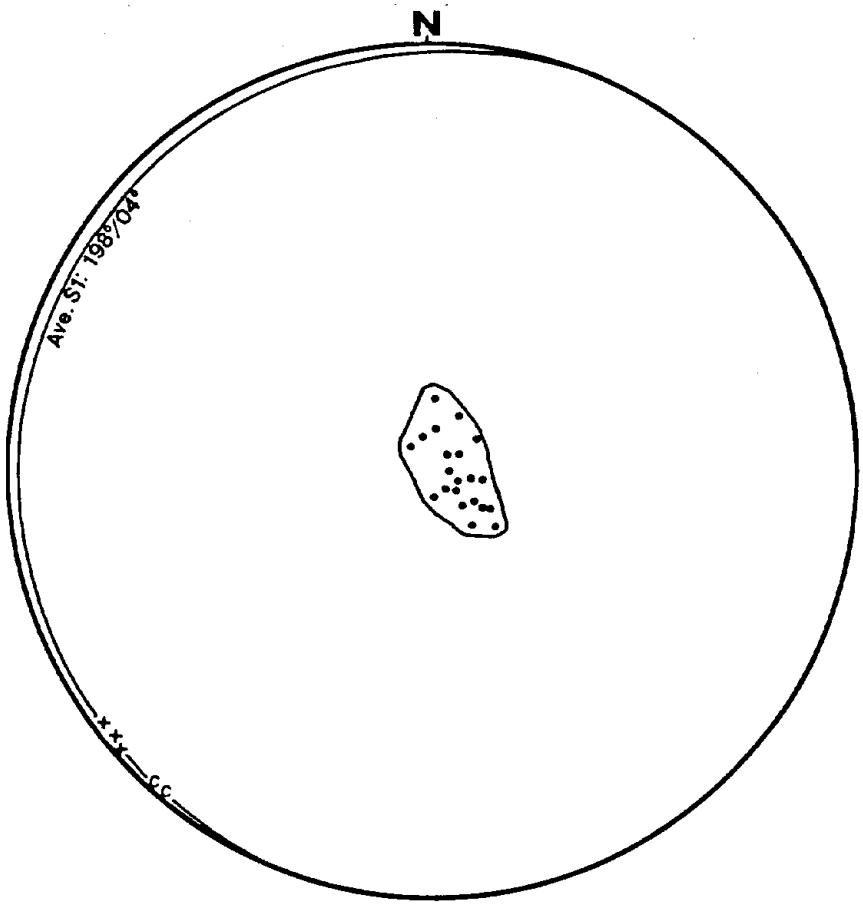
porphyroblasts of plagioclase (1-4 mm in diameter; Figure S3-9). This rock is very similar to the rock at Site C except for the presence of garnet. Garnet appears to be restricted to certain compositional horizons throughout the Martic Forge-Pequea area (although the rock at Site D does not appear to be much different in composition from that at Site C). The distribution of metamorphic minerals, such as garnet, is a problem for future work.

The foliation is defined by parallel arrangement of muscovite, biotite and chlorite, and by planar aggregates of quartz and plagioclase. The rocks of the Pequea Creek area have been referred to in the past as Wissahickon Formation: chlorite-albite schist (Knopf and Jonas, 1929; Cloos and Hietanen, 1941). It is clear from the minerals identifiable in hand sample that "chlorite-albite schist" is a gross generalization. Minerals such as biotite and garnet, although present in relatively minor amounts, characterize the metamorphic grade of the rock more accurately than do chlorite and albite.

Because the schistosity in this area is subhorizontal, the measured orientations of the foliation are scattered (Figure S3-10). Freedman et al. (1964) identified this foliation as S1, associated with the Taconian D1 deformation (Lapham and Bassett, 1964). Many small isoclinal folds (a few centimeters to a meter in amplitude) have axial planes parallel to this S1 foliation. The flat-lying foliation defines the minor Pequea Synform on the north limb of the Tuquan Antiform.

The schistosity surfaces exposed at this locality bear evidence for northeast-directed shear thrusting. Mineral lineations defined by thin (millimeters to centimeter) quartz rods are exposed on the foliation surfaces. These lineations trend generally 030° 050° (Figure S3-7). Just above the water level, small garnet porphyroblasts (2-7 mm in diameter) with

chlorite pressure fringe tails (Figure S3-9) define a lineation that trends parallel to the quartz lineations. These lineations tell us that shear thrusting occurred either to the northeast or



•Poles to S1 +Chlorite lineations
cPoles to crenulation lineations on S1

Figure S3-10. Schmidt net stereographic projection of structural features found at STOP 3.

the southwest. Oriented samples were cut parallel to the lineations and perpendicular to the foliation to reveal the profiles of the garnet-chlorite microstructures (Figure S3-11). The sense of shear thrusting was determined by the asymmetric distribution of chlorite pressure fringes about the host garnet crtystal. At this locality, the sense of shearing is top to the northeast. (See Chapter III for detailed discussion.)

Metamorphism

Polyphase metamorphism, another interesting feature at this locality, constrains the relative timing of the shear thrusting. The primary metamorphic mineral assemblage in this rock is muscovite-chlorite-biotite-garnet-plagioclase. This assemblage

of minerals is the result of the regional metamorphic episode (M1) interpreted to be Taconian (Lapham and Bassett, 1964). The mineral assemblage of chlorite-biotite-garnet places these rocks

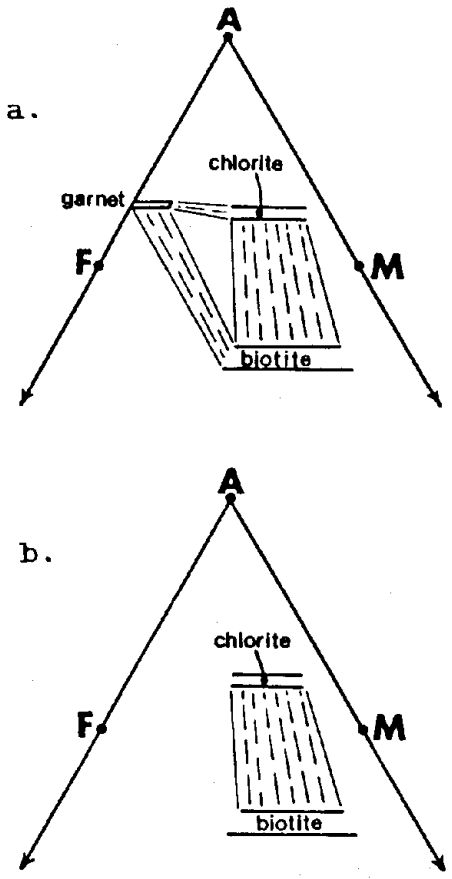


Figure S3-12. a. AFM diagram for the M1 metamorphic mineral assemblage observed in the area of STOP 3. b. AFM diagram for the M2 metamorphic mineral assemblage observed in the same area.

in the upper greenschist to lower amphibolite facies (Figure S3-12a). If you look closely at the garnets, you will notice that, in every case, they are surrounded by a corona of chlorite, including the pressure fringes. In thin section this corona is comprised mostly of chlorite and minor biotite. This is evidence of a second metamorphic episode (M2). This second metamorphism, however, is associated primarily with microstructures (such as the garnets with chlorite pressure fringes) and was not as intense or regional as the M1 metamorphic episode. The second metamorphism is characterized by the retrograde relationship of chlorite and biotite growth at the expense of garnet. This assemblage indicates lower to middle greenschist facies. (Figure S3-12b).

The correlation of the M2 retrograde metamorphism with the shear thrust microstructures indicates that the thrusting was

post-Taconian. An age of 320 Ma obtained by Lapham and Bassett (1964) for the M2 retrograde metamorphism clearly shows that the thrusting is later than Taconian. At present, the orogeny associated with this episode of metamorphism and deformation is uncertain. Although the date of 320 Ma suggests an Acadian age, the regional style of deformation, mainly transpression (see Chapter III), associated with this retrograde metamorphism is characteristic of Alleghanian deformation in other areas of the Piedmont.

SITE E

Note the floodplain on the opposite side of the stream. Although the material looks like alluvium from a distance, some of it is lake sediment deposited when the dam existed downstream. The area was covered with water to a point near Site F.

SITE F

At this location Pequea Creek makes a sharp right-angle turn and cuts across the general trend of topography (070°) at an orientation of about 292° (N22°W) (Figure S3-1). This is just one of many drainage channels that cut this ridge in the same manner. The drainage positions apparently are not controlled by fractures, because the orientation of fractures is about 310° (N40°W). Instead, it appears that the drainages have developed normal to the trend of topography in response to the development of the valley to the north.

The valley which extends northeast from Martic Forge (Figure S3-1) to Smithville was mapped as containing limestone of the Conestoga Formation in the area near Smithville by Knopf and Jonas (1929). Recent remapping (Berg and Dodge, 1981) extended the Conestoga Marble to Martic Forge and also showed more of the Conestoga Formation at Colemanville. However, there is no evidence that any marble occurs in this valley. Limited data from outcrops marginal to the valley and interpretation of limited exposures of saprolite indicate that the valley is underlain by a facies in the Wissahickon Formation which has been much more deeply weathered than surrounding rock. This rock is a biotite-microcline schisto-gneiss made up of 40-50% quartz, 10-15% biotite, 5-15% muscovite, and 5-20% plagioclase of which less than 10% is microcline. Complete weathering of this rock produces a fine-grained quartz sand which could be easily eroded to produce the valley. It is also probable, but currently unprovable that at some time in the past the valley was a channel for through-drainage which has now been pirated at one or more places. It may be presumed that, as the valley deepened relative to the more resistant adjacent rock, the tributary drainages developed normal to the topographic trend.

SITE G

Along this part of the trail there are several good views of floodplain alluvium exposed on the opposite bank. This material

appears to be entirely the result of fluvial deposition not influenced by the downstream dam.

SITE H

At this site a small tributary to Pequea Creek cuts across the topography at right angles. Fractures have an orientation of 315o (N45oW) and the schistosity has a low north dip. Again the argument must be made that the drainage developed normal to the trend of topography which is controlled primarily by the orientation and dip of the schistosity.

Reboard the buses at the parking lot of the Martic Forge Hotel.

(1.0 23.4) (Return mileage, via road, to Martic Forge). LEAVE PARKING LOT of Martic Forge Hotel, turning left, then immediately right onto Marticville Road (the eastward continuation of Pequea Boulevard, PA 324). Procede toward east, following valley which may be a "window" eroded through Wissahickon Formation into underlying Conestoga Formation, although recent work casts doubt on this interpretation.

0.8 24.2 BEAR RIGHT onto Red Hill Road (do not continue on Marticville Road through underpass beneath the "low grade" railroad line).

2.0 26.2 Tucquan Glen Road bears off to right. CONTINUE STRAIGHT AHEAD on Red Hill Road.

0.3 26.5 BEAR RIGHT onto Nissley Lane.

0.1 26.6 Cross Martic Heights Road. Continue straight ahead (south) on Nissley Lane, passing Martic Elementary School on right.

1.3 27.9 TURN RIGHT onto Drytown Road, heading southwestward. Note that this ridgetop drainage divide road is one of the few straight and relatively level roads in the area.

1.9 29.8 TURN LEFT onto Hilldale Road, heading south.

0.4 30.2 Cross Old Holtwood Road, continue south.

0.1 30.3 TURN RIGHT onto new Holtwood Road (PA 372). Procede toward southwest.

1.1 31.4 Muddy Run hydrolectric pumped storage reservoir, operated by Philadelphia Electric Co., on left.

0.6 32.0 Enter extensive roadcut in Wissahickon Schist at the east end of Norman Wood Bridge over the Susquehanna River.

0.8 32.2 Begin crossing the Norman Wood Bridge. The Holtwood Dam, with its associated hydroelectric and coal-fired steam driven power plants, is visible approximately one mile upstream to the right. When built in 1910, the Holtwood Dam was the largest in the United States. Lake Aldred, behind the Holtwood Dam, has a pool elevation of 169 feet. The Conowingo Reservoir, visible downstream to the left, is impounded behind the

Conowingo Dam in Maryland and has a pool elevation of 109 feet.

0.8 33.0 West end of the Norman Wood Bridge. Now in York County.

0.1 33.1 TURN RIGHT into the Pennsylvania Power and Light Company's Lock 12 Historic Area.

0.2 33.3 Parking lot and picnic area. LUNCH STOP. The directions for the self-guided tour will be distributed at lunch. Buses will leave this area for passenger pick-up at the completion of Stop 4 two and one-half hours after arrival. The pick-up point is approximately 0.7 mile to the south along River Road.

STOP 4. HISTORY AND RIVER-BED FEATURES NEAR LOCK 12 OF THE SUSQUEHANNA AND TIDEWATER CANAL

Leaders: Glenn Thompson and Bill Jordan

At this stop (Figures S4-1 and S4-2) we will have opportunities to see a partially restored lock of the Susquehanna and Tidewater Canal (see Chapter VI), other points of historic interest, and to observe some of the erosional features taken as evidence for intense Pleistocene flooding. This site is owned and maintained by Pennsylvania Power and Light Company (PP&L).

PLEASE BE AWARE THAT WATER LEVELS CAN CHANGE GREATLY AND RAPIDLY DEPENDING ON HOW MUCH WATER IS RELEASED FROM THE SAFE HARBOR DAM UPSTREAM. A SIREN AND FLASHING LIGHTS ON THE POWER HOUSE OF THE HOLTWOOD DAM (1 MILE UPSTREAM) SERVE AS A WARNING THAT WATER LEVELS ARE ABOUT TO RISE. YOU ARE STRONGLY ADVISED NOT TO GO OUT ONTO THE ROCK BED OF THE RIVER, WHICH MAY BE EXPOSED IF WATER LEVELS ARE LOW. THIS CONDITION CAN CHANGE VERY QUICKLY!

You have two and a half hours at this stop. During that time you may: 1) eat lunch, 2) visit the sites of historic interest in the vicinity of the picnic grounds, and 3) when you are ready, begin the self-guided tour of potholes and other features to be seen on and near Peavine Island, located south of the Norman Wood Bridge. Follow the trail under the bridge; assistants will be posted at key points to show the way (see Figure S4-1) Allow at least $1^1/_2$ hours for the self-guided walking tour.

The walking tour is moderately strenuous and requires a climb up a steep trail at the end. If you prefer, you may remain in the Lock 12 area until the buses leave for the pick up point.

