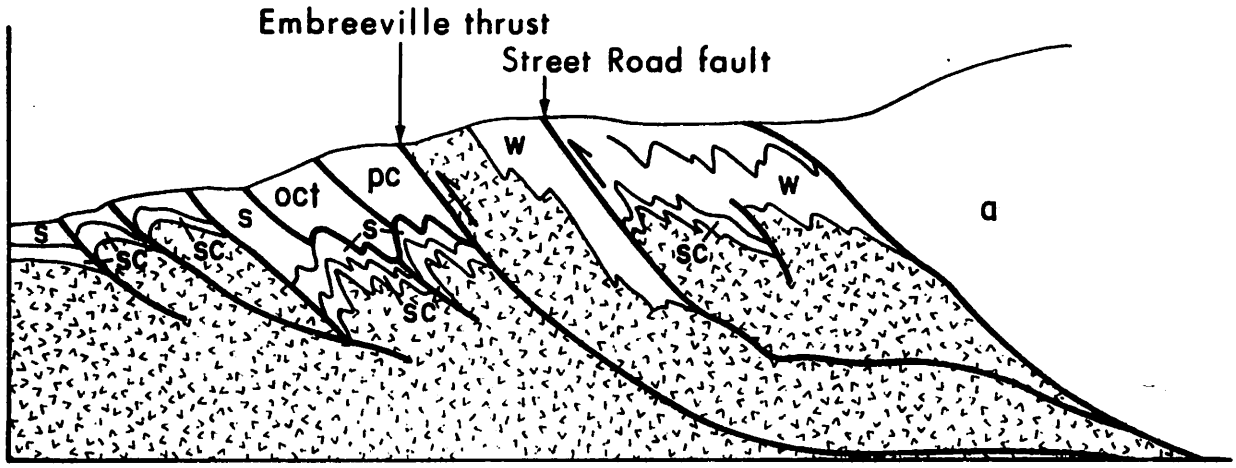
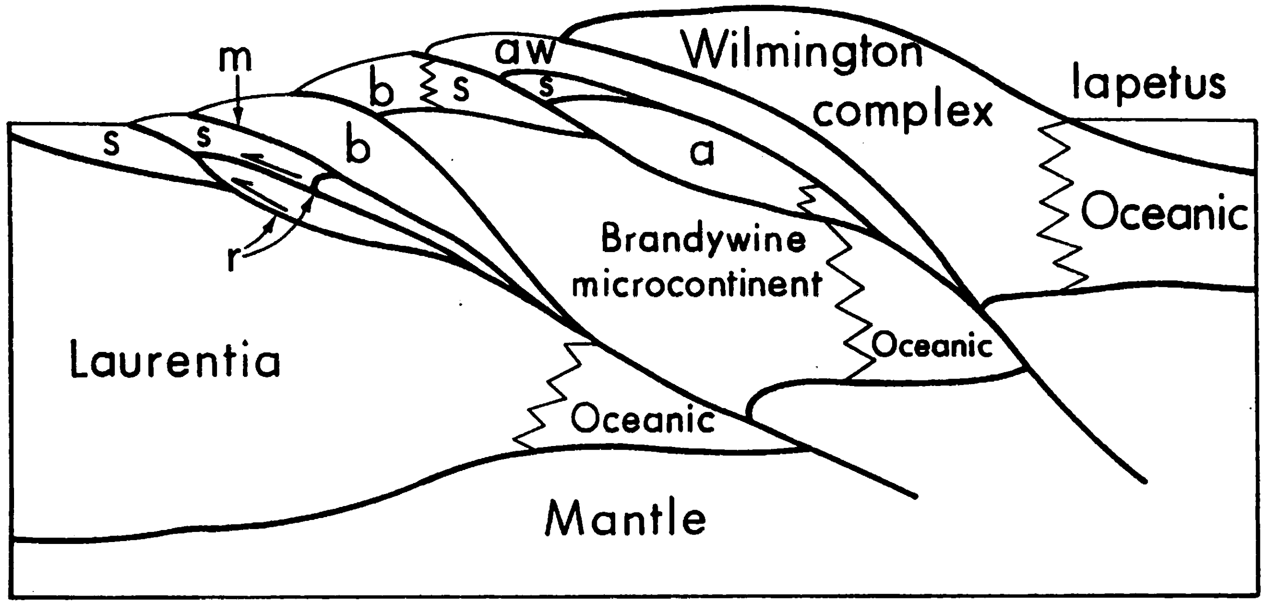
GUIDEBOOK

59th Annual Field Conference of Pennsylvania Geologists

Various Aspects of Piedmont Geology in Lancaster and Chester Counties, Pennsylvania





Hosts: Pennsylvania Geological Survey
West Chester University
Concord College

September 29 and 30,
and October 1, 1994
Lancaster, Pa.

Guidebook for the

59th ANNUAL FIELD CONFERENCE OF PENNSYLVANIA GEOLOGISTS

VARIOUS ASPECTS OF PIEDMONT GEOLOGY IN LANCASTER AND CHESTER COUNTIES, PENNSYLVANIA

Rodger T. Faill and W. D. Sevon, Editors
Field trip leaders and contributors
Samuel W. Berkheiser, Jr.
Alexander E. Gates
David B. MacLachlan
Robert C. Smith, II
Glenn H. Thompson
David Valentino
C. Gil Wiswall
Additional contributors
John H. Barnes
Thomas W. Gardner
Frank J. Pazzaglia

September 29 - October 1, 1994

Hosts: Pennsylvania Geological Survey
West Chester University
Concord College

Headquarters motel: Holiday Inn, Lancaster, PA

Cover: These cross sections illustrate two different interpretations of the Taconian assembly of the Piedmont of southeastern Pennsylvania (Faill and Wiswall, this guidebook, Figures 2D and 3C, p. 75 and 77).

Artwork: Jack Kuchinski

Guidebooks distributed by:
Field Conference of Pennsylvania Geologists, Inc.
P. O. Box 5871
Harrisburg, PA 17110-0871

Frontispiece. 1993 Field Conference of Pennsylvania Geologists group photograph at Rocky Gap State Park, MD.

TABLE OF CONTENTS

LIST OF FIGURES

LIST OF TABLES

VARIOUS ASPECT OF THE PIEDMONT GEOLOGY IN LANCASTER AND CHESTER COUNTIES, PENNSYLVANIA

Rodger T. Faill
Pennsylvania Geological Survey
P. O. Box 8453, Harrisburg, PA 17105-8453

INTRODUCTION

Southeastern Pennsylvania holds a plethora of mysteries about its diverse tectonic, sedimentologic, igneous, and metamorphic events, many of which relate to one another. Evidence of these mysteries is recorded (and disguised) in the Pennsylvania Piedmont, a 250 km-long belt that reaches from Adams County west of the Susquehanna River to Bucks County at the Delaware River. This year's Field Conference of Pennsylvania Geologists will examine various aspects of the Piedmont geology in Lancaster and Chester Counties, Pennsylvania.

The evolution of the Piedmont began with the formation of the Laurentian passive margin. The Neoproterozoic supercontinent Rodinia broke apart at the end of the Precambrian, separating (among other parts) eastern Laurentia from southwestern South America (present geographic facings)--between them the Early Paleozoic ocean Iapetus was born. The Laurentian passive margin developed over the transition between continental and oceanic crust.

An intracontinental rift formed near the margin during the latest Neoproterozoic, in which a large volume of Catoctin rhyolite and basalt accumulated. These chemically distinctive rocks were part of a very widespread volcanism and intrusion that permeated this part of Laurentia in the early rifting of Rodinia. Subsequently (in the Early Cambrian), this Catoctin rift and the rest of the Laurentian eastern margin was covered by terrigenous sediment (Chilhowee Group) derived from the Laurentian craton to the west. Some of this sediment passed over the margin onto the continental slope and perhaps even into the adjacent oceanic basin.