POINTS OF HISTORIC INTEREST

Besides the partially restored lock, other points of interest in the picnic area include lime kilns and remains of a sawmill. As you walk south toward Peavine Island, you will pass the abutments of a canal bridge, the foundation of a canalmen's store and tavern, a wall that supported part of the towpath, and

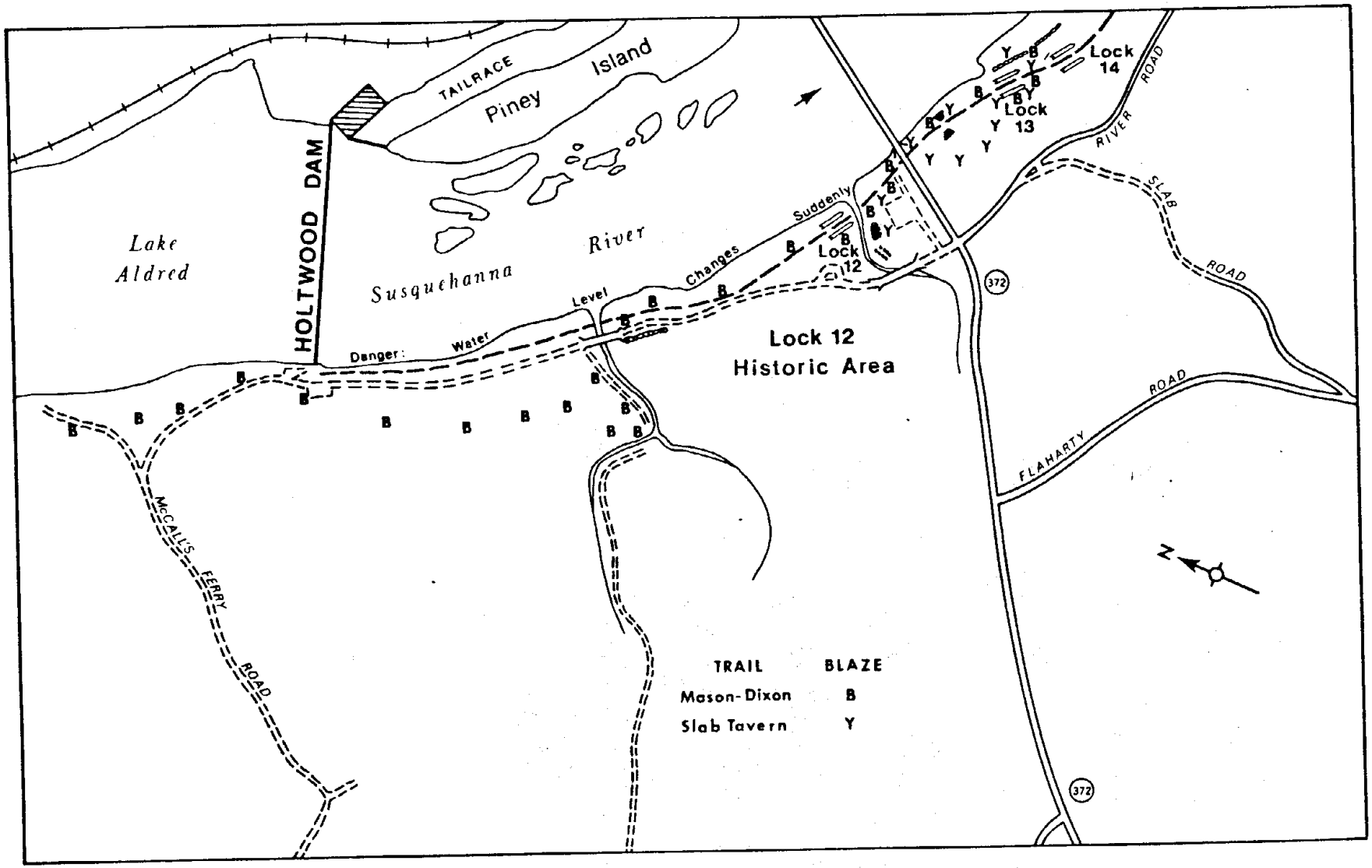
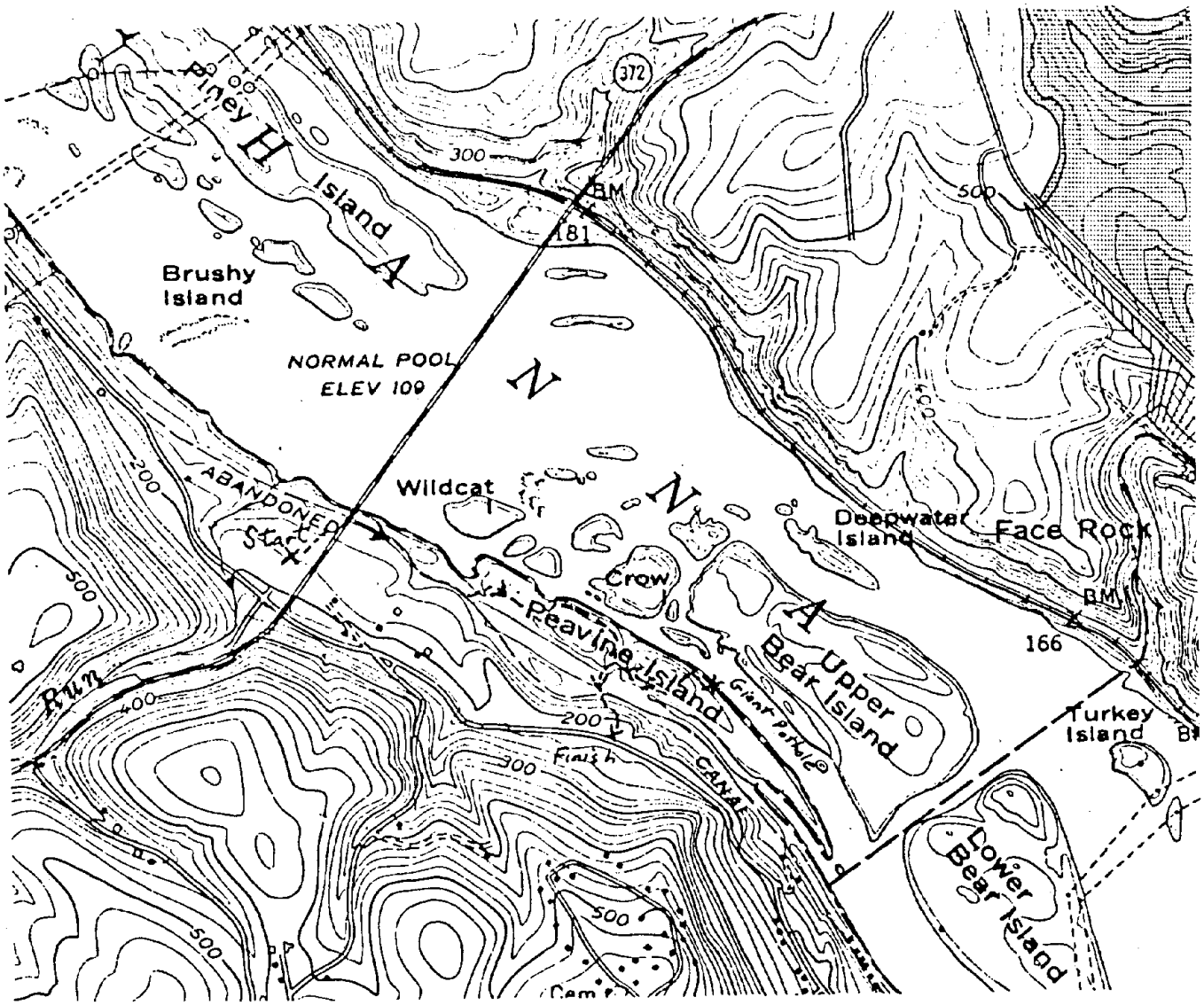


Figure S4-1. Map of the Lock 12 historic area, showing locations of locks and blazed trails



1 MILE 1/2 0

FEET 1000 0 1000 2000 3000

1 KILOMETER 1 .5 0

CONTOUR INTERVAL 20 FEET
NATIONAL GEODETIC VERTICAL DATUM OF 1929

Figure S4-2. Map of the vicinity of STOP 4, showing the route of the self-guiding tour of Peavine Island.

Lock 13. The remains of McCall's hotel and Lock 14 are slightly farther downriver. These features and other points of historic interest along the route of the canal are discussed in Chapter VI.

TOUR OF PEAVINE ISLAND

Because observation points are scattered and are in some places positioned in small or precarious locations, the trail is set as self-guiding. Each of you should have received a numbered description sheet correlated with numbered sites along the prescribed path. Caution is urged when moving about in rough or high places.

Beginning at the parking area at the west end of the Norman Wood Bridge, first follow the blue-blazed Mason-Dixon Trail to a point beyond the remains of Lock 13. From there the trail is indicated by flagging tape. After crossing Peavine Island, you will have a STEEP CLIMB back to the road to meet the buses. DO NOT ATTEMPT TO ABORT THE TRIP BY SHORT-CUTTING THROUGH TERRAIN CONGESTED WITH THICKETS OF RHODODENDRON, BLOWDOWNS, AND HIGH-LEVEL EROSION CHANNELS. Copperhead snakes also have been reported in the region. Assistants will be positioned for directional aid at critical points of the trail.

Specific features to be observed here are island shapes, accordant island heights, potholes, erosion channels and fluvial erratics.

		LEAVE parking area on River Road where blue-blazed Mason-Dixon Trail returns to mainland at south end of Peavine Island. Procede north on River Road.
0.5	33.8	TURN RIGHT onto PA 372. Procede northeast toward Norman Wood Bridge.
0.2	34.0	West end of Norman Wood Bridge.
0.8	34.8	East end of Norman Wood Bridge, continue on PA 372 (Holtwood Road).
0.8	35.6	Muddy Run hydroelectric pumped storage power reservoir on right.
0.7	36.3	TURN LEFT on River Road.
0.4	36.7	Cross Old Holtwood Road. Continue straight ahead toward north.
0.2	36.9	Cross Drytown Road. Continue straight ahead.
1.3	38.2	Pinnacle Road on left leads to Pennsylvania Power and Light Company overlook of the Susquehanna River at Pinnacle Hill, illustrated on the front cover of this guidebook. From here on River Road our view to the north is over Tucquan Glen and the northern part of the Susquehanna River gorge.
0.8	39.0	Cross Tucquan Creek.
0.2	39.2	Entrance to Lancaster County Conservancy's Tucquan Glen Nature Preserve on left.
0.6	39.8	Clark Run section of the Pennsylvania Scenic River System on the right. Continue north and northeast on

River Road through all intersections until reaching Martic Forge.

3.1 42.9 TURN LEFT onto Pequea Boulevard (PA 324) at Martic Forge. Cross Pequea Creek.

0.1 43.0 CONTINUE STRAIGHT AHEAD on Pequea Boulevard (PA 324).

0.4 43.4 TURN RIGHT onto Colemanville Church Road.

0.3 43.7 Colemanville Church. Park and DISEMBARK FOR STOP 5. Walk north over the bridge crossing the "low grade line" railroad bed. Procede for 0.2 miles to access point onto railroad grade for STOP 5.

For the location of this stop, see Figure S3-1: X marks the spot!

STOP 5. SAPROLITE NEAR COLEMANVILLE CHURCH.

Leader: Bill Sevon

This stop is along the now-abandonded low-grade railroad about 400 feet (120 m) east of the bridge over the track at Colemanville Church. The location is about 0.6 mi (1 km) northwest of the Martic Forge Hotel, the ending point of the walk at STOP 3.

The purpose of the stop is to examine several exposures which demonstrate the variable nature of saprolite in the Holtwood area. An exposure of colluvium also will be viewed.

The rock of the Wissahickon Formation in this area is muscovite-chlorite, plagioclase schist with some occurrence of biotite and garnet. Primary foliation (S1) is well displayed and varies slightly around an orientation of 235° strike, 35° dip (N55°E, 35°N). Feldspar in the weathered rock is quite noticeable.

SITE A

A prominent outcrop on the south side of the tracks across from the access point exposes material which has the appearance of rock, but has been weathered to the point of uncohesiveness which allows easy digging with a shovel or shaving with a knife. The structure of the rock is perfectly preserved and parts of the outcrop are more indurated than others. The lack of red color typical of much saprolite in the Holtwood area suggests that this material is close to the contact with the underlying weathered rock. Note the abundant muscovite, the quartz, and the coarse texture of the material.

SITE B

About 100 feet (30 m) east of Site A on the south side of the tracks is a good example of slumping. A well-developed head scarp at the top of the slope marks the present upper limit of slumping. The material being slumped is saprolite and failure presumably is occurring along the plane of primary schistosity which dips about 35° toward the railroad track.

SITE C

A few hundred feet (about 100 m) east of Site B is another slump somewhat hidden in underbrush and small trees. The slump scar exposes saprolite which contrasts to that present at Site A because of the bright red (2.5 YR 5/6) color and the fineness of its grain size. The red color is typical of saprolite occurring at higher topographic positions throughout the area. The schist from which the saprolite developed is a facies different from that present at Site A. The red color results either from penetration of iron-rich staining fluids to a greater depth than at Site A or from the presence of more iron-bearing minerals in the rock itself. The saprolite has a greasy feel and it is easy to imagine why slump failure would occur along the plane of schistosity. Orientation of schistosity here is 240° strike, 46° dip (N60°E, 46°N).

At the top of the saprolite there is a sharp contact with overlying colluvium. The lower colluvium comprises unconsolidated layers of quartz and mica derived from the saprolite. Fragments of schist are rare. The layers have a slight downslope dip and variable colors varying from red (2.5 YR 5/6) to reddish yellow (5 YR 6/8) to yellow (10 YR 6/8). The zone is up to 0.5 m thick and has a sharp contact with the overlying coarse colluvium. This colluvium has a brown color and abundant schist fragments up to 30 cm long and 10 cm thick surrounded by quartz and muscovite matrix. There are more fragments than matrix. The fragments have a crude alignment of flatness subparallel with the slope. This zone is up to 2 m thick and is typical of much colluvium occurring on slopes in the Holtwood area. The two colluviums may represent two different times of colluviation or two phases of the same event.

SITE D

Another 100 feet (30 m) or so farther east on the south side of the tracks is an outcrop of fairly fresh rock which shows good schistosity with orientation 235° strike, 35° dip (N55°E, 35°N) as well as a good crenulation cleavage.

SITE E

A few hundred feet (about 100 m) farther east on the north side of the tracks is a good outcrop of weathered rock which is not quite weathered to the saprolite stage. There is some variability in hardness of the rock in this outcrop. Some of it is weathered sufficiently to exfoliate and some of it is relatively hard. This rock is coarse-grained like the saprolite at Site A. Feldspar is abundant and has not been totally weathered to clay. Orientation of schistosity is the same as at Site D.

This is a good place to ponder the variability noted within a relatively small stratigraphic and areal distance. Obviously

there are lithologic variations within the Wissahickon Formation. These lithologic variations presumably contribute to variations in the depth of weathering which produces saprolite. An additional variable, not seen here, is the variation in dip of schistosity which strongly affects the depth to which weathering will proceed and the thickness of resultant saprolite.

LEAVE Colemanville Church on Church Road, heading south.

0.3 44.0 TURN LEFT onto Pequea Boulevard (PA 324), procede toward east.

0.4 44.4 Cross Pequea Creek, pass through Martic Forge area (for third time), heading east. Pequea Boulevard is now called Marticville Road.

0.8 45.2 Continue on Marticville Road (PA 324), passing through underpass beneath the "low grade line" just beyond intersection with Red Hill Road. Marticville Road turns sharp right at other side of underpass.

0.3 45.5 Marticville Road (PA 324) turns left. CONTINUE STRAIGHT AHEAD toward east, parallel to the "low grade line," on Pennsy Road.

1.3 46.8 Entrance on right (through stone arch bridge beneath "low grade line" embankment) to Lancaster County Conservancy's Trout Run Nature Preserve.

0.7 47.5 Cross Rawlinsville Road. Continue straight ahead eastward on Pennsy Road. The valley area to the right, paralling Pennsy Road and the railroad is another possible "window" eroded through the Wissahickon Formation into Conestoga Formation below.

1.4 48.9 TURN LEFT, toward the north, on Willow Street Pike (PA 272), a divided highway.

0.3 49.2 Cross Martic Line, now traveling on Conestoga Formation, with other carbonate and basal clastic units outcropping in parallel east-west striking bands ahead. There are 5 bands of clastics (mostly Antietam Formation) north of the Martic Line; E. Cloos mapped these as bounded by thrust faults.

0.8 50.0 Cross Pequea Creek.

1.2 51.2 In the field to the right is the barely visible trace of a sinkhole that opened as a consequence of the 1984 "Easter Sunday" Martic Earthquake.

0.8 52.0 Entering the town of Willow Street. Continue straight ahead.

1.4 53.4 Intersection with Beaver Valley Pike (US 222). Continue straight ahead on Willow Street Pike which extends north as US 222.

0.7 54.1 Excavated blocks of Conestoga Formation showing multiple generations of folding are used for decorative landscaping on the right.

0.6 54.7 Cross Mill Creek.

1.4 56.1 Cross Conestoga River. This location was the head of the Conestoga Navigation Company's slackwater canal from the Susquehanna River, at Safe Harbor, to Lancaster. Entering City of Lancaster.

1.2	57.3	Lancaster Chamber of Commerce and information center in old Southern Farmers' Market building on left.
0.1	58.4	Penn Square, with its 1874 Civil War Soldiers and Sailors Monument, is the center of Lancaster City.
0.2	58.6	TURN RIGHT onto Chestnut Street at the Brunswick Hotel.

END OF FIRST DAY.

ROAD LOG--DAY 2

Mileage Inc	Cum	
0.0	0.0	START. Leave from Chestnut Street entrance of Brunswick Hotel.
0.1	0.1	TURN RIGHT and procede one block south on Duke Street. Lancaster City Hall on right, St. James Episcopal Church on left.
0.1	0.2	TURN RIGHT onto Orange Street. Procede west, Lancaster County Court House is on the left.
0.2	0.4	Cross North Queen Street; Central Market one-half block to the left on North Market Street. Continue west on Orange Street.
0.4	0.8	BEAR RIGHT at Getty convenience store onto Marietta Avenue, continue west.
0.2	1.0	Cross College Avenue. St. Joseph Hospital on right; the Franklin and Marshall College campus is two blocks to the right on College Avenue.
0.5	1.5	Cross President Avenue. The Lancaster Historical Society's building and Wheatland, home of the 15th President of the U.S., James Buchanan, are on the left. Wheatland, in the country and surrounded by wheat fields at the time of Buchanan's residence, is considered one of the best preserved and most authentic of all presidential homes. Built in 1828, James Buchanan occupied Wheatland from 1848 through his presidency (1857-1861) until his death in 1868.
0.9	2.4	Cross Little Conestoga Creek.
0.4	2.8	TURN RIGHT onto Farmingdale Road. Procede north.
0.3	3.1	Cross Conrail railroad tracks.
0.8	3.9	Cross Little Conestoga Creek for the second time. Now proceeding east.
0.2	4.1	TURN RIGHT into Toys-R-Us parking lot. Walk to STOP 6.

STOP 6. LONGS PARK EXPOSURE OF CONTACT BETWEEN THE KINZERS AND LEDGER FORMATIONS

Leaders: John Taylor and Charles Scharnberger

The exposure here is along the eastbound exit ramp from Route 30 at Harrisburg Pike. Longs Park (a Lancaster city park) is directly across Harrisburg Pike from the top of the exit ramp (Figure S6-1).

GENERAL DESCRIPTION

After an orientation in the Toys-R-Us parking lot, you will walk down the driveway of the parking lot, across Farmingdale Road and follow the marked path down onto the berm. Walk down the berm (where there are some outcrops of the Longs Park Member

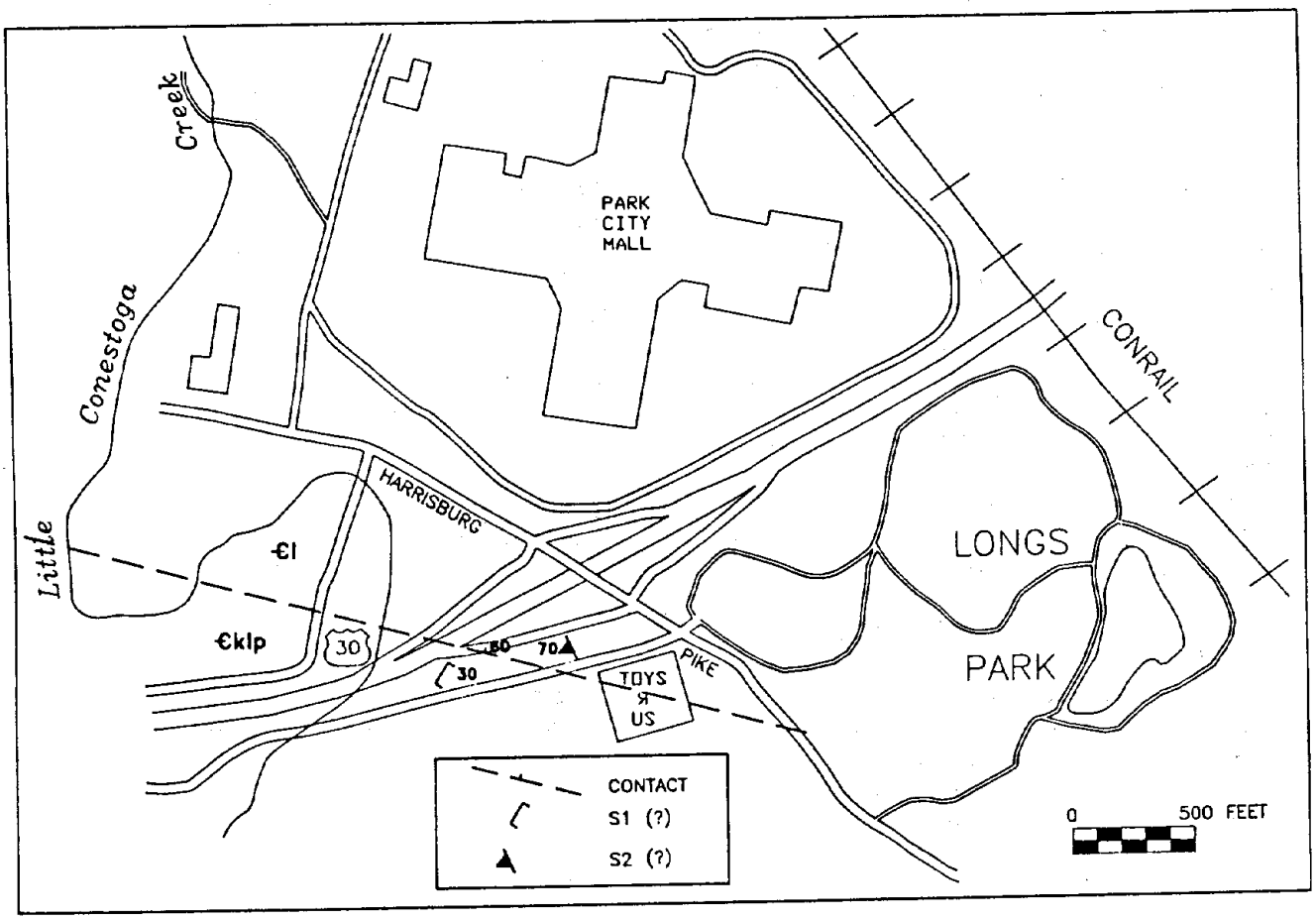


Figure S6-1. Location map for STOP 6, showing geologic features; €l=Ledger Fm., €klp=Longs Park Mbr., Kinzers Fm.

of the Kinzers Formation) to the pavement at the foot of the Route 30 exit ramp. PLEASE BE CAUTIOUS OF TRAFFIC EXITING FROM ROUTE 30! Turn right and walk UP the ramp, observing the rock exposed in the ramp cut.

You begin in the Longs Park Member and walk toward the Ledger. The contact between the Ledger Formation and the underlying Longs Park Member of the Kinzers is about 100 feet (30 m) up the ramp from the point where you come off the berm. The attitude of the contact is deceptive because the direction of the ramp is only about 25° different from strike; the contact is dipping obliquely toward you. The contact strikes about 285° (N75°W) and dips about 60°N. Bedding generally is obscure.

Cleavage in the Longs Park Member dips shallowly (about 30°) to the southeast. This cleavage, which is not obvious in the Ledger Formation, *may* be related to the S1 cleavage that you saw yesterday and will see again at STOP 7. At a point 20 feet (6 m) up the ramp from where you begin your walk, the S1(?) cleavage is folded into an antiform that plunges about 50° in the direction S30°E. Just below the contact there appears to be a small fault (thrust?) dipping northward. If this fault pre-dates the tilting of the rocks to their present orientation, it could have originally dipped on the opposite direction. There is a prominent spaced cleavage in the Ledger that dips steeply (70°) to the southwest (strike is about 155° or S25°E). This *may* be related to S2 seen at STOPS 1 and 7. The strike of this cleavage, however, is unusual for S2 in the Conestoga or Wissahickon Formations.

Continue walking to the top of the ramp (a total distance of about 900 feet (275 m), turn right, re-cross Farmingdale Road near its intersection with Harrisburg Pike, climb up the grassy slope to the Toys-R-Us parking lot (stepping over a low fence) and return to the buses.

SIGNIFICANCE OF THIS EXPOSURE

This stop provides the first opportunity for the field conference to examine the contact between dark, shaley off-platform deposits at the top of the Kinzers Formation and the pure dolomites of the overlying Ledger Formation, which are believed to have formed in carbonate platform environments (Chapter IX). Recovery of the trilobite genus *Peronopsis* from shales within the upper Kinzers (Campbell, 1969) at this locality established a Middle Cambrian age for the transition to shallow platform conditions brought about by progradation of the carbonate margin. This is the type section proposed by Gohn (1976) for his Longs Park Member, the upper shaly member of the Kinzers Formation. Gohn (1976) and Campbell (1969, Figure 4) correlated (physically and temporally) this interval in the Lancaster area with the siliciclastic-rich upper member of the Kinzers in the York area (Stose and Stose, 1944), envisioning a laterally extensive tongue of shaley basinal lithofacies produced by a single transgressive pulse at some time in the Middle Cambrian. Lower Cambrian fossils recently recovered from the upper Kinzers in the York area (STOP 11) require a revised stratigraphic model (see Chapter IX). Accordingly, a conservative approach is followed in assigning the name

Greenmount Member to the upper member of the Kinzers in the York area (see Chapter VIII), given the uncertain spatial relationship of those strata to the interval identified as the Longs Park Member here in the Lancaster area.

Try to keep what you see here in mind for comparison with the Kinzers-Ledger contact at STOP 11, the last stop of the day.

Unfortunately, this stop also provides a proper introduction to some of the problems confronted by geologists in the Conestoga Valley throughout the years--specifically, poor exposure, structural complexity, and strong diagenetic overprint. The problem of limited exposure is obvious in the Kinzers; note in particular the nearly complete loss of access to shaly intervals, which have provided most of the faunal data. The remaining exposures of strongly cleaved shaly limestone demonstrate the difficulty of distinguishing sedimentary features from those of tectonic origin. The exposures of Ledger Dolomite provide an excellent example of the nearly complete obliteration of primary depositional textures by diagenesis.

		LEAVE parking lot by Harrisburg Pike exit. TURN RIGHT onto Harrisburg Pike, procede southeast towards Lancaster. Longs Park of the City of Lancaster on the left.
0.3	4.4	Pass under Conrail railroad tracks.
0.5	4.9	TURN RIGHT onto President Avenue at Faulkner Chevrolet dealership.
0.8	5.7	Cross Marietta Avenue. Lancaster Historical Society and Wheatland on the right.
0.3	6.0	TURN LEFT onto Columbia Avenue. Procede east, passing former Hamilton Watch Company factory on the left.
0.4	6.4	TURN RIGHT onto West End Avenue, procede south.
0.6	7.0	Cross Manor Avenue, continue straight ahead. West End Avenue has now changed name to Hershey Avenue.
0.6	7.6	TURN RIGHT onto Wabank Street. Procede west.
1.1	8.7	TURN LEFT into parking lot of the Miller Quarry of Eastern Industries, Inc. STOP 7.

See Figure III-1 (Chapter III) for the location of this stop relative to the Lancaster Valley Tectonite Zone and the locations visited on the first day of the conference.

STOP 7. H. R. MILLER QUARRY--EXPOSURE OF A TECTONITE ZONE IN THE CONESTOGA VALLEY, LANCASTER COUNTY, PA.

Leaders: Rodger Faill and Dave MacLachlan

INTRODUCTION

The Conestoga Limestone (or Marble) underlies much of the southern part of the Lancaster Valley in central Lancaster County. These carbonates, probably deposited during the Cambrian