A marked decrease of terrigenous material reaching the margin late in the Early Cambrian allowed the development of a carbonate bank that would grow into a rimmed carbonate shelf hundreds of kilometers wide (MacLachlan, this guidebook). This shelf persisted for more than 100 m.y. into the early Late Ordovician. The early shelf rim (Early to Middle Cambrian) extends through Lancaster and Chester Counties, providing us an opportunity to see the shelf-rim facies (Ledger Formation oolites) and the upper slope facies (Conestoga Formation). These will be seen at Stops 1-4.

The argillaceous carbonates of the upper continental slope graded downslope into the basinal pelites and psammites, represented today by the Octoraro and Peters Creek Formations. The question of whether this basin was the western part of Iapetus or a basin (possibly an embayment) separated from Iapetus by microcontinents is moot. It devolves on whether the Brandywine massifs to the southeast (West Chester, Avondale, and Woodville) were originally part of Laurentia or came from a microcontinent (Brandywine) separated from Laurentia by an oceanic basin (Octoraro seaway).

Several observations support the microcontinent model: (1) tectonically emplaced ultramafic bodies on the northwest side of the West Chester massif probably had an oceanic crust origin; (2) metabasalts in the Peters Creek Formation have ocean-floor basalt protoliths, and chemically are not at all like the Laurentian Catoctin volcanics (Smith and Barnes, this guidebook); (3) lithically, the Brandywine gneisses are quite different from those of the Honey Brook Upland, the Trenton Prong, and the Reading Prong, all of which are similar to the gneisses of the Adirondack Highland terrane of the Grenville orogen; and (4) all three Brandywine massifs lack intrusive rocks of Catoctin affinity, which are common in the Grenville-age gneisses to the northwest (Smith and Barnes, this guidebook).

Regardless, the phyllites and schists of the Octoraro and Peters Creek Formations had sedimentary protoliths that were deposited on attenuated continental to oceanic crust. Although the Peters Creek now lies southeast of the Octoraro, the original relation between the two is not certain. The provenance of the feldspathic quartzose schists of the Peters Creek may have been Laurentia, to the northwest, or South America (or a microcontinent), to the southeast. A Laurentian source either would have required the transport of the coarse sediment across the broad carbonate shelf, slope, and Octoraro seaway during the Early Paleozoic, or, to avoid this, would restrict the Peters Creek deposition to the Early Cambrian (pre-carbonate shelf), correlative only with the Chilhowee Group. A southeastern source would only need to be continental, requiring no time constriction on Peters Creek deposition within the Early Paleozoic. The Octoraro Formation will be seen at Stops 8, 9, and 10 (Valentino: Octoraro Formation, this guidebook); the Peters Creek Formation will be visited at Stops 6, 7, 12, 14, and 15 (Valentino and Gates, this guidebook).

The metabasalts in the Peters Creek Formation constitute a minute fraction of the rock volume, but their significance is great. Their protoliths are basalt flows, so they are excellent indicators of bedding orientation. In addition, their chemistry is distinct from that of the widespread Laurentian Catoctin volcanics and intrusives, which indicates the metabasalts originated in a quite different tectonic setting (Smith and Barnes, this guidebook). Metabasalts will be seen at Stops 7, 11, and 14.

The Peach Bottom slate and adjacent Cardiff conglomerate are distinctive units within the Piedmont. The Peach Bottom structure they occupy has traditionally been thought to be a syncline, which would make the two formations the youngest in the Glenarm Supergroup. However, if the structure is an anticline (some evidence for this), then the Peach Bottom slate is the oldest unit, succeeded by the Cardiff, Peters Creek, and Octoraro Formations going northward, or by the Cardiff, Peters Creek, "Glenarm Wissahickon", Cockeysville, and Setters Formations to the east. More likely, the Piedmont that is underlain by the Glenarm Supergroup is an *assemblage* of nappes and thrust blocks, and no coherent encompassing stratigraphy can be discerned. The presence of ultramafic layers on either side of the Peach Bottom structure may indicate that tectonic emplacement of structural blocks was a significant factor and that we may be seeing a melange (Smith and Barnes, this guidebook). Chemically, the protolith of the Peach Bottom slate was a shale (Smith and Barnes, this guidebook); yet the strong shearing of the adjacent Cardiff conglomerate suggests that the dark color and very fine grain size of the slate may have been enhanced by transpressional shear (Valentino: Peach Bottom Problem, this guidebook). These, and other aspects, will be seen at Stop 14.

The metamorphosed siliciclastic rocks of the Pennsylvania Piedmont can be divided into two belts, each of which has a different metamorphic pattern and presumably a different origin for the metamorphism. The northwest belt centers on the Tucquan antiform and appears to have been heated primarily from below; the southeast belt lies southeast of the Pleasant Grove/Huntingdon Valley shear zone and appears to have been heated primarily from a hot source above. These differences will not be demonstrated in the Field Conference stops, but are discussed by Valentino and Faill (this guidebook).

The latest movements on the subvertical Cream Valley fault were strike-slip, but the juxtaposition of high-grade Grenville gneiss of the West Chester massif with low-grade metasediments suggest that a large vertical displacement must also have occurred on an early Embreeville thrust fault, the precursor to the Cream Valley fault (Faill and Wiswall, this guidebook). The fabrics associated with both movements will be seen at Stops 6 and 7.

Regional shear zones have only recently been recognized in the Pennsylvania Piedmont, and they are prominently featured in this Field Conference. These wide (up to 1 km) zones are characterized by fine grain size (resulting from the shearing) and numerous rotational features that indicate dextral strike-slip regional transpression (Valentino: Peach Bottom Structure, this guidebook). They appear to transect all other structures in the Piedmont, and thus are probably of late Alleghanian age. One such zone occurs in the Conestoga Formation at Stop 4. The Drumore zone is discussed at Stop 9 and 12; and the Peach Bottom zone, at Stop 14.

If all these bedrock mysteries leave you bewildered, you can always kick the dirt. Saprolite and colluvial deposits, combined with the land forms of this part of the Piedmont, reveal a great deal about how old this Piedmont landscape is, and how it probably formed (Sevon, this guidebook). In particular, notice the accordancy of hilltops on the drive between Stops 14 and 15, as discussed in the road log.

If you kick the dirt in the field where we will disembark for Stop 14, you may uncover a "fossil potato". These "potatoes" are large pebbles of quartz that were deposited by the Susquehanna River millions of years ago and now are found as remanant terrace deposits (Pazzaglia, this guidebook). Note where we stand relative to the present Susquehanna River.

These are but some of the various aspects of the Piedmont geology in Lancaster and Chester Counties. The Field Conference of Pennsylvania Geologists hopes that you find this trip stimulating and enlightening, and that you carry away with you an appreciation of this wonderfully complex and exciting geology.

SOME ASPECTS OF THE LOWER PALEOZOIC LAURENTIAN MARGIN AND SLOPE IN SOUTHEASTERN PENNSYLVANIA

David B. MacLachlan
Pennsylvania Geological Survey
P. O. Box 8453, Harrisburg, PA 17105-8453

"... she gave all subsequent Appalachian geologists a basic picture to modify and perfect - a picture which was largely blank when she began her work" - GSA memorial to Anna Isabel Jonas Stose, 1881-1974

INTRODUCTION

In 1966 or 1967 Don Wise told me that John Rodgers came through asking for guidance to some good exposures of carbonates in the Lancaster Valley. He said Rodgers claimed that the Conestoga Limestone was the down-slope facies of the adjacent platform margin carbonates. This was welcome news. I had found the supposed great unconformity beneath the Conestoga Limestone irreconcilable with my perception of what must be occurring seaward of my bailiwick in the Taconic foreland nappes of the Great Valley. The great regional synthesis that Rodgers (1968) produced was scarcely dependent on the confirmatory detail he acquired here, but it is a benchmark paper in Pennsylvania geology - it delivered rocks whose names we knew and some of us should know something about into the embrace of "the new global tectonics". At the 35th Annual Field Conference (Crowley and others, 1970) it was demonstrated that it was possible to bring some modern sense into another package of familiarly named but poorly understood rocks. They solved part of this problem by renaming most of the rocks, and the Maryland Geological survey has subsequently abandoned use of the term Wissahickon - right on!

In the latter part of the 1980's the Pennsylvania Geologic Survey concluded that the implications of these observations had matured sufficiently that it might be a prudent investment of some tax money to reexamine this most populous part of the state about which we knew least. An Eastern Regional Section of the Geologic Mapping Division was organized composed of Rodger Faill, Bill Sevon, myself, and a vacancy - to be filled for a few stimulating years by Dave Valentino. During my first encounters with Stose and Jonas' limited work in the Great Valley east of the Susquehanna I was both awed and irritated. How could they be so nearly right, in an intellectual environment obviously hostile to their ideas, as reflected in a number of exchanges in the contemporary literature, and still be wrong. Entering their home turf I decided to proceed with circumspection. In the (mostly) conspicuously stratified rocks of the northwestern Piedmont there was some reason to wonder if they might not be substantially correct in many particulars. Armed with a radically different concept of the genesis of eutectonic orogenic belts and a hasty brush-up on modern principles of sedimentation I resolved, insofar as possible, to reexamine the rocks they had mapped unbiased by the work of subsequent writers about these particular rocks.

This paper is only a progress report on the regionally significant aspects of these investigations. Nor can the four localities presented for your inspection on the field trip hardly do entire justice to the full flavor of my impressions.

THE LEGACY

In 1922 George Stose and Anna Jonas published on the Lower Paleozoic section of southeastern Pennsylvania. This was followed in 1923 by a hypothesis of Ordovician overlap (Conestoga Limestone). The latter paper I was tempted to characterize as a mistake, but in 1960, the year G.W. Stose died at 90 and I joined the Pennsylvania Geologic Survey, the text books had yet to be rewritten . They had recognized an anomalous stratigraphic relationship and given the best explanation consistent with the prevailing paradigm - good enough work. These papers document the stratigraphic framework of an interval of intense geologic reconnaissance and mapping which was sufficiently advanced that the 1931 Geologic Map of Pennsylvania (Stose and Ljungstedt, 1931), southeast of the Lower Mesozoic basins, is essentially identical in most particulars to the 1960 edition (Gray and Shepps, eds., 1960). Most of the dozen 15-minute quadrangles and two partially redundant county reports including this predominantly carbonate section, all of which bear the name of one or both of these authors and occasional co-workers (sometimes hereinafter collectively identified as Jonas, Stose and Co.), appeared after the 1931 map was compiled, the last in 1944 (A.J. and G.W. Stose, 1944). However, it is apparent that the conceptual framework and an amazing portion of the supporting detail had been established in the 1920's before the 1931 State Geologic map was prepared.

Of the few emendations to the product of Jonas, Stose and Co. appearing in subsequent editions of the State Geologic Map only two merit particular comment. The 1960 map (Gray and Shepps, 1960) introduced changes in the Martic Line area of the McCalls Ferry - Quarryville district (Knopf and Jonas, 1929), in part from information submitted by D. U. Wise at compilation time, but more significantly from Cloos and Heitanen (1941). The salient point, which may be intentionally obscured by artful color selection (get your hand lens and look at the thin pale orange band west of Quarryville) is that there is an arbitrary cut-off between "Antietam-Harpers" and "Wissahickon". I address this matter further in discussing the Chilhowee Group below, but in summary: my final score in the great Martic controversy - Knopf and Jonas 3, Cloos and Heitanen 2, and I am going to have to change both the map and the State Correlation Chart (Berg and others, 1983).

In 1971 Meisler and Becher established that Jonas and Stose (1930) had made a significant stratigraphic *faux pas* in the Lancaster (and adjacent) quadrangles. The northern part of the Lancaster County carbonate belt had to be completely recompiled for the 1980 Geologic Map of Pennsylvania (Berg and others, 1980). Fortunately, senior geology students at Franklin and Marshall College had been mapping Meisler and Becher's (1968) stratigraphy in the area almost since the paper appeared. South of Meisler's Mechanicsville thrust, however, Jonas and Stose's contacts required little adjustment. Simply put, the rocks they designated Conococheague and Elbrook in superposition order should be called Elbrook and Zooks Corner. Where the former names recur in the eastern Chester Valley, I find they were also mapped fairly consistently with that adjustment except in the Coatesville-Downington area. A more subtle change from 1960 to 1980 is the deletion or truncation of a number of faults, some of which will be restored with augmentation (Hill, 1989; MacLachlan, 1990; Valentino, *1990; ibid.* and others, 1994) on the Millennial State Geologic Map (Godknows and Politics, not in prep.). The working proposition was apparently that Jonas and Stose mapped too many faults. This is locally true, as in the Lancaster area, but my overall impression is that they are more to be "faulted" for misinterpreting some of the faults they did map and failing to map a few that do have significant impact on the accuracy of their mapping in some areas.

The period reflects the limitations under which this herculean task was performed. Geologic formations were conceived as time bound events often with little known of their depositional environment beyond marine or terrestrial. Deep water deposits, perhaps especially carbonates, were notably poorly known and understood. The perceived wisdom at the U. S. Geological Survey was generally hostile to low angle thrust faulting. Stose and Jonas (1927) and Jonas (1929) had already made the first of a number of observations which were later to embroil them in considerable controversy on the side of the angels in this matter. The last cited paper elevated the Taconic "disturbance" of New York to a major orogeny impacting all of the eastern part of the Appalachian disturbed zone, and tends to establish Anna as the real idea "man" in Jonas, Stose & Co. It is thus doubly ironic that their first joint effort was fatally flawed by their failure to recognize that they were missing section under one of their faults.

The New Holland and Lancaster quadrangles (Jonas and Stose, 1926; 1930) were among the first completed maps. These two contain much of the most complete, unambiguously Lower Paleozoic section anywhere in the Piedmont, and the type sections or areas of five new formations they erected in 1922. The Vintage, Kinzers, Ledger, Conestoga, and one misnamed unit - "Elbrook" [=Zooks Corner (Meisler and Becher, 1968)] formations represent aspects of the Lower and, at least in part, Middle Cambrian carbonate shelf margin and slope. The Cocalico Shale of the northern part of the map area is the local expression of the Upper Ordovician flysch basin. It apparently contains both autochthonous Martinsburg equivalents and Taconic elements. If this unit is adequately remapped and subdivided the name might be restricted or replaced, but it currently stands as a vast improvement over the Second Survey map (Smith and Lesley, 1893) which included these rocks in the Mesozoic basin.

The Chilhowee Group appears beneath the bank margin carbonates. For the lowest formation they adopted the local name Chickies Quartzite, used by the Second Pennsylvania Geological Survey in an extended sense for the "Primal sandstones" of southeastern Pennsylvania (Lesley, 1892, Ch. 16), though the name was first used specifically for the light "scolithus" bearing quartzites at Chickies Rock as distinct from the associated "hydromica schists". (Skolithos is "scolithus" bearing quartzites at Chickies Rock as distinct from the associated "hydromica schists". (Skolithos is now the approved spelling, but I shall herein-after use the prior form without the embarrassment of quotes, as it so appears in most of the relevant descriptions.) It was subsequently applied by Jonas (1905; Bliss and Jonas, 1916) to metasediments of the Doe Run area, subsequently assigned to the Setters Formation (Bascomb and Stose, 1932). No specific type section is given for the type area in the Middletown 15-minute quadrangle (Stose and Jonas, 1933) where "quartz schist" is (barely) mentioned. The often visited section of the Chickies anticline along the railway (Wise and Kauffman, 1960) is a superb structural exposure but a poor sample of the abundant, less-quartzitic rocks. A salutary corrective is the traverse along PA Route 441 on the north flank of Chickies Ridge up to the parking lot near the crest. Those seeking to provoke students may profitably continue the traverse down the south flank to locate the Harpers/Antietam contact (in fresh road cuts) approximately where mapped by Stose and Jonas. It is a real challenge (look for bedding beginning to be resolvable from the cleavage). My reconnaissance impression is that, barring exceptional exposure, the contact is not really mappable in this area. Jonas and Stose (1926) had an easier task on the west edge of the Honey Brook upland in the New Holland 15-minute quadrangle, where I had no doubt that I could see what they were mapping.