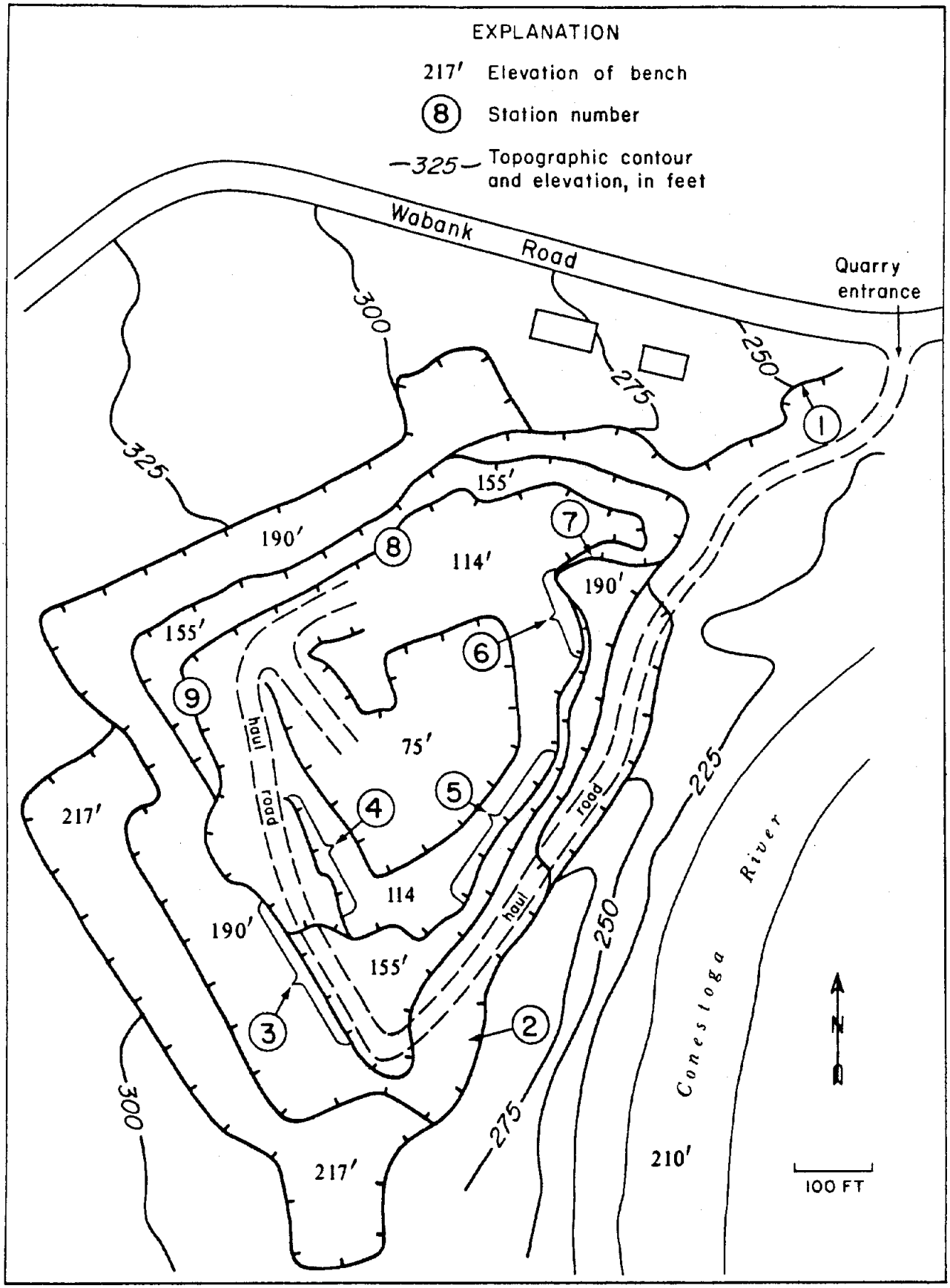


Figure S7-1. Map of the H. R. Miller quarry (Eastern Industries), southwest of Lancaster, Pennsylvania

and Ordovician Periods, represent a transitional facies between the purer carbonates of the continental shelf and the basinal siliciclastic rocks deposited during the expansion of Iapetus. During the Late Ordovician Taconian Orogeny (D1), the Conestoga Formation was metamorphosed to the biotite grade, greenschist facies, and a moderately strong S1 foliation developed parallel to the lithic layering (bedding=S0). Subsequently (perhaps during the Middle to Late Devonian Acadian Orogeny), a 4-5 km wide east-northeastwardly trending tectonite zone formed, transecting the Conestoga Valley just south of the city of Lancaster (Valentino and MacLachlan, 1990). The rocks within this zone (including those in the H. R. Miller Quarry) were retrograded to the lower greenschist facies. The exposures in this quarry are exceptional because they exhibit the structures, fabrics, and textures associated with: 1) the early prograde metamorphism, 2) the Lancaster Valley Tectonite Zone and retrograde metamorphism, and 3) subsequently developed fabrics (S3) and faults.

THE H. R. MILLER QUARRY

The H. R. Miller quarry was recently acquired by Eastern Industries, Inc. The quarry lies in the center of the Lancaster Valley in the southwest part of the Lancaster 7 1/2 minute quadrangle, on the southwest outskirts of Lancaster City and 2 km east of Millersville. The Conestoga River is the principal drainage for this valley--one of its numerous meander loops passes within than 300 feet east of the quarry. The quarry is quite old, having been first opened in the 19th century. But it wasn't until the latter part of the 20th century that intensive quarrying created the sizeable pit we see today. At the present time, the quarry is nearly 1200 feet across (longest dimension), and more than 225 feet deep. Remnants of four earlier working levels (at the 114, 155, 190, and 217 foot elevations) remain as benches at various places around the sides of the quarry (Figure S7-1).

Permission to enter quarry should be obtained from the quarry superintendent, Scott Handwerk, at the quarry office. You are advised to call several days in advance to insure that your visit does not coincide with scheduled quarry blasts or heavy haul road usage.

Appreciation is given to Charles Scharnberger, Millersville University, and Albert Mabus, Eastern Industries, for their recent survey of the quarry from which the topographic base map in Figure S7-1 was derived.

LITHOLOGIES

The quarry lies in the middle of the wide Conestoga Limestone outcrop belt that underlies much of central Lancaster County. The Conestoga comprises the "impure" carbonate deposition on the continental slope between the purer carbonates

of the continental shelf, and the argillaceous sediments of the adjacent ocean basin.

The Conestoga Limestone is rather heterogeneous, but the major fraction of it consists of rythmites, an interbedded sequence of thin to medium bedded microcrystalline to finely phanerocrystalline limestone alternating with partings and very thin to thin beds of phyllite. The limestone beds were apparently dilute turbidites derived from the carbonate bank on the craton to the northwest. The phyllites represent inter-turbidite intervals of argillaceous deposition. Other facies of the Conestoga formation include coarse conglomeratic limestone: some are autoclastic; others are clearly alloclastic. Dolomitic orthoquartzitic sandstone is also present in places. The thick conglomerate beds are exposed along the north sideof the quarry (at Stations 1 and 8); they possibly represent a subaqueous debris flow. The rythmites occupy the remainder of the quarry. Pyrite cubes are very common in some of the beds.

The rocks in the H. R. Miller Quarry have been metamorphosed to the biotite grade, greenschist facies, and were subsequently retrograded to the lower greenschist facies within the Lancaster Valley Tectonite Aone (Valentino, Chapter III, this quidebook).

STRUCTURAL GEOLOGY

The rocks in this immediate area have been subjected to three tectonic (deformational) events, labeled D1, D2, and D3. Fabrics, such as foliation (S), crenulation (C), or folds (F), among others, related to a specific deformation are correspondingly identified (for example, S1, C2, or F3). The original sedimentary bedding is labeled S0. The first deformation produced a high grade (biotite) greenschist facies metamorphism and imparted a bed-parallel foliation (S1) to the rocks. This deformation is believed to be Taconian (Late Ordovician) in age (Freedman and others, 1964). The second deformation (D2) produced the regional tectonite zone, which expresses itself with a strong subvertical cleavage (S2), tight upright folds (F2), and a gently northeastward plunging crenulation (C2) associated with a low-grade greenschist facies retrograde metamorphism. The linear fabric is characteristic of the tectonite zone, and almost entirely disappears within a short distance to the north of the quarry. Some elements of the zone persist only sporadically to the south, but the transition out of the tectonite zone is almost equally abrupt in that direction as it is to the north. The age of this deformation is not definitively established. Radiometric cooling dates in southeast Pennsylvania, ranging from 380 to 320 Ma and clearly associated with or subsequent to the D2 metamorphism, indicate it was older than Alleghanian, and younger than Taconian. The third deformation (D3) created a moderately northwest-dipping foliation (S3) that appears only sporadically in the quarry. This deformation may be a late phase of D2, or a separate orogenic

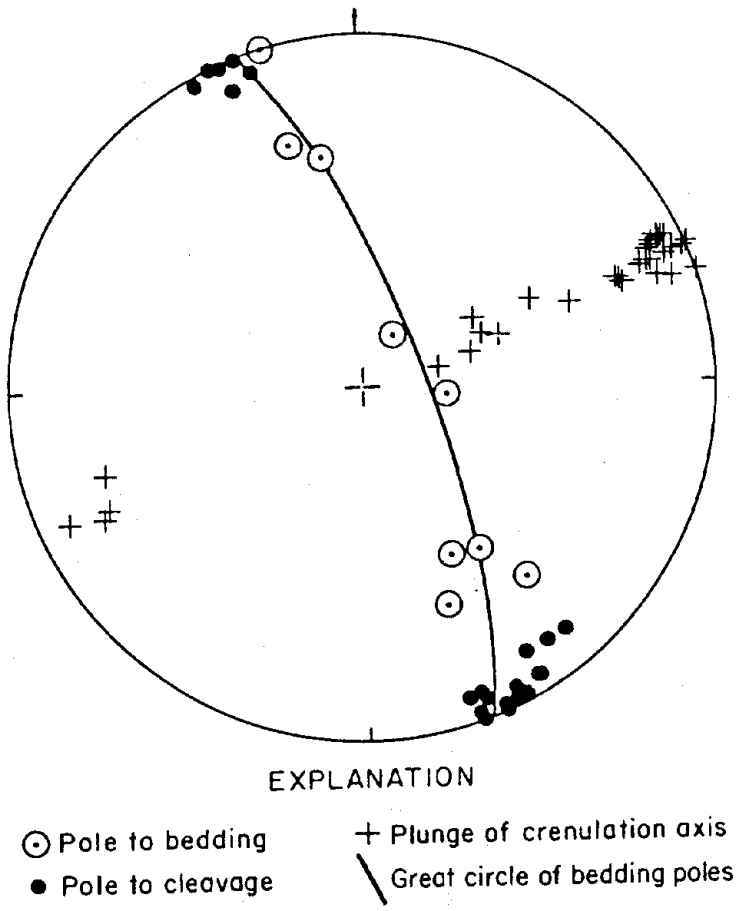


Figure S7-2. Stereogram of bedding/foliation (S0/S1), cleavage (S2), and crenulation (C2) measurements in the H. R. Miller quarry, illustrating the geometric concordance of the structural elements. The pole to the great circle fit to bedding poles plunges 260°-14°.

event. Its age is not known, but it is certainly pre-Late Triassic, and probably Alleghanian in the usual sense.

S0 & D1 Structures

The original sedimentary bedding (S0) is still quite evident throughout the quarry, although tracing individual beds any great distance can be problematic because of the impress of later D2 structures. The metamorphism associated with D1 had scant mesoscopic impact on the carbonate fraction of the rock. On the other hand, the clay fraction was thoroughly recrystallized and given phyllitic aspect. Chlorite and muscovite are the dominant minerals. Corroded shreds of remnant prograde biotite have been determined at some localities, but have not been established as definitively present here.

The S1 foliation in these rocks is caused by the strong alignment of these new phyllosilicate minerals with bedding. Four facts or inferences can be made about this foliation. First, the close parallelism between compositional layering (S0) and the S1 foliation is a pervasive aspect, not only in this quarry but elsewhere in the Conestoga outcrop belt. Second, the good alignment of the phyllosilicate minerals over such a large area implies that a significant differential stress regionally pervaded these rocks. Third, the apparent regional persistence of the major lithic units, and the absence of any other D1 structures, suggests that the compositional layering (bedding) was largely horizontal during metamorphism. Fourth, the absence of local concentrated metamorphic isograds (Valentino and Faill, 1990) indicates that the source of heat was the "normal" geothermal gradient, and cannot be attributed to a singular source such as an intrusion. From these inferences, it can be deduced that the S1 foliation was produced by a graduated vertical stress gradient throughout a large depositional basin, and that the source of the stress gradient was gravity. These are necessary and sufficient conditions for burial metamorphism. The only other signs of D1 deformation are some calcite veins and fractures parallel, and at a low angle, to bedding that have been folded by F2 folds.

D2 Structures

The principal large structure exposed in the quarry is a single, third-order anticline that plunges gently to the west southwest (Figure S7-2), concordant with other folds throughout the Conestoga Valley. The fold is not immediately obvious in the quarry because of the fairly homogeneous lithology and the profusion of meso-scale structures, especially the small tectonite zones. This fold appears to be upright, and it affects only bedding (S0) and the early foliation (S1), so we interpret it to be an F2 fold. The hinge of the anticline crosses the northern part of the quarry (Figure S7-1) so most of the quarry exposes the generally southeast dipping beds of the southern limb. The absence of conglomeratic beds in the south limb does not necessarily require a fault--conglomeratic bodies in the Conestoga Formation characteristically grade laterally into rythmites within very short distances.

The Lancaster Valley Tectonite Zone extends from the Susquehanna River at Turkey Hill eastward across the Conestoga Valley to the Honey Brook Upland in Chester County (Valentino & MacLachlan, 1990). Valentino (1989, and Chapter III, this guidebook) has described this tectonite zone in the Wissahickon Schist exposures at Turkey Hill. The tectonite zone is characterized by a strongly developed subvertical cleavage (S2), tight upright folds (F2), and gently northeast-southwest plunging crenulations (C2). Also characteristic of the tectonite zone is the reflective sheen that is common on bedding surfaces, and which is absent outside the zone. The zone itself may be a subvertical tabular structure.

As is true with many large tectonite zones, the Lancaster Valley Tectonite Zone is not a single homogeneous structure. Rather, it is quite heterogeneous, consisting of meso-scale zones of strongly developed vertical cleavage interspersed with meter-scale blocks (mesolithons) relatively free of the S2 fabric. This alternating pattern can be seen in the quarry, especially on the 114-foot level (Stations 4 to 9), where 2 to 5 meter-wide zones can be traced across the quarry in an ENE direction. The cleavage (S2) generally dips very steeply to the northwest and southeast (Figure S7-2). In the intervening mesolithons, the S2 cleavage is usually only weakly developed (to varying degrees), and open, upright folds with wavelengths of a meter or more are also present.

Smaller, centimeter to decimeter-scale folds (F2) usually congregate at the margins of the mesotectonite zones. These folds tend to be upright, angular, and rather tight; their axial surfaces verge northwestward (relative to the S2 cleavage), forming a 20 to 30 degree angle to S2.

The S2 cleavage has folded and locally transposed the phyllosilicate minerals comprising the S1 foliation to form a moderately pervasive millimeter-scale crenulation (C2). The crenulation is best developed where S2 is strongest, both within and adjacent to the mesoscale tectonite zones. The intersection of S0/S1 with S2 constitutes the crenulation axis, which plunges gently to the northeast and southwest (Figure S7-2).

Pyrite crystals are somewhat stronger than the enclosing limestone and thus they were deformed less than the surrounding carbonate minerals during the D2 deformation. This relative rigidity of the pyrites resulted in pressure shadows (fringes), which are areas on either side of the pyrite crystal in the direction of the minimum principal stress, in which calcite and quartz were precipitated. The alignment of these low-pressure shadows of many pyrites in the subhorizontal, east-northeast direction, indicates that this was the direction of extension during D2 compression. Some of these pyrites are elongated in this same direction, suggesting that they too were deformed, but to a lesser extent than the carbonate material.

Boudins are present in some of the purer, thin-bedded limestones of the rythmites. Some of the boudin separations are narrow, whereas others are quite wide; they are generally filled with calcite. The sharp boundaries of the separations, the absence of any necking of the limestone beds, and the absence of any phyllite filling indicate that moderately brittle conditions prevailed during the boudinage. The boudin axes appear to plunge gently to the east-northeast, subparallel to the crenulation axes.

In summary, the D2 deformation is geometrically a relatively simple tectonic event, despite its manifestation in various guises (hence the variety of structures). Although the most obvious element is the planar S2 cleavage, a strong linear aspect

is present which one can sense even in a hand specimen. This lineation plunges gently to the northeast and southeast throughout the quarry. Its strongest expression is the intersection of the S2 cleavage with the S0/S1 bedding/ foliation. The crenulations are a variant of this intersection. Other linear manifestations are the mesoscopic fold axes; extension and pressure shadow directions of the pyrite crystals; and the boudin axes. All these are co-linear with the S2 X S0/S1 intersections.