The remaining parts of the section were ascribed to units presumed correlative to the section previously described by Stose (1906, 1908) in Franklin County. Unfortunately they failed to realize that their Lancaster County section was incomplete. This resulted in compressing the entire Franklin County section into the Lancaster partial section, largely invalidating their proposed correlations. It is not true that Jonas and Stose (1930) completely failed to recognize the Mechanicsville thrust in the Lancaster quadrangle (Meisler and Becher, 1971), subsequently called Oregon thrust by some Pennsylvania Geologic Survey members (e.g. Faill and MacLachlan, 1989; 1991) in informal publications. They mapped a mosaic of normal faults approximately coincident with the trace of the Mechanicsville thrust as mapped by Meisler and Becher, but the crop of an Alleghanian decollement was more than they were prepared to deal with. They came a lot closer (Stose and Jonas, 1939) in a regional sketch map for the York County report on which the Chickies thrust is more-or-less correctly projected into the Mechanicsville thrust as dashed line separating the Lancaster-York valley from the Great Valley (which is essentially correct except in the eyes of regional physiographers). If they had had the courage of their convictions to add the barbs, they would have been in a position to redeem their early error; but they were too old with too many other irons in the fire.

Considering the quality of the available base maps, the absence of air photos, and the great speed with which this large area was mapped by relatively few people, it is much to the credit of Jonas, Stose & Co. that, with appropriate stratigraphic reinterpretation, much of the area mapped by them remains sufficient for 1:250,000-scale presentation. Except for the area of the Philadelphia Folio (Bascom and others, 1909) which was hastily patched to accommodate the stratigraphy adopted for the 1931 geologic map and for a few areas where they failed to recognize significant faults, the general picture they presented stands. If this were their sole product it would be respectable. Considering all the other matters with which they concerned themselves concurrently and subsequently, it is truly remarkable. The subsequent sections necessarily involve supplementing their observations with information to which they had no access and occasionally carping about things they might have done better. I do so with respect. Few, if any, who see this guidebook will have the opportunity to contribute as much to their profession as they did.

TECTONIC FRAMEWORK OF THE SHELF MARGIN

Iapetan Rifting. The Late Neoproterozoic rifting of Rodinia (Bond and others, 1984) yielding Laurentia and other cratonic fragments with allegedly Grenvillian margins, now widely scattered around the world according to numerous recent authors (e.g. Powell and others, 1993), is a fitting end to an era. The general character of this event for the Atlantic margin of Laurentia is summarized (Rankin and others, 1989), with specific reference to the Ledger-Conestoga contact as approximating the main breakup rift margin in this area, as well as regional characteristics of the Catoctin rift volcanics which is augmented by new details for the Pennsylvania area by Smith and others (1991, Smith and Barnes, this guidebook). This spate of papers, global, regional, and provincial, shows that a new era of geological science is fully established. It depicts a world unimaginable to the generation of geologists who largely defined how our Piedmont is portrayed today. In the light of radically different concepts of how such things came into being, the picture they left us requires considerable modification and perfection. But as long as people find information about the rocks they trod useful or interesting, many of their contributions will persist.

MacLachlan (1991) concluded the main body of Pennsylvania's South Mountain with its rhyolite dominated volcanics and medial shelf cover was distinct from the Blue Ridge trend extending from South Mountain (Maryland) via the Pigeon Hills to the Honey Brook upland. I more speculatively projected the obvious dividing fault into Lancaster County. I now deduce that the rhyolite dominated Catoctin of the former occupies a failed rift extending northwestward from a triple point where the Blue Ridge trend changes. The failed rift, in segmenting the Laurentian basement, appears to have significant impact on subsequent tectonic events. The relation between the two South Mountains has been forshortened and obscured by Paleozoic thrusting (Faill and MacLachlan, 1991).

Smith and Barnes (this guidebook) give the basalt at Pigeon Hills a distinctly oceanic character. Structurally continuous cover of northern Piedmont type runs up to bank margin carbonates in the West York area. The Jurassic reefs of the Atlantic rift margin are similarly situated (Sheridan and Grow, 1988). The slice between the Glades and Highmont thrusts in the Hellam Hills, bottoming in Loudoun Formation (R.C. Smith, personal communication) and metarhyolite, appears to be a horse of western South Mountain affinity entrained in the Chickies thrust system.

The crystalline rocks of Reading Prong and the Honey Brook uplift both have metadiabase dikes that are probably more extensive than mapped. Every one who has worked these areas has bemoaned the paucity of outcrop with more than customary vigor. I know from personal experience how easy it is to lose a fairly thick Jurassic dike, which was mapped (discontinuously - it probably is not) in the Honey Brook upland (Bascom and Stose, 1938), when attempting to trace it northward from the Chester Valley. The Catoctin affinity of these dikes has been confirmed chemically by Smith and Barnes (this guidebook).

The Honey Brook upland is known to have had a volcanic cover from rare (confirmed - Smith, personal communication) occurrences of debris at the base of Chickies Formation. As this contact is ordinarily colluviated, they may be more common than we know. He (Smith) also suggests that the Hexenkopf complex of the Easton quadrangle may be associated with this volcanic cycle.

Modeling the development of the numerous known Lower Mesozoic rift basins occurring seaward of the Newark-Gettysburg basin and its kin under the Coastal Plane deposits and the continental shelf results in the conclusion that the basins are generated by listric extension shears which descend to a basal detachment, presumably at the base of the sialic crust. During the extension, the mantle stretches plastically while the crust is thinned in a more brittle way by the cumulative effect of the extensional ramps. Simply, approximately vertical extensional ruptures provide access for basaltic magma forming dikes, sills, and flows which increase the density of the crust. The combination of these effects produces the "transitional crust" of geophysical profiles in the neighborhood of the breakup line.

These observations suggest that, similar to the more recent Atlantic rifting, the Iapetan rifting produced a number of basins inboard of the breakup. The associated extensional ramps are an ideal locus for reactivation as thrusts in subsequent compressional orogenies. In short, the Iapetan extension predefined the root zones of the Laurentian cored, parautochthonous Taconic nappes of the Ordovician shelf, and probably influenced low angle Alleghanian thrusts and the Lower Mesozoic rifting.

Drift stage - breakup, thermal rebound, and breakup unconformity.

The southwesternmost shelf rocks having a probable Taconic fabric appear near the west margin of the Gettysburg basin beneath Triassic red beds in northernmost Adams County. *Not* coincidently I think, they have about the same position along strike as the western limit of Hamburg klippe related rocks in the Martinsburg Formation of Cumberland County (Root and MacLachlan, 1978). Beekmantown Group rocks appear in a comparable position near Fairfield in southern Adams County. These show no evidence of Taconic tectonism and appear to be dropped from the cover of South Mountain (Stose and Bascomb, 1929). A subtidal, probably fairly deep (R. B. Neuman, personal communication) equivalent of the latter rocks, appears above upward-shallowing Cambrian slope-facies rocks as the Grove Limestone (Stose and Jonas, 1936; Reinhardt, 1974) in the Frederick Valley. The Grove is also without Taconic fabric and is probably separated structurally from the Fairfield area by only one substantial Alleghanian thrust deduced as precursor of the fault bounding the western margin of the Gettysburg and Culpepper basins in Maryland. The contrast of this situation with the belt of Taconic nappes with shelf facies cover, for which I estimate a minimum width of about 30 miles prior to Alleghanian shortening, lying seaward of Cumberland Valley rocks of the western South Mountain, is striking. It seems unlikely that the two known Alleghanian thrusts, The Antietam Cove (MacLachlan, 1991) and "Blue Ridge Summit" (MacLachlan, 1993, new name; locality mentioned by Smith and Barns, this Guidebook, in relation to the Pigeon Hills) appearing between the Fairfield rocks and the western South Mountain blocks has sufficient displacement to conceal the Taconic nappes.