D3 Structures

Scattered here and there in the quarry is a shallowly northwest-dipping planar fabric which crosses F2 folds, and thus must be post-D2. This fabric, a spaced foliation, is better developed on the south limbs (where it is at a larger angle to bedding) than on the north limbs. Locally this S3 fabric is enhanced by parallel fractures. The significance of this D3 deformation is not understood.

Recognizable faults are not abundant in this quarry, probably because the interbedded, anisotropic character of the rythmites allowed other mechanisms to operate. However, they are common in some of the very thick beds, especially those in the hinge of the anticline. The faults appear to be rather late in the deformation, possibly D2, but more likely D3.

TOUR OF THE QUARRY

Tour begins near the quarry entrance, at Station 1 (see Figure S7-1). Proceed southwest to lip of quarry for overview; then begin descent down the haul road past Station 2 at the 190-foot bench level, past Station 3 (approximately the 155-foot bench level) to the 114-foot bench. Proceed counterclockwise around the quarry on the 114-foot level (Stations 4-9), then walk back up haul road to the buses at the entrance. PLAN TO BE BACK TO THE BUSES BY 11:00 A.M.

Station 1: [near the quarry entrance, northwest of the crushing equipment, just east of the piles of processed stone.]

The conglomeratic facies of the Conestoga Formation is exposed here. Bedding (S0) dips moderately to the northwest and is best seen on the northeast-facing wall at the west end of the exposure. The rock is a sequence of very thick to thin bedded crystalline limestone and conglomeratic limestone. Some of the thicker layers contain carbonate clasts at various orientations which imparts a conglomeratic appearance to the beds. However, it is not clear whether these beds are true sedimentary conglomerates or represent a deformational breccia.

Many of the bedding surfaces exhibit the sheen that is characteristic of tectonite zones. In addition, the subvertical S2 cleavage is well represented here, trending 050°, so well-developed in places that bedding is obscured. The wall facing

the crushers consists essentially of S2 surfaces. Tight F2 folds with S2 as axial plane cleavage are very common. This folding may have contributed to the apparent brecciation of the conglomerate beds. The well-developed lineation (seen especially well on the S2 cleavage surfaces) that plunges gently northeastward is the intersection of bedding/foliation and cleavage (S0/S1 X S2).

Station 2: [partway down the haul road along the southeast highwall is a pull-off on the left--Station 2 is along the highwall just above the 190-foot bench].

The rythmites that dominate this part of the quarry comprise interbedded 2-4 cm thick limestone beds and 1-10 mm phyllite beds. For the most part, bedding (S0) dips moderately to the southeast. However, a mesotectonite zone approximately 5 meters wide is also present. The trend of the tectonite zone is oblique to the face of the highwall, and a complete traverse across the zone from outside to within and out again can be made along the quarry face.

Within the zone, the bedding/foliation (S0/S1) has been rotated (transposed) to a subvertical orientation, parallel to the S2 cleavage. This S0/S1 transposition and the S2 cleavage reinforce each other to produce the intense fabric characteristic of tectonite zones. Other structures within this mesotectonite zone include: gently northeastward-plunging crenulations; elongated pyrite crystals with pressure shadows; and smearing lineations.

The tectonite zone boundaries are quite different. Through this transition out of the zone, the bedding/foliation (S0/S1) progressively diverges from the S2 cleavage. In addition, decimeter-scale tight, angular folds are common--they are not present at all within the zone. And some of the limestone beds exhibit boudinage. Other beds are clearly disrupted.

Beyond these transitional boundaries, bedding dips moderately to the southeast. The S2 cleavage is present outside of the zone, transecting bedding, but it is not nearly as intensely developed.

These mesotectonite zones appear to be relatively straightforward transpressional events, but the various structures present in them suggest three strain components, which together form a deformational mosaic, the D2. The first is a subhorizontal flattening perpendicular to the northeast trending lineation. It manifests itself primarily by the S2 cleavage, and the S0/S1 transposition into the S2 cleavage. The complementary extension to the northeast-southwest appears as the elongation of the pyrite crystals and the growth of pressure shadows, both in the lineation direction. The second component is a simple shear about the lineation direction, principally up-on-the-north, as evidenced by the rotation of bedding/foliation into the tectonite zone, and by the boudinage. Locally, a complementary down-on-

the-south simple shear is manifest by the small decimeter-scale folds at the zone boundaries. The third strain component is a simple shear with rotation about a vertical axis, producing a dextral horizontal displacement parallel to the tectonite zone. This rotation is evidenced by the asymmetry of the pressure shadows.

The D2 deformation producing the mesotectonite zones is evidently a rather complex tectonic event, calling upon several deformational mechanisms despite the relatively simple colinear geometry.

Station 3: [extends west of the haul road downward from the sharp bend just below the 190-foot bench to the 114-foot bench level].

A number of large boulders are placed along the east side of the haul road to help prevent equipment from going over the precipice. A short distance past the sharp turn is a boulder with pyrite crystals sporting aligned pressure shadows--an excellent photo opportunity! Along the highwall west of the haul road, bedding is quite apparent. The metamorphic fabric S1 is reinforced by the D1 calcite veins and fractures, all of which are folded by F2 folds. At the sharp turn in the haul road, beds are subhorizontal, but moderately open, upright folds with a fairly well developed S2 cleavage are just to the north. Farther north along this high wall are two additional mesotectonite zones (6-8 m wide), both with the vertical S2. In the intervening mesolithons, the bedding is subhorizontal to moderately dipping. The S2 vertical foliation is present (to a lesser extent) in the mesolithons, as are open folds (F2) with vertical axial surfaces. In addition, the moderately northwest-dipping S3 foliation transects S2 and the F2 folds.

[Upon reaching the 114-foot bench, turn right to follow the bench around the quarry in a counterclockwise direction.]

Station 4: [extends from the haul road southward to the southeast corner along the highwall below the haul road]

Along this stretch, two mesotectonite zones are encountered. The one at the beginning of the exposure is 2 m wide and is the same as the last one seen along the haul road above (in Station 3). Halfway south to the corner is an offset in the highwall, at which another 4-5 m wide zone occurs (which is also present in Station 3). The intervening rocks are subhorizontal with S2 absent or very poorly developed.

South of the second tectonite zone, beds are generally subhorizontal again, forming a syncline with one or two small open folds. At the corner, bedding has steepened to a moderate northwest dip. Some upright, tight F2 folds, with strongly developed S2, constitute the north margin of another mesotectonite zone. The gently to moderately northwest-dipping D3 foliation appears sporadically along here, enhanced by fractures, transecting the small F2 folds.

Station 5: [extends northeastward for nearly 300 feet from the southwest corner at the end of Station 4 along the irregular highwall on the southeast side of the quarry].

Some 75 feet from the corner is the same mesotectonic zone seen at the offset in Station 4. Bedding is locally northwest-dipping, reflecting mesoscale structures within the south limb of the major anticline. Another 50 feet to the northeast is another mesotectonite zone which aligns with the one at the beginning of Station 4, and the end of Station 3. Boudins are present here. Additional small folds and other structures are present farther to the northeast to the end of Station 5, where the rythmites give way to the underlying massive beds.

Station 6: [east of the deepest level of the quarry].

This is the hinge of the major anticline, which extends west-southwestward across the quarry to the west highwall. The beds on this east side are so thick that bedding is very difficult to make out. But thinner beds high on the wall show that bed dip gradually decreases from the south and becomes subhorizontal at this station.

Thick beds in the Conestoga such as these usually contain clasts, indicating that the beds are essentially subaqueous debris flows. Although the clasts don't jump out here, a careful search will reveal a few. The S2 cleavage is not well developed in these beds. A number of faults are present in these beds, with various orientations; slickenlines of various plunges indicate that the fault movements were more complex than the D2 movements. Because they do not fit geometrically the D2 deformation, they are assigned to the later D3 deformation. The presence of these faults in the very thick beds and their apparent absence in the rythmites suggests either that the faults are obscured by the other fabrics (S0 and S2) or that they developed only in the thick beds that lack mechanical anisotropy.

At the north end of this station, where the highwall protrudes most into the quarry, is a tight, upright anticline and syncline pair, with many of the D2 structures seen at the previous stations. This same fold pair is present in the opposite west highwall.

Station 7: [eastward projecting pocket of the quarry].

Along south wall of this pocket, the bedding and S2 foliation are parallel and subvertical. Faults cut across this fabric, suggesting a northeast-southwest directed maximum principal stress, clearly not congruent with the D2 deformation. Crenulations are common, and plunge in numerous directions, indicating a multiplicity of movements. Along the opposite north wall, bedding dips steeply to the north, being within the northern limb of the major anticline.

Station 8: [north high wall of the quarry].

Along this north highwall, the rock is thick bedded and occasionally conglomeratic, as at Station 1 with which these beds correlate. Bedding is difficult to ascertain, even from the relatively inaccessible 155-foot bench. The S2 cleavage is not well-developed, which is probably a function of the thick bedding (as at much of Station 6).

Station 9: [northwest corner of the quarry, along the west high wall].

The tight syncline and anticline seen at the north end of Station 6 are present in the thin-bedded rythmites near the corner of the quarry. These structures can be better seen from the 155-foot bench level (but access in this part of the quarry is hazardous). Overlying these beds are the thick beds present along the north high wall (Station 8).

[Proceed up the haul road to the top of the quarry, and return to the busses].

		Leave parking lot of Miller Quarry. TURN LEFT onto Wabank Road, procede west.
0.8	9.5	TURN RIGHT onto Millersville Road, procede north. The campus of Millersville University is located approximately one mile to the left (west).
0.7	10.2	Cross Manor Avenue, continue straight ahead to the north.
0.7	10.9	Cross Little Conestoga Creek (for the third time).
0.2	11.1	TURN LEFT onto Charlestown Road, procede west.
0.9	12.0	TURN RIGHT onto Centerville Road, procede north.
1.7	13.7	Cross Columbia Avenue, continue straight ahead to north. High ground ahead is east-west striking Chestnut Ridge underlain by basal Cambrian clastic units (Antietem, Harpers, and Chickies Formations).
0.6	14.3	Cross Conrail tracks.
0.2	14.5	Cross over limited access U.S. 30. TURN RIGHT onto entrance ramp to U.S. 30 West. Antietem Formation outcrops to left along the entrance ramp.
0.2	14.7	Procede west on U.S. 30. This highway, from here to York, follows the strike valley of Cambrian carbonate units stratigraphically above and south of the ridge of basal Cambrian clastics (Chestnut Ridge) to the right of the highway.
4.0	18.7	View ahead to west, across the valley of the Susquehanna River. The Mt. Pisgah ridge, also underlain by basal Cambrian clastics, is to the left (south) and the Hellam Hills, the western continuation of Chestnut Ridge, is to the right (north).
0.8	19.5	Exit ramp for PA 441. Antietem Formation outcrops in cut along exit ramp. Continue straight ahead.
0.2	19.7	Begin crossing the Wright's Ferry Bridge over the Susquehanna River. Spectacular view upstream to the

north (right) of the water gap eroded through the Chestnut-Hellam Hills ridge. Chickies Rock, a prominent landmark overlooking the Susquehanna on the Lancaster County shore, is visible in the middle distance on the right. Chickies Rock is the type locality of the Chickies Formation and collecting site of the first described specimens of the trace fossil *Skolithus*. If the atmosphere is clear, higher ground underlain by Triassic age rocks is visible in the far distance to the north. View downstream to the south (left) is of the Susquehanna River watergap cut through the Mt. Pisgah-Manor Hills ridge of basal Cambrian clastics. It is bounded on its north by the Stoner Fault, one of the "overthrusts" mapped by early workers in the region. Exposed in the County Line Quarry, visible on the ridge flank on the York County shore, are the Vintage, Antietem and Harpers Formations. If the day is clear, visible in the far distance downstream (left) is Turkey Hill (Stop 1 yesterday) and the high ground of the River Hills underlain by schists of the Wissahickon Formation. Visible in the immediate foreground to the left is the "longest multiple-arch highway bridge in the world," constructed in 1929-1930 for the Lincoln Highway (old U.S. 30, presently PA 462). In front of the reinforced concrete bridge are the ruined piers of several wooden covered bridges (and the subsequent replacement steel bridge) that crossed from Columbia in Lancaster County to Wrightsville on the York County shore. The second of the covered bridges served also as towpath for canal boats crossing between the southern terminus of the Pennsylvania Canal system in Columbia and the northern end of the Susquehanna and Tidewater Canal in Wrightsville (remains of the Susquehanna and Tidewater canal basin are still present just south of the concrete bridge). The first covered bridge, built in 1812 with a length of 5,690 feet, was the longest covered wooden bridge in the world. It was destroyed in 1832 by a flood and ice jam, and rebuilt in a slightly different location. This second bridge carried a covered double deck (one for each direction) canal boat towpath on its downstream side. The second covered bridge was burned June 28, 1863 by civilian volunteers and milita opposing troops of the Army of Northern Virginia moving east immediately prior to the Battle of Gettysburg. General Jubal A. Early's soldiers had occupied York on June 27th, and the following day forces under General John Gordon advanced on the river, crossing at Wrightsville. They were unsuccessfully opposed by Union fire from rife pits dug west of town. Failing at this defense, the local milita (27th Regiment of Pennsylvania Volunteers) and the supporting civilians fled east across the bridge, burning it in their retreat. The Wrightsville action marked the farthest penetration of Confederate troops to the northeast

during the war. The third Columbia-Wrightsville bridge was also a wooden covered bridge, but with a two-span metal truss fire break built into it. It was destroyed by a windstorm in 1896. The 1897 to 1964 replacement was a steel-truss bridge carrying both railroad and highway traffic, as did its predecessor. The still used concrete Lincoln Highway (old U.S. 30) bridge was built in 1929-30. The present Wright's Ferry Bridge (new U.S. 30) was opened to traffic in 1972.

1.1	20.8	York County end of Wright's Ferry Bridge.
0.4	21.2	Outcrop of Vintage Dolomite in roadcut on right.
0.6	21.8	EXIT divided highway (U.S. 30 West) at Wrightsville exit ramp.
0.4	22.2	Stop sign. TURN LEFT (south) onto Blessing Road at head of ramp. Mt. Pisgah ridge visible ahead to the south.
0.8	23.0	Cross Columbia Avenue (Lincoln Highway, now PA 462). Continue straight ahead to south on Cool Creek Road.
0.5	23.5	Cross Kreutz Creek. Road begins climbing north-facing slope of the Mt. Pisgah ridge, crossing Vintage and Antietem Formations in cuts on right.
1.0	24.5	TURN RIGHT onto Mount Pisgah Road, procede toward west.
0.3	24.8	Outcrop of Chickies Slate on left.
0.3	25.1	TURN LEFT at entrance to Samuel S. Lewis State Park which encompasses the summit of Mt. Pisgah. LUNCH STOP AND STOP 8.

STOP 8. SAMUEL S. LEWIS STATE PARK: THE HELLAM CONGLOMERATE AND OVERVIEW OF THE CONESTOGA VALLEY

Leaders: Bill Sevon and Charles Scharnberger

This is our lunch stop; after lunch, you may want to examine outcrops of the Hellam Conglomerate Member of the Chickies Formation (see Chapter I) that are scattered through the area. This point, atop Mt. Pisgah, affords an outstanding view of the Conestoga Valley and adjacent regions if the atmosphere is not too hazy (Figure S8-1). The description below is based on the guidebook for the 2nd annual field trip of the Harrisburg Area Geological Society (Sevon, 1983).