These discrepancies are readily resolved if the Iapetan breakup utilized the failed rift as a local transform to step cratonward toward the east flank of Catoctin Mountain (Maryland Blue Ridge). The western margin of the Culpepper basin (Maryland and northern Virginia) probably represents Lower Mesozoic reactivation of an Alleghanian ramp originating at the Iapetan breakup line. The Frederick Valley is displaced eastward from Catoctin Mountain by the extent of the Mesozoic extension. The basement north of the failed rift transform would extend many miles seaward prior to the Taconic thrusting and be much more susceptible to the Taconic docking stresses. Gates and Valentino (1991) and Thomas (1993) also deduce transform faulting during opening of the Iapetus on several different grounds. This would explain not only why Taconic fabrics are found in shelf rocks in the northeastern rocks, whereas they occur only in slope and basinal clastic rocks to the southwest, but possibly also why the present contact of these slope and basinal rocks shifts about 15 miles northwest in the vicinity of the Susquehanna River.

A decade ago the Catoctin volcanics were believed to be at least 100-200 m.y. older than the base of the Cambrian. This implied a substantial unconformity even if the Chilhowee group below the (sparingly) fossiliferous Antietam Formation is assigned to the Eocambrian, as I did in the Correlation Chart of Pennsylvania (Berg and others, 1983). The Catoctin volcanics have now been shown to range up to 570 Ma (Badger and Sinha, 1988; Aleinikoff and others, 1991). Rift to drift transition (breakup), thermal rebound with development of a breakup unconformity, and the seaward thickening shelf wedge deposited during thermal relaxation [about 2000 feet of Chilhowee (Jonas and Stose, 1944) and over 3000 feet of certifiably Lower Cambrian predominately carbonate rocks (Taylor and Durika, 1990) in the West York block] must be compressed into 30 m.y. This interval is closely commensurate with timing of such events at the Jurassic margin of the Atlantic (Vogt and Tucholke, 1989) and the whole business begins to make sense.

The extent of unconformity at the Pigeon Hills is problematical owing to lack of geologic data within the basalts, but it is clearly present in the Hellam Hills where there is substantial angular discordance between units mapped in the volcanics (Stose and Jonas, 1939) and the Chilhowee contact. It is more profoundly apparent on the Honey Brook upland where an unknown thickness of volcanics and Laurentian gneiss were stripped before Chilhowee deposition. I believe the Honey Brook basement continues westward along the plunge at least as far as the vicinity of York with presumably comparable erosion.

Thermal rebound after the breakup produces maximum uplift, with consequent lag of subsidence with thermal decay, in the platform proximal to the breakup line. Erosion from this uplifted area will discharge both seaward - "Chickies slate" of Stose and Jonas (1939) exposed only above the Stoner thrust (Urbana Formation of Edwards, 1988) - and landward - Hellam "member" of the Hellam Hills and probably Weaverton Formation of the South Mountain area.

Subsequent deposits represent the interval of rapid subsidence accompanying thermal decay. These grade up to at least Lower Middle Cambrian shelf margin and shelf carbonates that form a westward-thinning coastal wedge extending westward to or somewhat beyond the Cumberland Valley. It should be recognized that the development of the depositional shelf break initiates where the open water high energy wave base impinges on the initial slope to rework the material and form a marginal shoal, which may prograde or recede in response to fairly minor changes in water level, subsidence rate, or sediment supply. It is naturally limited seaward by the break-up line, but need not prograde so far. In an environment hospitable to carbonate deposition, this shoal may colonize when the pulse of sediment readily available from the regolith of the antecedent platform has been diminished to the point where the water is not excessively turbid and the marginal shoal is sufficiently stabilized. This results in the production of a marginal bank that is not necessarily at the breakup line, as it appears to be at the Pigeon Hills. On the flanks of the Honey Brook upland the marginal carbonate bank is obviously above extended Laurentian basement, and it this condition that probably prevails at the crop of this unit throughout Chester, Lancaster, and eastern York Counties.

Lower Paleozoic Passive Margin Subsidence. The Iapetus was no "pacific" Atlantic quietly separating continental plates from a single medial spreading center, but more a Pacific-type Ocean that contained island arcs, one of which was to dock here to generate the Taconic orogeny and terminate the purely passive character of the Laurentian margin of the Iapetus. The Avalon composite terrane was also already being assembled from microcontinental fragments somewhat farther out. The docking of the Avalon terrane is reflected in the Acadian orogeny of Maritime Appalachians. It seems to have arrived somewhat later, in the Upper Devonian, opposite Pennsylvania. Emplacement was apparently transcurrent, as no tectonic impress of this event is definitely recognized within the state, but the event is well marked by the Upper Devonian and Lower Mississippian flysch and molasse sequence of the foreland (Faill, 1985).

The Iapetus coast of Laurentia, however, was grossly similar to the contemporary Atlantic margin. Allowing for the environmental differences of low latitude, abundant carbonate deposition, and hinterland of apparently much lower relief, which gives the two cycles a different stratigraphic aspect, comparison of Figure 1 with a contemporary reconstruction of Atlantic Coastal Plain and shelf (e.g. several in Sheridan and Grow, 1988) shows a geometrically and genetically similar epi- to pericratonic wedge evolved in comparable intervals. The section is derived from Read (Rankin and others, 1988, p. 51) with modifications especially near the eastern margin. It is not palinspastically restored. While slightly revised for clarity, the relations between Lower to Middle Cambrian facies near the bank margin in the original diagram are substantially the same as advocated here.

Figure 1, based on the known or projected thicknesses assumed to be present in the reference sections, is effective in portraying the overall geometry of the depositional wedge. It is much less suitable for portraying the relationship between the diachronous facies appearing in the neighborhood of the bank margin. Figure 2 is drawn with a vertical time axis, arbitrary compressed or expanded according to detail comprised in each interval, to emphasize the temporal relationships. Marietta is located in the slice between the Mechanicsville and Chickies thrusts west of their divergence. Despite the abrupt bend in the section, the attempt is to portray a schematic section normal to the bank edge which aspires to be more-or-less palinspastic.

Inclusion of rocks above and below those exposed near the section line of Figure 2 are necessarily speculative. Extension of the Henderson Marble (see end of this paper) west to Lancaster County is particularly dubious, but it serves to illustrate the possible relationship of these rocks that we will see in part at Stops 3 and 4.

Meisler and Becher (1971) suggest a pre- or lower-Whiterockian disconformity at the northern edge of Lancaster carbonate belt. This may be dynamically equivalent to the Black Riverian diastem which appears within the Chambersburg Formation near Chambersburg (Craig, 1949) and increases northeastward though the Great Valley to encompass all of the Black Riverian in the Lebanon Valley and also the Chazyan and at least most of Whiterockian in the Lehigh Valley. I interpret this hiatus to result from a flexural bulge complementary to the initial attempted subduction of Laurentia. As this bulge would migrate cratonward as the subduction progressed, the locus of the hiatus would actually transgress time. Under the circumstances I hesitate to speculate what rocks might overlie those shown in Figure 2, but I surmise that much of it would have a basinal character, perhaps like some of the *Nemograptus*

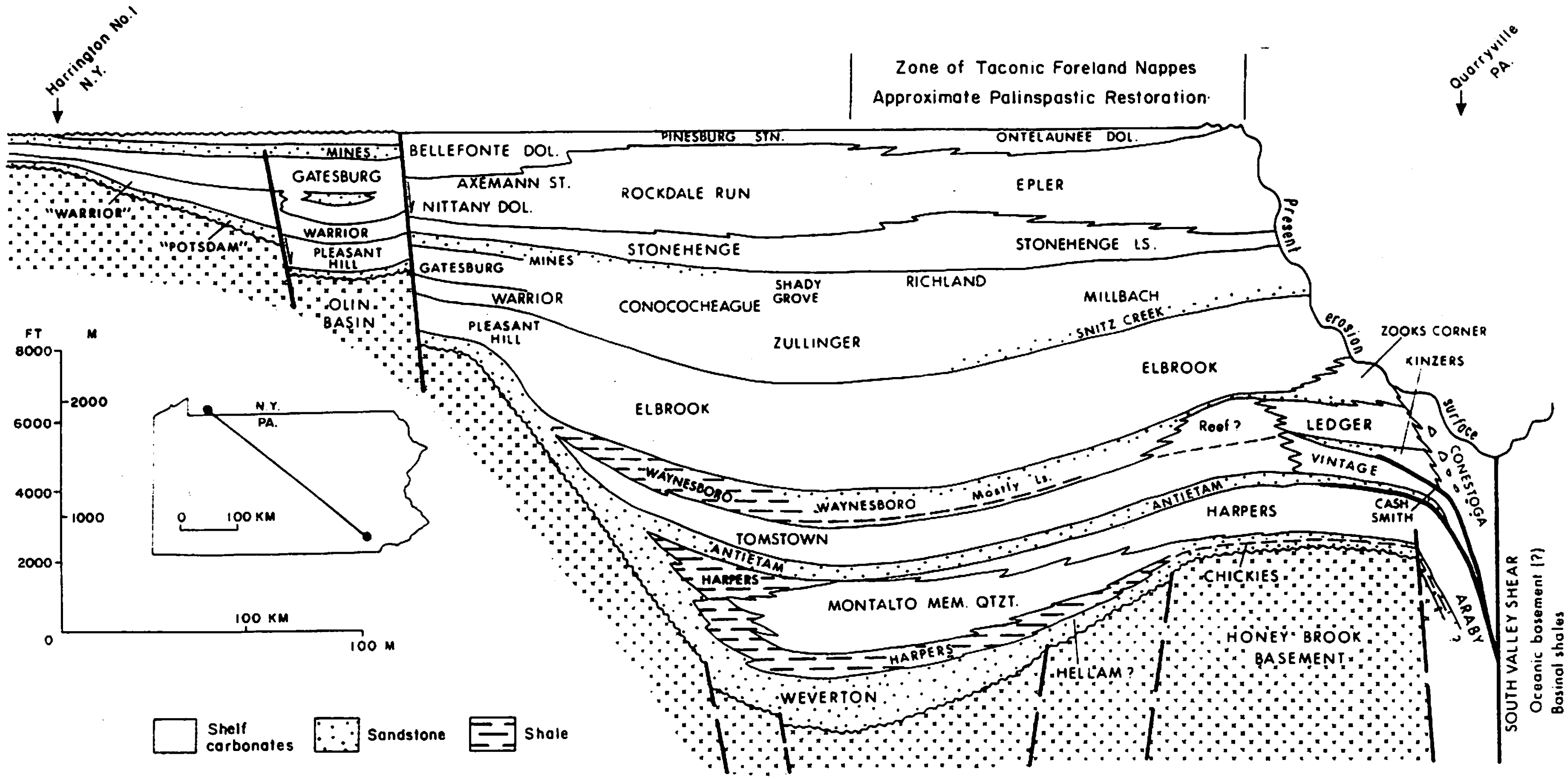
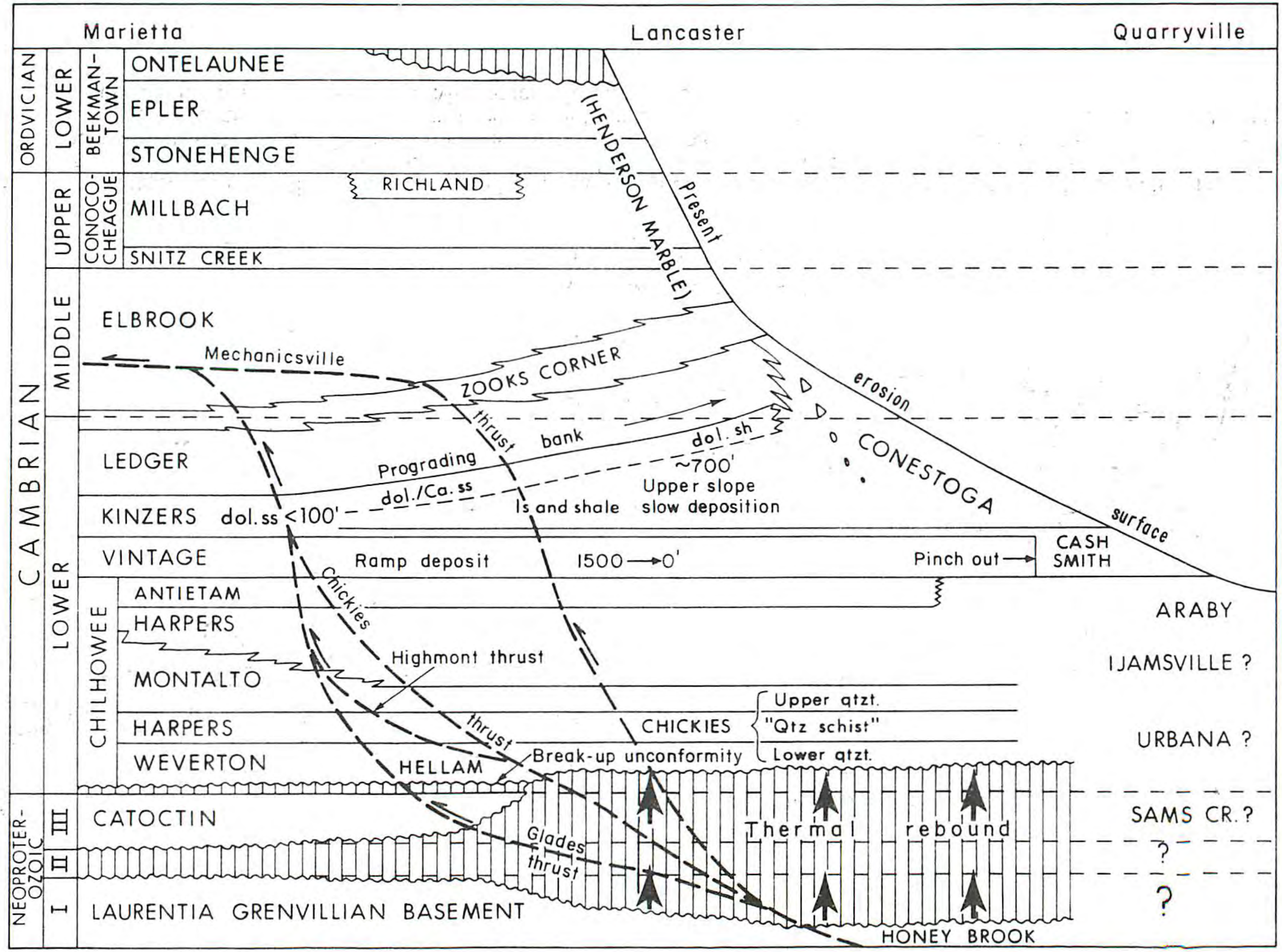


Figure 1. Stratigraphic cross section - Harrington, NY. to central Lancaster County, PA. The western half of this section is slightly modified from Read (p. 51 in Rankin and others, 1988). Alleghanian shortening between the Allegheny Front and the Taconic nappe front is about 70 miles and is not palinspastically restored. Approximate restoration in the Taconic nappe zone includes Alleghanian thrusting within it, but possibly large, although indefinite, displacement on the high-level Alleghanian decollements (Yellow Breeches and Mechanicsville thrusts) is not expressly accomodated.


Marietta
Lancaster
Quarryville
ORDVICIAN
LOWER
BEEKMAN-TOWN
ONTELAUNEE
EPLER
STONEHENGE
(HENDERSON MARBLE)
Present
erosion
surface
CAMBRIAN
UPPER
CONOCO-CHEAGUE
RICHLAND
MILLBACH
SNITZ CREEK
MIDDLE
ELBROOK
Mechanicsville
ZOOKS CORNER
thrust
bank
dol. sh
Prograding
~700'
LEDGER
dol./Ca. ss
Upper slope
slow deposition
KINZERS
dol.ss <100'
ls and shale
CONESTOGA
VINTAGE
Ramp deposit
1500→0'
Pinch out
CASH SMITH
LOWER
CHILHOWEE
ANTIETAM
ARABY
HARPERS
Chickies
Highmont thrust
IJAMSVILLE ?
MONTALTO
Upper qtzt.
"Qtz schist"
Lower qtzt.
HARPERS
CHICKIES
thrust
URBANA ?
WEVERTON
HELLAM
Break-up unconformity
NEOPROTEROZOIC
III
CATOCTIN
SAMS CR. ?
Thermal rebound
Glades thrust
II
?
I
LAURENTIA GRENVILLIAN BASEMENT
?
HONEY BROOK

gracilis zone ribbon limestones found in the Hamburg klippe. In any event the Taconic arc is drawing nigh, as reflected in the occurrence of thin ash beds through much of the Chambersburg Formation.