Samuel S. Lewis State Park consists of 75 acres at the top of Mt. Pisgah 2 miles south of Wrightsville. Mt. Pisgah (elevation 865 feet) is the high point on a series of prominent hills bordering the south side of the York Valley and extending 12 miles southwestward from the Susquehanna River. These hills are formed on a broad, fault-bounded anticlinal belt of black slate, quartzite, and conglomerate known collectively as the Chickies Formation (Cambrian). Some of Mt. Pisgah is composed of black slate, but the crest is upheld by the more resistant conglomerates which come to the surface here. Good outcrops of the slate occur in the deep roadcut on Mt. Pisgah Road northeast of the park entrance. A poor outcrop of weathered slate occurs as a small parking lot downhill to the south of the picnic

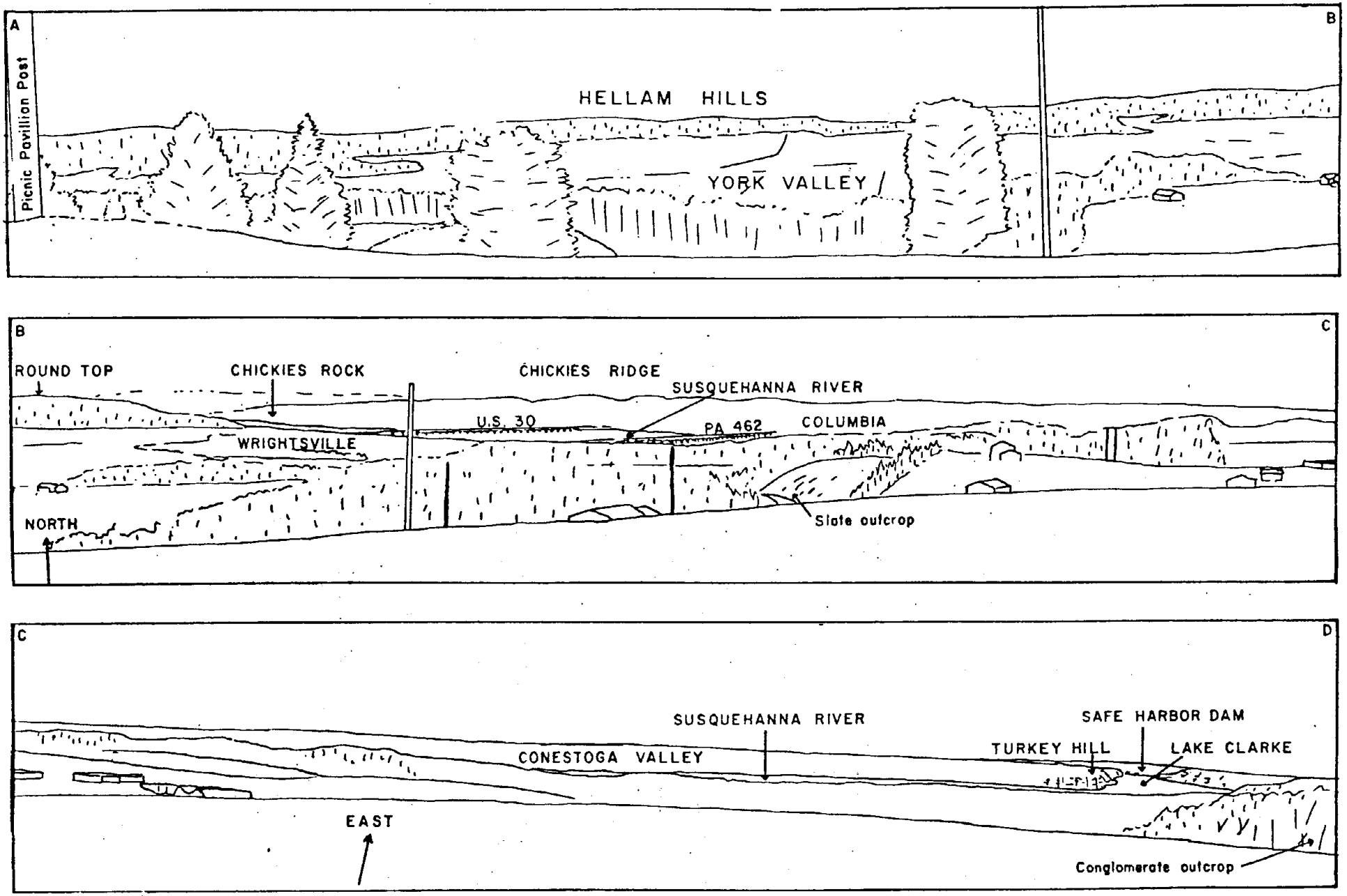


Figure S8-1. Panoramic view from Mt. Pisgah (STOP 8); from 2nd Annual Field Trip Guidebook, Harrisburg Area Geological Society, 1983.

pavilion. Moderate outcrops of the conglomerate, the Hellam Member, occur in the woods at the crest of Mt. Pisgah, but a better exposure is in the woods south of the road downhill from the picnic pavilion. These outcrops show a light gray, quartz pebble conglomerate with pebbles generally less than one inch in diameter, vague crossbedding, and flattening and elongation of some pebbles as a result of intense deformation. Numerous quartz veins cut through the conglomerate.

On a clear day the view from the top of Mt. Pisgah is impressive and encompasses several hundred square miles to the north, east and south. The accompanying panoramic sketch (Figure S8-1) locates some of the things seen from the field just east of the picnic pavilion. The following itemization starts looking north and moves to the east and then south.

1. The valley north of Mt. Pisgah is the York Valley which is underlain by Cambrian carbonates. The valley is generally one to two miles wide and extends southwestward from the Susquehanna River at Wrightsville 28 miles to Hanover. The southern boundary of the valley here is the Stoner Fault (see Chapter I and description for STOP 9), which runs along the base of Mt. Pisgah. The Martic Line is about 3 miles to the south of Mt. Pisgah.

2. To the north of the York Valley are the Hellam Hills which are a rocky belt of wooded hills developed mainly on the Cambrian Chickies and Antietam Quartzites and Harpers Phyllite. Some Precambrian metavolcanics also outcrop there.

3. Round Top is located southwest of a sharp bend in the Susquehanna River and in underlain by the Chickies Quartzite.

4. Chickies Rock exposes a large anticline in the Chickies Quartzite, verging to the north, with numerous second-order folds and meso-scale faults. To the west, the Hellam Hills seem to have controlled the course of the Susquehanna River and forced it to flow eastward. At Chickies Rock the river has managed to cut through the resistant quartzite and resume a southerly course. Only 7 miles farther east the Chickies Formation disappears beneath carbonates, thus removing the barrier, yet the river cuts through the quartzite ridge here rather than taking the "easy" path around the end. This situation, of course, is not unusual for the Susquehanna River in the Appalachian region. Another point to be noted here is that the Chickies Anticline, so prominent on the eastern shore of the river, does not seem to be present on the western (York County) shore. A bit farther west, however, near Highmount, there is a tight, overturned (to the north) anticline involving the Hellam Member. This fold is cored by metavolcanic rocks, thus suggesting that the Hellam is (at least in places) at the base of the Chickies Formation. Neither the metavolcanics nor the Hellam Member crop out on the Lancaster County side of the river.

5. Chickies Ridge (also called Chestnut Hill) is a narrow ridge of Chickies Quartzite which extends eastward to a point

near Rohrerstown, near the location of Stop 6. The north side of Chickies Ridge (and the Hellam Hills) is probably a fault, the "Chickies Overthrust" of Stose and Stose (1944). Whether this is a low-angle fault (a true "overthrust") or a high-angle reverse fault, however, is not certain.

6. On a really clear day, the Triassic Furnace Hills may be visible about 18 miles to the north. The highest point is Governor Dick Hill, 1150 feet above sea level, a knob of resistant diabase.

7. Columbia sits on an extension of the York Valley and is underlain by the Conestoga Formation.

8. The foreground hills with houses and school buildings are the continuation of Mt. Pisgah and are underlain by the less resistant Chickies Slate. This ridge is cut by the Susquehanna River and east of the river the ridge is called the Manor Hills.

9. The Lancaster Valley is the eastern, broader portion of the Conestoga Valley.

10. Turkey Hill was the location of STOP 1 yesterday. At this point the Susquehanna narrows from about 1.5 miles to less than 0.75 mile as it enters its lower gorge.

11. The Safe Harbor Dam marks the approximate axis of the Westminster Anticline hypothesized by Campbell (1933). The relatively flat-appearing upland surface seen to the left of the river and developed on the Wissahickon Schist possibly represents the dissected remnants of the warped Harrisburg Peneplain.

		Leave Samuel S. Lewis State Park. TURN RIGHT onto Mount Pisgah Road, procede east.
0.5	26.6	TURN LEFT onto Cool Creek Road, procede north downhill.
1.0	27.6	Cross Kreutz Creek.
0.5	28.1	Cross Columbia Avenue (PA 462). Continue straight ahead to north.
0.8	28.9	TURN LEFT onto entrance ramp to U.S. 30 West. Procede west on U.S. 30 along strike valley underlain by carbonate rocks. The high ridge of the Hellam Hills to the right is underlain by Cambrian clastics, including the Hellam Conglomerate at the base of the Chickies Formation, resting unconformably on Precambrian metabasalts and "metarhyolites."
3.8	32.7	Hellam exit. Continue straight ahead (west) on U.S. 30.
1.2	33.9	Unique "Shoe House" on the left now serves as an ice cream parlor.
1.6	35.5	Exit ramp for Mt. Zion Road and Rocky Ridge County Park. Continue straight ahead (west) on U.S. 30. Rocky Ridge Park contains good exposures of the Hellam Conglomerate.
1.0	36.5	Caterpiller Tractor Company plant on left.

0.9	37.4	Cross North Hills Road. Limited access highway ends. Continue straight ahead on U.S. 30 west.
1.3	38.7	Intersection with Eden Road. Harleigh-Davidson Motorcycle plant and museum to the right off Eden Road. Continue straight ahead on U.S. 30 west.
0.3	39.0	Cross channelized Codorus Creek.
0.4	39.4	Pass under I-83. Continue straight ahead, entering heavily congested area.
0.2	39.6	Cross North George Street (Business I-83). The City of York lies to the south (left). York served as capital of the United States from September 30, 1777 until June 27, 1778 following evacuation of Philadelphia by the Continental Congress. Confederate troops occupied York briefly in 1863 immediately prior to the Battle of Gettysburg, exacting a tribute of $100,000 as payment to spare the city from pillage and destruction. Their recall to Gettysburg was so swift that only $30,000 was collected.
0.1	39.7	Emigsville Member of the Kinzers Formation crops out behind McDonalds and other buildings on the left.
0.5	40.2	Cross Pennsylvania Avenue, entering York city limits. Continue straight ahead on U.S. 30 West.
0.6	40.8	Cross Roosevelt Avenue, leaving City of York. Continue straight ahead on U.S. 30 West. Limited access highway resumes.
1.0	41.8	Exit ramp for Carlisle Road (PA 74). Continue straight ahead on U.S. 30 West.
1.9	43.7	Former Medusa Cement quarry (now West Gate quarry) on right, developed in argillaceous carbonates of the Kinzers Formation. This pit is offically active, but is not being worked at present.
0.9	44.6	Limited access highway ends. EXIT RIGHT. U.S. 30 West merges with PA 462 (West Market Street). Get into left lane and continue straight ahead.
0.1	44.7	TURN LEFT onto Trinity Raod (PA 616).
0.1	44.8	TURN LEFT, almost immediately, onto Woodberry Road, procede toward east, then southeasterly.
1.3	46.1	Cross bridge over railraod tracks. TURN LEFT immediately and park on unpaved road. STOP 9.

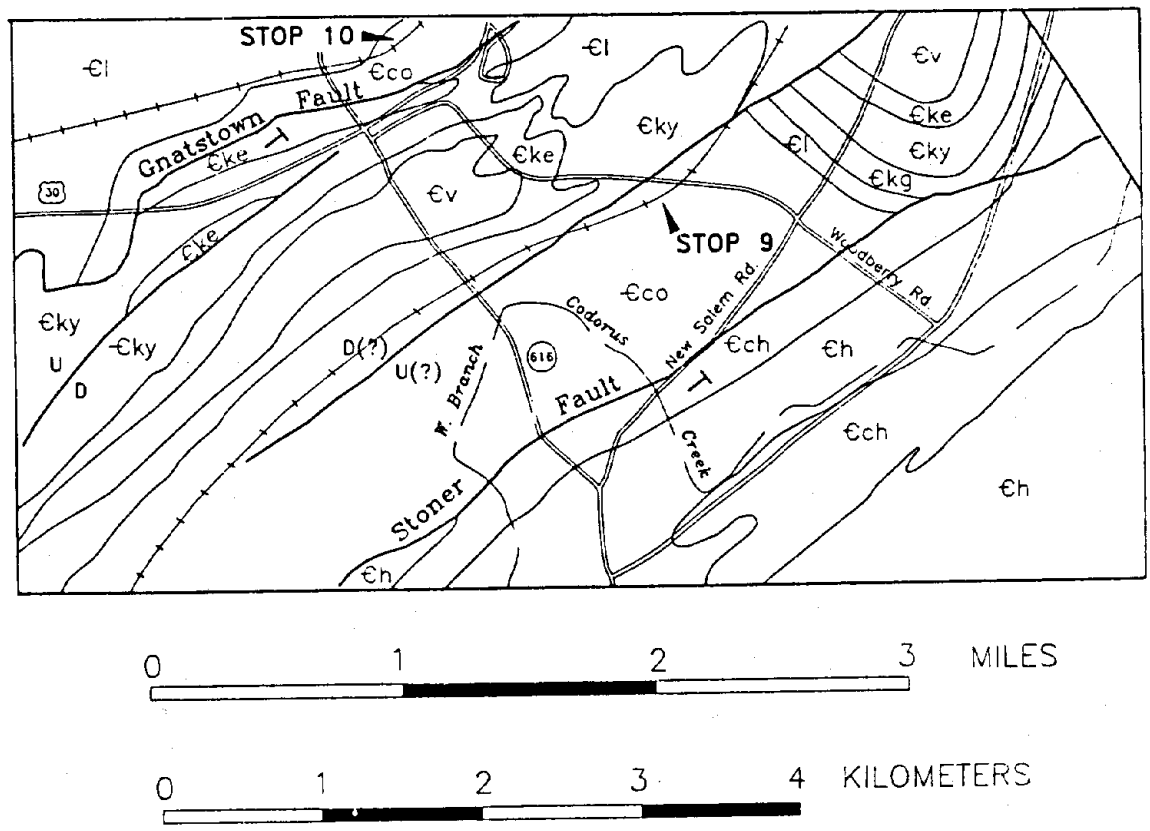


Figure S9-1. Location map for STOP 9 (also shows location of STOP 10). Geology is from Stose and Stose (1939b), but stratigraphic nomenclature of Ganis and Hopkins (Chapter VIII, this guidebook) is used for the Kinzers Fm. (Note, however, that the validity of these names has not been establised outside the West York Block (i.e., south of the Gnatstown Fault). €ch=Chickies Fm., €h=Harpers Fm., €v=Vintage Fm.; Kinzers Fm.:-€ke=Emigsville Mbr., €ky=York Mbr., €kg=Greenmount Mbr.; €co=Conestoga Fm.