Taconic Orogeny. The only clear imprint of the Taconic on the (future) upper plate of the Mechanicsville thrust is biotite grade (upper greenschist facies) metamorphism at about 440 Ma or little earlier (I deduce about 458 Ma from graptolite data in the Hamburg klippe) allowing for unloading and cooling to argon retention. The Taconic has profound impact on the Piedmont at large, but only the few features necessary to set the stage for the Alleghanian in the bank margin area and the discomfiture of Stose and Jonas are discussed.

Hoersch and Crawford (1988) recognized Taconic detachment of the Honey Brook-Mine Ridge massif, amplifying ideas they had expressed in 1984. Faill and MacLachlan (1989) recognized that this detachment must be expressed by the Mechanicsville (Oregon) thrust as a significant terrane boundary. In the carbonate valley of low relief the low dip of this thrust is not readily apparent, but is very conspicuous in its Chickies thrust splay, which has a conspicuously curved trace and an outlying klippe near the area of their convergence. Stose and Jonas recognized all of this (except the Mechanicsville thrust) and used it in 1933 portraying the relationships at Chickies Rock in the Middletown quadrangle. The Mechanicsville thrust transects Taconic fabric in its lower plate and carries folds of Alleghanian style and orientation in its upper plate which plicate strata bearing the clear impress of the Taconic mimetic recrystallization. This style of upper plate deformation persists, following the change in the Alleghanian grain, to the West York area where the Pigeon Hills basalt is separated from South Mountain only by the Lower Mesozoic extension (MacLachlan, 1991). The exclusively Alleghanian deformation of the last, above the breakup unconformity, has been asserted by numerous distinguished authors and, to the best of my knowledge, disputed by none. The Mechanicsville thrust has the clear aspect of a high-level Alleghanian decollement.

Alleghanian orogeny. Hoersh and Crawford (1988) suggest that folding of the Honey Brook-Mine Ridge massif and its cover, and retrograde metamorphism of the basement were a result of the Taconic detachment. I must respectfully dissent. Their speculations on the genesis of the pegmatites are interesting. If these are in fact Paleozoic (a radiometrically testable hypothesis?), they might have induced some deuteric alteration of the basement gneiss in the Taconic, but Bloomfield (1989) offers an alternative hypothesis which is more compatible with events in the cover rocks apparently not impacted by such events. Maximum prograde metamorphism was attained beneath the vast hot (Faill and Wiswell, this guidebook) overburden of exotic oceanic Taconides which subsequently shed their detritus to much of the sediment in the Appalachian basin. Retrograde metamorphism (as biotite grade to chlorite grade greenschist facies) appears in the cover rocks, but only where they are impacted by Alleghanian shears (Valentino, 1990; Valentino and Faill, 1990; 1993). The reasons for doubting essentially Taconic folding are enumerated in the preceding paragraph, though my observations do not preclude the possibility of some Taconic warping of the upland region which did not tectonize the cover rocks.

These considerations do not imply that the Taconic detachment must be rejected; rather it is required, and the only dispute is with regard to its impact on the deformation of the

Figure 2. Correlation diagram of rocks on the upper plate of Mechanicsville thrust system: Marietta-City of Lancaster-Quarryville, Lancaster County. It should be noted that the Elbrook, in part, and all younger formations presently underlie the Mechanicsville thrust in a generally inverted nappe sequence below the Alleghanian Mechanicsville thrust. The upper and lower parts of the sections exposed in the Marietta and Lancaster areas were not in depositional proximity. In the eastern Chester Valley the Zooks Corner Formation is depositionally succeded by the Henderson Marble which appears to extend to the Lower Ordovician. This unit sufficently resembles the Elbrook-Conococheague-Beekmantown succession that it seems appropriate, for want of something better, to extend the existing names to represent the formerly present platform carbonates in the Lancaster area. The heavy dashed lines represent the stratigraphic locus of future Alleghanian thrusts.

exposed rocks. It is precisely because the Honey Brook plate was the highest deck of the Laurentian-cored Taconic foreland nappe pile that the upper plate of the (future) Mechanicsville thrust escaped the pervasive Taconic tectonitization and overturning apparent in the lower nappes. I deduce that the protoMechanicsville thrust ramped up from the basal Taconic detachment, probably utilizing a relict Iapetan extensional ramp, possibly to the Elbrook Formation. Additional inboard ramps, that developed out of the same basal detachment, generated the lower nappes. During the Alleghanian docking of Gondwana, the Taconic thrust system was reactivated and advanced as a decollement sheet at a high structural level with respect to the sole thrust of the Valley and Ridge fold system. The structurally similar Yellow Breeches thrust (MacLachlan and Root, 1966; MacLachlan, 1967) is undoubtedly very late Alleghenian and presumably arose from a more inboard Taconic ramp and passed into high-level decollement in the Upper Ordovician flysch. These late Alleghanian low angle thrusts, apparently involving no rocks accreted to the original Laurentian margin except the allochthonous pelites, wyldflysch, and miscellaneous rocks of the Hamburg klippen, are essentially in the style of Appalachian foreland deformation and are part of the out-ofsequence thrust family of Faill (1991). They are believed to have large displacement and have played an important part in thickening the proximal foreland wedge. The Mechanicsville thrust extended at least as far as Reading because there is no other plausible source of klippen of bank margin rocks that occur in that area (MacLachlan, 1983). The late high-level decollement system in general has been proposed to have formed a significant portion of the very thick cover above the Llewellen Formation in the anthracite region at the end of the Paleozoic (MacLachlan, 1985; in press).

A different aspect of Alleghanian deformation, transpressional shearing, is addressed by Valentino (this guidebook). South of the Brandywine fault, east of the Susquehanna River such deformation has impacted Laurentian basement and its cover to varying degree (Bloomfield, 1989; Hill, 1989; Valentino, 1990; Valentino and MacLachlan, 1990). The extent to which this process has modified the older depositional patterns and structural fabric is not entirely clear. As presently mapped it may superficially appear that the Chester-White Marsh Valley approximately parallels the margin of the Lower Paleozoic bank, thus presumably also the Iapetan rift trend. In fact the valley trend is defined by the late shearing which is somewhat inclined to the depositional trend, as can be demonstrated in the stratigraphy of some of the units. This would be more apparent on the map if the shelf facies rocks of the eastern part of the Valley presently mapped as Conestoga Formation were reassigned. The Henderson Marble is proposed below to receive these rocks. My present opinion is there are no slope facies rocks in the White Marsh Valley, but to say they are poorly known is understatement - they haven't really been mapped at all beyond the hasty reconnaissance level (not really adequate for contemporary standards of 1:250,000-scale work) except as the "Shenandoah Limestone". West of the Susquehanna the northwestern limit of such shearing is yet to be clearly defined, and the indicated presence of oceanic basement in southern York County along the strike of Laurentian basement to the east might materially impact how it is expressed.