STOP 9. THE CONESTOGA FORMATION EXPOSED ALONG THE RR TRACKS WEST OF WOODBERRY ROAD

Leader: Charles Scharnberger

The buses will let you off on an unpaved road off of Woodberry Road just south of the tracks and east of the bridge that carries Woodberry Road across the tracks (Figure S9-1). Climb down to track level (there will be a rope available to assist) and walk along the tracks to the bridge, a distance of about 250 feet (75 m). **BE CAUTIOUS: THESE ARE ACTIVE TRACKS.** A large block of vein quartz lies to the left of the tracks about 125 feet (38 m) beyond the bridge. The outcrop of Conestoga Formation (Wrightsville Member?) begins on the left about 75 feet (23 m) past the quartz block.

GENERAL DESCRIPTION

The Conestoga exposure in this cut has not been studied in detail and the description below is very tentative. Any relevant observations or interpretations that you might wish to call to the leader's attention would be much appreciated. The main purpose of stopping here is to see the Conestoga Formation at what might be called an "intermediate" level of deformation and metamorphism between what you saw this morning in the H. R. Miller Quarry (STOP 7) and what you are going to see next at STOP 10. STOP 7, you will recall, was located within the Lancaster Valley Tectonite Zone. Here at STOP 9 we have some aspects of the tectonite zone (strong cleavage, tight folding), but, on the other hand, there is, generally, only one cleavage present and argillaceous layers lack the sheen characteristic of rocks within the tectonite zone (see discussion for STOP 7 and Chapter VII).

STRUCTURAL SETTING

Figure S9-1 is taken from the geologic map of York County drawn by Stose and Stose (1939b) 50 years ago. It appears, on the basis of that map, that STOP 9 is located in a fault-bounded block that, in turn, is between the Stoner Fault ("overthrust") to the south and the Gnatstown Fault to the north. (STOP 10 is located just across the Gnatstown Fault.) Stose and Stose show the fault just to the south of our location as down-on-the-north. But, considering that the relatively older Kinzers Formation is mapped on the north side of the fault, this interpretation seems questionable. The fault farther north (but still south of the Gnatstown Fault) was mapped by Stose and Stose as up-on-the-north, which seems consistent with the presence of the lower member (Emigsville) of the Kinzers Fm. on the north side and the middle member (=York Mbr.?) on the south. Also, according to the Stose and Stose map, this location is on the west limb of a fairly simple-looking anticline, plunging south, that is confined to this fault block.

STRUCTURAL DETAILS

As was stated above, the rock in the cut has not been intensively studied. The prominent cleavage generally strikes about 050° and dips about 50° southeast. In many places bedding seems parallel to the cleavage, or nearly so, but about 25 feet (8 m) past the point where the outcrop begins, on the left side of the tracks (as you walk west), bedding can be seen in tight folds to which the cleavage bears an axial-planar relationship. A point to consider is how this folding and cleavage may be related to the D1 and D2 events discussed and illustrated earlier in the field conference in Lancaster County.

About 350 feet (107 m) farther down the tracks there may be a second cleavage present, dipping more steeply (70°) than the predominant cleavage. This seems most apparent in the outcrop on the right (north) side of the tracks, but is still rather vague. So there may be an S1 and S2 present in the rocks here, after

So there may be an S1 and S2 present in the rocks here, after all. But are these related to the same D1 and D2 as S1 and S2 at STOP 7? What do you think?

Stop here, turn around, walk back under the bridge and return to the buses.

DISCUSSION

The rock here is located between the Stoner Fault (to the south) and the Gnatstown Fault (to the north). As discussed elsewhere (Chapter I), these may or not be true overthrust faults. But, whatever they are, they seem to mark significant transitions in the geology of this part of the Piedmont. The Stoner Fault forms the southern limit of carbonate bedrock (though a few small outliers of Conestoga Formation were mapped by Stose and Stose between the Stoner Fault and the Martic Line). The carbonate rocks between the Stoner and Gnatstown Faults are tightly folded, as we see at this stop, but are not as "metamorphic-looking" as in the Lancaster Valley Tectonite Zone. Then there is a dramatic change in deformational character once the Gnatstown Fault is crossed and the West York Block is entered (see discussion in Chapter VIII). This will be illustrated at the next two stops, STOP 10 and STOP 11.

An interesting speculation is prompted by Valentino's (Chapter III, this guidebook) suggestion that 1) the Stoner Fault is a westward extension of the Brandywine Manor Fault mapped east of the Susquehanna River, and 2) the Brandywine Manor Fault (and associated minor faults) have experienced almost 20 kilometers of left-lateral strike slip. Restoring 20 km of sinistral movement on the Stoner Fault would put the location of STOP 9 just north of Mt. Pisgah (STOP 8). If the Stoner Fault has significant strike-slip displacement, what about the Gnatstown Fault? Could it too be, at least in part, a strike-slip fault? And if so, what are the implications for understanding the abrupt change in deformational style that occurs across that fault? These questions are raised just as that: *questions*, intended to stimulate more questions and thought.

		Leave parking area. TURN RIGHT, crossing railroad bridge and returning northwesterly on Woodberry Road.
1.3	47.4	TURN RIGHT onto Trinity Road (PA 616). Procede north.
0.1	47.5	Almost immediately, cross West Market Street (U.S. 30 and PA 462). Continue straight ahead to north onto Baker Road.
0.3	47.8	Cross railroad tracks.
0.2	48.0	TURN LEFT into parking area for the J. E. Baker Company plant and quarry. STOP 10.

STOP 10. EXPOSURES OF CONESTOGA LIMESTONE ALONG RR TRACKS SOUTH OF J. E. BAKER QUARRY

Leaders: John Taylor and Dave Hopkins

For the location of this stop, refer to Figure S9-1. You will get off the buses in the parking lot of the J. E. Baker Company. Walk a short distance back (south) along Baker Road and then left (east) along the railroad spur to the first outcrop, near the switch. PLEASE BE CAREFUL BOTH OF TRUCK TRAFFIC ON THE ROAD AND OF TRAINS: THESE ARE ACTIVE TRACKS.

LITHOFACIES

Three lithofacies common within the Conestoga Formation are represented at this stop: polymictic carbonate breccia, limestone-shale rhythmite, and lithoclastic lime grainstone. The first two are the most characteristic lithologies of the formation. The rhythmite, in particular, is not found in any of the associated Lower Cambrian carbonate units and produces a distinctive regolith with shale fragments that is very useful in recognizing areas underlain by the Conestoga. All three lithologies are assigned to periplatform or toe-of-slope environments.

Polymictic Carbonate Breccia

This most conspicuous lithofacies in the Conestoga Formation typically consists of cobble to boulder sized carbonate clasts, now recrystallized to calcitic marble, enclosed in a matrix (often dolomitized) of peloid sand or silt. The small exposure at this stop shows the very poor sorting, variable clast composition, generally massive character, and lenticular morphology of a typical periplatform carbonate breccia. Conestoga breccias at other locations include clasts more than 10 meters in diameter (Stose and Stose, 1944; Gohn, 1976). The breccias formed when submarine debris slides transported sediment from outer shelf/upper slope environments to the base of the slope creating chaotic deposits, often with a lenticular morphology mirroring the shape of the submarine channel. Unfortunately, recrystallization of the large clasts in this formation has erased all primary fabric. By analogy with better preserved toe-of-slope breccias of similar age in the northern (James, 1981; James and Stevens, 1986) and southern (Barnaby and Read, 1990) Appalachians, the large light-colored clasts are presumed to have been algal framestones, well-winnowed lime grainstones, and other lithologies created by synsedimentary cementation in shallow shelfbreak environments.

Limestone-Shale Rhythmite

This most characteristic lithology of the Conestoga is also the least resistant. This is well illustrated by its poor representation in the exposures at this stop. This lithofacies, which is dominant in strata deposited in distal toe-of-slope

environments or areas far removed laterally from the channels in proximal toe-of-slope settings, represents hemipelagic shale deposition occasionally interrupted by deposition of carbonate sediment as dilute turbidites.

Intraclastic Lime Grainstones

These thickly bedded to massive lime grainstones, consisting primarily of sand to granule sized intraclasts and some bioclasts, accumulated as periplatform lime sands transported downslope by gravity flow. They are interstratified with the limestone-shale rhythmites. Bedding is not obvious but is discernible, defined by thin bioclastic laminae. The attitude determined from these laminae is consistent with that displayed by the associated rhythmites, confirming that the massive grainstone exposures are, in fact, intervals rather than megaclasts. An excellent analog for these intraclastic grainstones is found in the limeclast sand facies of the Upper Shady Dolomite in southwestern Virginia (Barnaby and Read, 1990)

STRATIGRAPHY

The Conestoga Formation in this area directly overlies the Ledger Dolomite. Small exposures and drill core data from the fields to the north of this stop demonstrate that the strata in this railroad cut lie approximately 1000 feet (stratigraphically) above that formational contact. Fossils recovered from the intraclastic grainstone facies at this stop include inarticulate brachiopods (*Prototreta sp.*) and trilobites (*Modocia*) that establish a Middle Cambrian age for these strata. These stratigraphic relationships, along with proximal-distal trends (northwest-southeast) documented by Gohn (1976) for the Conestoga Formation, indicate: 1) that the shelf margin retreated to the northwest (present coordinates) sometime in the Middle Cambrian, and 2) that the boundary between the Lower and Middle Cambrian in the West York Block lies within the Upper Dolomite Member of the Ledger Formation or is represented by an unconformable contact between the Ledger and Conestoga Formations (see discussion in Chapter IX).

Also noteworthy at this stop (our first in the West York Block) is the relatively undeformed nature of the strata. Cleavage is only weakly developed in these rocks and the previously mentioned drill core data from the fields to the north document an essentially uninterrupted homoclinal sequence from the Upper Dolomite Member of the Ledger through 1000 feet or more of the Conestoga Formation. The limited extent of deformation in the West York Block is more convincingly demonstrated at our next stop in the Delta Carbonate Quarry.

If time allows, a drive-through of the J. E. Baker Co. quarry and plant will be taken.

THE J. E. BAKER QUARRY

The J. E. Baker Company's quarry is located in a thick sequence of high-purity dolomite of the Lower Cambrian Ledger Formation. The dolomite typically contains less than 1% total impurities. The Ledger Formation at this site is generally a light gray mottled with dark gray, coarsely crystalline, low-porosity dolomite. Locally, the dolomite contains spherical and undeformed ooliths. The pure nature of the carbonate and the presence of bedded oolite suggests that this is probably a shelf-derived carbonate.

Stose and Jonas (1939b) estimated the Ledger to be about 1000 feet thick in this general area, but current estimates are that the Ledger is substantially thicker here. Structural complexity in the quarry usually is masked by the lack of identifiable bedding, extensive faulting and the massive nature of the dolomite. The geologic complexity encountered through mining has led to the evolution of highly selective quarrying, with an intensive quality control program that ensures the quality of the various products manufactured.

The unique combination of physical and chemical characteristics of the dolomite allows the Baker Company to manufacture a high-purity refractory dolomite grain by burning the stone in a single pass through a high-temperature rotary kiln. This "grain" (granular sintered dolomite) is used to manufacture a refractory dolomite brick that finds widespread use in steel ladles, AOD vessels, specialty products for use in steel-making, and in lining the burning and transition zones of rotary cement and lime kilns. The Baker Company is the only company in the United States that produces a grain that is suitable for the manufacture of refractory dolomite bricks. Baker products are marketed both domestically and internationally.

DRIVE-THROUGH TOUR

The quarry tour begins by proceeding down the first-level ramp. The quarry face to the left of the ramp preserves some interesting features that are associated with a previous unconformity between the Ledger Formation and the overlapping sedimentary rocks of Mesozoic age. Erosion has removed the Mesozoic rocks that formerly overlay this area. The current border of the Mesozoic basin is located approximately 1000 feet to the northwest, visible as a low hill to the left at the top of the first ramp. The paleo-karst features visible to the left of the ramp appear to have been caves and possibly crevices that filled with Mesozoic sediments.

The tour will continue past a primary Universal Impact Crusher that dates to 1959. Three stockpiles of run-of-quarry dolomite are maintained near the primary crusher. Each pile contains dolomite with different impurity levels. As the tour proceeds down the second-level ramp, note the white area in the

faces directly ahead (to the east). These areas are high-calcium "marble" that strongly resembles the "white marble" found in the middle member of the Kinzers Formation. This white "marble" zone cuts across the local structure.

The tour concludes at the bottom of the fourth-level ramp where the buses will turn around and exit the quarry by the same route followed on the way in.

SOME FACTS ABOUT THE QUARRY

-- Quarry was started in 1952

-- The quarry covers approximately 70 acres

-- The quarry currently is worked on four levels, each level approximately 50 feet in height. It is planned to take the quarry down two additional levels.

-- About 900,000 tons of rocks are removed yearly.

-- Quarry areas are diamond-core drilled on a 75 foot grid pattern, and all blast hole cuttings undergo chemical analysis in order to insure only the proper quality dolomite is quarried.

-- Quarry shots typically consist of 5 to 11 holes 6" in diameter. Up to 4000 lbs of explosives are detonated during each shot, which brings down 7000 to 14,000 tons of stone. Shots are completed in the quarry 2 or 3 times per month.

-- The quarry is operated one shift per day, five days a week.

-- Quarry equipment consists of two 50-ton haulers, one 13.5 cu yd bucket loader, an excavator, and a wagon drill.

-- Reclamation: After all quarrying activity is completed on the property, the following will occur:

1. Any pits will be allowed to fill with water to form lakes.
2. Any exposed quarry faces will be sloped to 30°.
3. All berms will be leveled.
4. All buildings will be removed.
5. All disturbed areas will be covered with topsoil and seeded.

Leave J. E. Baker Company parking area. TURN LEFT on Baker Road, procede northeast.

0.4 48.4 Clay-soil stripping operation of J. E. Baker Company on right.

0.5	48.9	TURN RIGHT onto East Berlin Road (PA 234). Procede east.
0.6	49.5	Pass over limited access U.S. 30. "Dormant" West Gate Quarry (formerly Medusa Cement) visible to left.
0.7	50.2	BEAR LEFT (intersection is poorly marked) onto Bannister Street. Procede eastward past West York Area High School on right.
1.7	51.9	BEAR RIGHT at intersection with Carlisle Avenue (PA 74) at the northwest corner of the York Fairgrounds. Originally established as a semi-annual fair in 1765, the York County Agricultural Society re-established the fair as an annual early fall event in 1853. Procede southeast on Carlisle Avenue.
0.3	52.2	TURN LEFT onto Maryland Avenue opposite entrance to the York Fairgrounds on right. Procede northeast.
0.6	52.8	TURN LEFT onto Roosevelt Avenue. Procede northwest.
0.2	53.0	TURN LEFT toward parking area for Delta Carbonate (formerly York Stone and Supply) Quarry. STOP 11.

STOP 11. THE LEDGER AND KINZERS FORMATIONS IN THE DELTA CARBONATE QUARRY

Leaders: Bob Ganis and John Taylor

The busses will take us to the floor of Pit #2. Stations A and B are located in this pit. After visiting Station A, you will re-board the busses to drive to Station B. Station C, in Pit #1, will be visited if there is time, but probably will be viewed only at a distance from a vantage point near Station B.

GENERAL DESCRIPTION

The extensive exposures of the Kinzers and Ledger Formations (Figures S11-1 and S11-2) in this quarry provide an exceptional view of the complex lithofacies mosaic produced by deposition along the margin of a carbonate platform. An order-of-magnitude increase in thickness of the middle carbonate member of the Kinzers Formation, from 100 feet or less in the Lancaster area to over a thousand feet in West York Block, is one of the more dramatic lines of evidence that the strata of the Conestoga Valley represent the transition from carbonate platform to off-platform environments. Other evidence that the carbonates of the West York Block formed in shelfbreak environments include: 1) interfingering of light-colored, pure platform carbonates with dark off-shelf lithofacies containing abundant fine-grained siliciclastic sediment, 2) large blocks of dark, laminated, fine-grained carbonate that are rotated and internally deformed within the light-colored, massive platform deposits, and 3) a contrast between faunas in the deep water lithofacies, which include trilobites restricted to off-platform deposits (e.g. the trilobite genus Pagetides), and algal boundstones in the platform carbonate lithofacies.

Figure S11-1. Photograph of Delta Carbonate, Inc., Pit 2, looking north. Є1wr=Willis Run Mbr. of the Ledger Fm., Є11d=Lower Dolomite Mbr. of the Ledger Fm., Єkg=Greenmount Mbr. of the Kinzers Fm.

GENERAL STRATIGRAPHIC SECTION FOR DELTA CARBONATE (FORMERLY YORK STONE & SUPPLY)

		THICKNESS	DESCRIPTION	EXPOSURE
LEDGER FORMATION	Upper Dolomite Member	not measured	coarse crystalline, highly fractured, grey to light brown, oolitic, v. pure dolomite	occurs just north of operation
	Willis Run Member	210	fine to med. grained; argillaceous, black, platey bedded limestone	exposed in Pit 2
	Lower Dolomite Member	200	coarse crystalline, highly fractured, grey to light brown, oolitic, pure dolomite	exposed in Pit 1
KINZERS FORMATION	Greenmount Member	0–50	impure limestone, weathering to a shaley or sandy appearance; dark grey when fresh; partly laminated, fine and med. gr. massive bedded	
	York Member	1000–1200	top 200' is impure grey, oolitic; remaining section highly variable containing 10–100' beds of white limestone, colored limestone, grey limestone, some pure, some impure	
	Emigsville Member	200	greenish phyllite; weathers brown	not exposed; penetrated by drilling

Figure S11-2.

A geologic map (Figure S11-3) is provided for the area surrounding the quarry. (The aerial photo used as the frontispiece of the guidebook covers about the same area). The structure of Pit #2 is monoclinal, consisting of massive bedded units dipping from 15 to 20 degrees to the southwest, truncated by high-angle transverse faults. Between Stations B and C, a panoramic view of Pit #1 will provide a view of the broad, open syncline in which the pit has been opened. This exposure is the best and largest for the York Member within the West York Block. The mildly deformed nature of the West York Block is well illustrated at this stop.

STATION A

An opportunity is provided here to examine the contact between the Kinzers and Ledger Formations as it appears in the West York Block. The thin (30 feet) interval of dark, somewhat nodular limestone below the contact is the Greenmount Member, an impure off-platform lithofacies that has yielded the characteristic off-platform eodiscid trilobite *Pagetides*. The Greenmount Member at this locality has, in fact, yielded two distinct trilobite faunas that may prove useful in correlation within and beyond the Conestoga Valley (see Chapter IX). The high concentration of fine-grained siliciclastics within this member confirms that it is a distinct sedimentary package, rather than an interval differing only in diagenetic history from the units below (York Member of the Kinzers) and above (Lower Dolomite Member of the Ledger). Also common in parts of the Greenmount are clusters of quartz crystals that display euhedral growth lamellae throughout, suggesting an entirely authigenic origin. They are, however, hollow and appear to be concentrated in portions of the rock with other bioclasts, suggesting that they originally were small, siliceous sponge spicules. Comments on the origin of these crystals would be very welcome.

STATION B

The limestones in this part of the quarry represent the middle member of the Ledger Formation, the Willis Run Member. In this area the Willis Run is essentially monofacial, consisting of moderately bioturbate, thinly bedded, lime mudstone to wackestone with abundant dolomitic laminae that are disrupted to varying degrees by burrowing and compaction. It includes no well-winnowed lime grainstones to suggest the influence of wave action even during storms. The limestone beds do not, however, display the characteristic normal grading or parallel laminations characteristic of limestone turbidites; the three-dimensional pattern of burrowing is also indicative of platform, rather than off-shelf, deposition. This facies of the Willis Run is interpreted as a deep platform deposit that accumulated below storm wave base in an intrashelf basin. Still deeper conditions are suggested by a thin siliceous, pyritic seam visible here as a reentrant in the quarry wall. Note that the limestones immediately above the seam are highly organic and grade upward

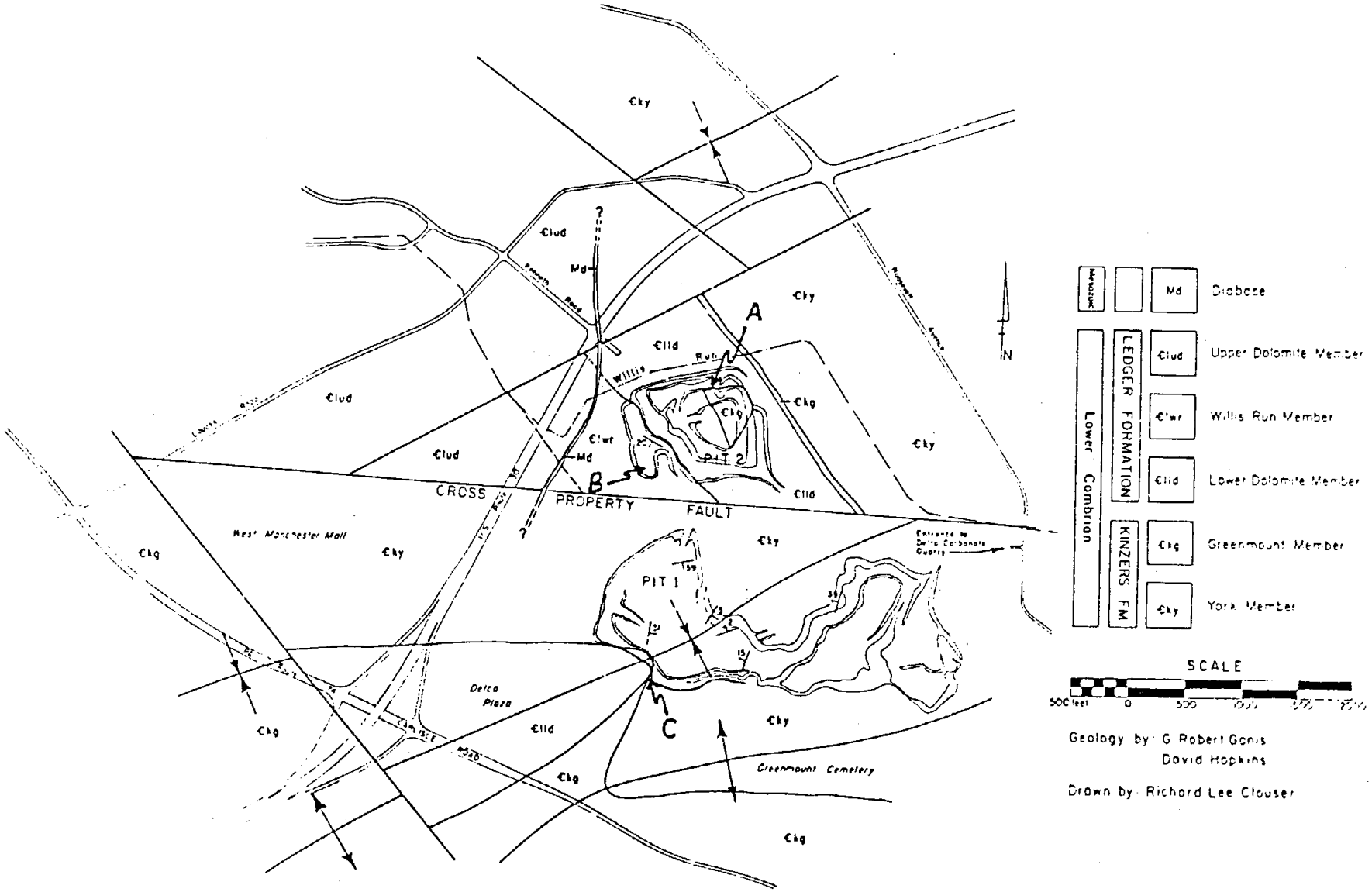


Figure S11-3. Geologic map of the Delta Carbonate quarry and vicinity, showing locations of Stations A, B and C.

into the more typical bioturbate facies of the Willis Run at this location.

Elsewhere, the member includes a much higher percentage of coarse-grained, well-winnowed shallow water lithologies such as oolitic, pisolitic, and bioclastic lime grainstone. The same exposures provide the only example of well-preserved algal reef facies in the West York Block. The reef facies consists of stromatolitic boundstone with shelter voids that are lined by botryoidal fibrous marine cements and floored by laminated internal sediment. Fossils are common in the Willis Run Member, weathering in relief from the dolomitic laminae on bedding plane exposures. Trilobites, brachiopods, and other elements of the fauna collected from such exposures establish a Lower Cambrian age for this unit and the underlying Lower Dolomite Member.

STATION C

In this area of the Delta Carbonate Quarry, the Greenmount Member is missing and the Lower Dolomite Member of the Ledger rests directly atop light-colored bioclastic lime grainstones assigned to the York Member of the Kinzers. However, within the massive dolomites at the base of the Ledger occur rotated and internally deformed blocks (Figure S11-4) of dark, laminated, impure limestone (Greenmount lithology) suggesting that the Greenmount Member was deposited but subsequently disrupted by slumping of the unconsolidated platform margin sediments. The loose carbonate sands of the overlying Ledger would have moved easily by grain flow down the slope with no internal features to record the transport. In contrast, the cohesive, fine-grained sediments of the Greenmount Member separated as oversized sedimentary boudinage that were rotated and deformed as they were carried along in the unconsolidated sands.

ECONOMIC GEOLOGY

This is quarry number 4 of the list in Chapter X. Both Delta Carbonate (a subsidiary of Millington Quarry, Inc.) and York Building Products Company produce construction aggregate at this site. The underground mine, visible in the panoramic view of Pit#1, was opened to follow a bed of particularly pure white limestone in the York Member that is desirable as whiting material (inert filler that imparts no color). The underground operation is not active at the present time. It is interesting to speculate on possible use of the very dark-colored limestone of the Willis Run Member as filler in products that are intended to be black, such as tires.

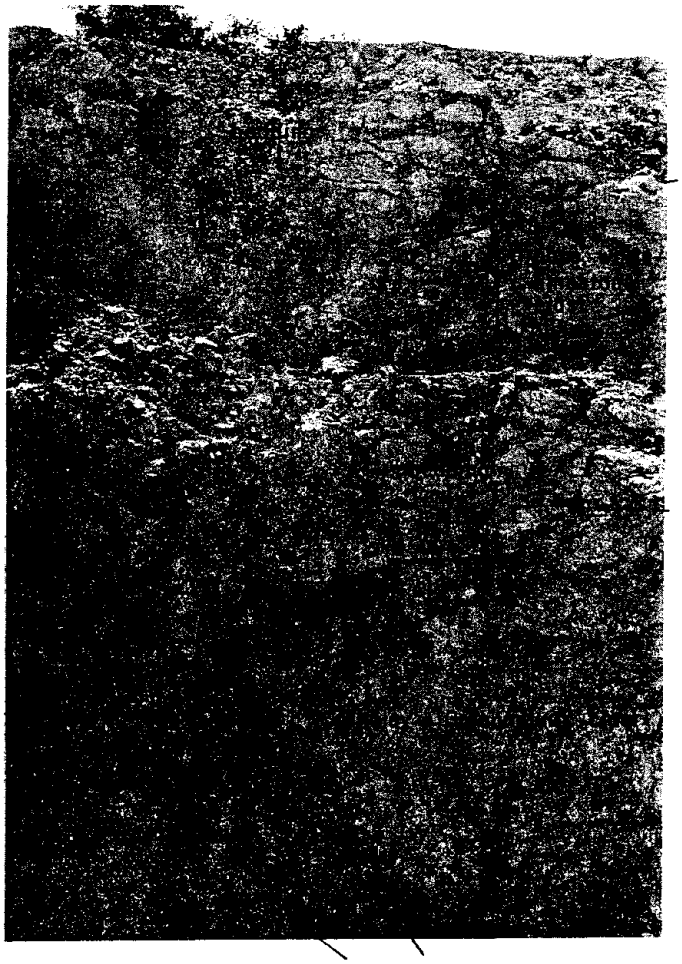


Figure S11-4. Gravity slide megabreccia exposed on the south face in the west end of Pit 1, Delta Carbonate, Inc. Both pure limestone and "shaley" carbonate (which resembles the Greenmount Member) have been incorporated into the Ledger Formation. Height of exposure is about 20 meters (65 feet).

		Return to Roosevelt Avenue, TURN LEFT and procede northwest.
0.8	53.8	TURN RIGHT onto U.S. 30. Procede east following U.S. 30 to Lancaster.
1.6	55.4	Pass under I-83.
1.9	57.3	Cross North Hills Road. U.S. 30 resumes as limited access highway. Continue straight ahead toward east.
3.5	60.8	"Shoe House" on right.
6.4	67.2	Begin crossing the Wright's Ferry Bridge over the Susquehanna River.
1.0	68.2	Lancaster County end of Wright's Ferry Bridge.
8.8	77.0	Rohrerstown Road exit to the right for Millersville University. Continue straight ahead on U.S. 30 East.

0.9	77.9	EXIT RIGHT onto ramp for Harrisburg Pike. Pass this morning's Stop 6 to right, on exit ramp.
0.2	78.1	TURN RIGHT onto Harrisburg Pike. Procede southeast toward Lancaster City. Longs Park on left.
0.5	78.6	Pass under Conrail railroad tracks.
0.5	79.1	Cross President Avenue. Continue straight ahead toward the southeast on Harrisburg Avenue.
0.3	79.4	Passing campus of Franklin and Marshall College on right.
0.7	80.1	End of Harrisburg Avenue. TURN RIGHT onto Water Street, then TURN LEFT after one block onto Lemon Street.
0.1	80.2	TURN RIGHT onto Prince Street. Procede south.
0.4	80.6	TURN LEFT onto Chestnut Street. Procede east.
0.2	80.8	Cross Queen Street. Chestnut Street entrance to Brunswick Hotel on right.

END OF FIELD TRIP. HAVE A SAFE JOURNEY HOME.