The Huntingdon Valley-Liberty shear zone (Valentino and others, 1994) trends more northeasterly than shearing in the Chester-White Marsh Valley and northward, and it may have much greater displacement. The Huntingdon Valley segment, of the Valentino and others major shear, separates clearly Laurentian rock and its Lower Paleozoic cover of the White Marsh Valley from highly metamorphic rocks of oceanic provenance to the south - the **REAL** Wissahickon schist, not to be lightly confused with other rocks often borrowing the name. To the extent it impacts the rocks here addressed it must be considered. I leave the rest to other discussants. The Huntington Valley shear zone was called "Martic" by Hill (1989). It indeed adjoins part of the "Martic line" as defined by Jonas (1929). Hill now realizes that designation generated more heat than light, and her principle concerns are simply not what Jonas was addressing. Nearly horizontal translation, deduced by Valentino and others (1994) to be in the 100+ km range, may be to a degree concentrated at the Huntington Valley fault (Wissahickon border) of long established usage, but a wide band of Grenvillian gneiss of the Trenton prong is strongly sheared with a distinct northern border. Lindline and Hill (1992) ran a show for skeptics along Pennypack Creek near the Bucks County line. Without their guidance I might have missed much, but I was convinced. I was gratified to note that the northern limit was directly on strike with major discontinuity in the carbonates of the Whitemarsh Valley. The latter projects to the foot of the South Valley Hills near King of

Prussia, and it extends thence westward along the south side of the Chester Valley to at least the vicinity of Quarryville in Lancaster County. Exposure conditions in the White Marsh Valley preclude directly establishing mechanics associated with the lineament there, but late foliation and lineation in the Octoraro phyllite adjacent to the Chester Valley appear to reflect a transpressional regime. That the south side of the Chester Valley is a substantial shear in its own right is most evident in the fact that it throws off several splays which in turn can be shown to impose a strong subhorizontal transport fabric on carbonates exposed in their vicinity in the Chester Valley. For want of a better name that might avoid confusion I call this the South Valley shear. This is the most embattled part of the classic "Martic line". I have seen the Martic thrust in several places and it is real, but at the south side of the Chester Valley it is either cut off or transpressionally modified to a steep fault.

There are at least two subordinate shears in the Chester Valley which diverge from the South Valley shear at an acute angle and cut across to the Honey Brook upland. The Exton fault emerges near Malvern and cuts the Chilhowee north of Thurmont in the eastern Coatesville quadrangle. Its probable continuation into the gneiss is unknown. The Coatesville fault emerges near PA Route 100 in the western Malvern quadrangle and either merges into the Conestoga/Harpers contact or cuts into the Chilhowee northwest of Coatesville. Stops 3 and 4 are located close to these two faults respectively, and structures related to transpression as well as stratigraphy unfamiliar to most will be featured. The "Gap overthrust" (Bascom and Stose, 1938) somewhat to the north of these localities may have its origin in the same late Alleghanian shearing, but it clearly has been extensionally reactivated, if this is the case. I consider it sufficient to regard it simply as a Lower Mesozoic fault.

Atlantic cycle rifting. There are a number of faults in the Paleozoic rocks in the general area of the bank margin which have known or suspected Lower Mesozoic displacement. Some actually cut the margin of the Newark-Gettysburg basin. A notable example of the suspect category is the unnamed high angle fault that complicates the geometry of the junction of the Mechanicsville and Chickies thrusts in the Columbia East quadrangle. This fault cuts across the Chickies ridge followed by the grade of the former Wilmington and Columbia division of the Reading Railroad and serves to divide Chickies Ridge proper from Chestnut Hill to the east (herein Chestnut Hill fault for reference below), and it extends well into York County trending parallel to the north margin of the Mesozoic basin. Such faults do not ordinarily have significant impact on interpreting the evolution of the Lower Paleozoic bank edge, and could be ignored in this context.

Correct interpretation of the "Gap overthrust" of Bascom and Stose (1938), however, bears on interpretation of facies in the Chickies Formation where two belts appear on the south side of the Honey Brook upland in the Downington quadrangle. This fault is easily traced across the Honey Brook upland and is marked by sporadic to often abundant large blocks of white bull quartz. It crosses a terrain of considerable relief without apparent deflection so that it is clearly very steep and the dip direction is ambiguous. It passes eastward into a series of (steep?) fault bounded slices in the Chester Valley in the Malvern quadrangle where field checks show Bascomb and Stose (1938) to be fairly reliable. The Valley Forge quadrangle has not been officially remapped since 1909 (Bascom and others), but a reconnaissance sheet for the Norristown 15-minute quadrangle was obviously produced for the 1931 Geologic Map of Pennsylvania, as the "Shenandoah Limestone" was (crudely) subdivided into Ledger, Elbrook (at least mostly Zooks Corner), and Conestoga Formations. To the extent that the ancillary data is reliable, the faults of the Malvern quadrangle appear to extend to the confirmed fault near Bridgeton which offsets the contact with the Upper Triassic Stockton Formation. To the west at Gap, Chickies is allegedly thrust over the Vintage dolomite; however, despite the deep Amtrack cut, contact relations are obscure along the railroad.

While none of the above details are particularly consistent with thrust origin of the Gap fault, it is easy to suppose that it is related to the formation of the Mine Ridge anticlinorium in the Gap area. A transpressional shear comparable to those in the Chester Valley might serve the case without violence to the observed geometry, but no positive evidence for such a feature has been observed. The quartz previously mentioned obviously formed in an extensional regime. Occasional blocks of this quartz contain large (up to 2 feet) inclusions of well foliated chlorite which is sometimes contorted. I discussed these occurrences with R. C. Smith (Pennsylvania Geological Survey), presenting the hypothesis that these might represent a phyllonite from a fault precursor to the Mesozoic extension. He said that the mineralogy was much more compatible with the obvious hydrothermal activity of the extensional phase "where everything goes to chlorite" than with compressional shearing of the dry Grenvillian gneiss. A Mesozoic fault with a maximum displacement in the order of 1000+ feet down to the north, presumably antithetic to the low angle extension fault at the north margin of the Newark basin, is sufficient to explain all observed features of the "Gap overthrust". It is possible, however, that the fault at Gap is not the same as the "Gap overthrust" of the Downington quadrangle. The map evidence (Bascom and Stose, 1938) for projecting the trace across the Wagontown quadrangle is far from compelling, but I have not attempted field validation.

STRATIGRAPHY AND EVOLUTION OF THE SHELF MARGIN

Chilhowee Group, Lower Cambrian. Named from Chilhowee Mountain in the Tennessee Blue Ridge (Safford, 1856), the Chilhowee Group comprises the Lower Cambrian basal clastic sequence deposited on the rapidly subsiding Laurentian platform adjacent to the Iapetan rift margin. The defining concepts here are more modern than the work of many of the stratigraphers who have used this convenient collective, but they used it consistently in the central and southern Appalachians, with regionally variable formational subdivision.

The formations introduced by Stose (1906) for the South Mountain area of Pennsylvania (Loudoun, Weaverton, Harpers, and Antietam) were standard usage in Maryland (Clark, 1897) from types near the Potomac River in the Blue Ridge-South Mountain of northern Virginia and Maryland. Clark excluded the (slope to basinal) clastic rocks on the east side of the Frederick Valley, referring them to Cambro-Siluric schists. Possibly encouraged by Lesley's (1892) reference to the Chickies of southern York County, Stose applied his southeastern Pennsylvania standard section (Chickies, Harpers, and Antietam) (Stose and Jonas, 1922) to rocks above the Stoner overthrust (Stose and Jonas, 1939) in Adams County (Stose and Bascom, 1929; Stose, 1932).

Meanwhile, having established a reputation, Jonas (1924, 1928; Knopf and Jonas, 1923, 1929b), abandoned the penury of the Pennsylvania Geological Survey for Maryland, and carried this interpretation of the southeastern Pennsylvania stratigraphy east of the Frederick Valley where it was subsequently used in a couple of other Maryland publications (Jonas and Stose, 1938; Stose and Stose, 1946). For reasons I deem better than mere provincial chauvinism, even though the initial decision was in part influenced by an errant fragment of the Braillier(?) Formation, the Maryland Geologic Survey subsequently rejected this correlation. The "Antietam" formation was renamed Araby Formation (Reinhardt, 1974). This and other slope to basinal equivalents of the Chilhowee Group were remapped and correlated to their platformal parents (Edwards, 1986). The Araby Formation of the northernmost Frederick Valley, thrust over the Frederick (slope) Limestone (Edwards, 1988) lies only about 12 miles south along strike from the southernmost Hanover Valley where the "Antietam" and "Harpers" are thrust over the Conestoga (slope) Limestone. The concealed interval lies merely beneath Triassic cover at the eastern, insignificantly faulted margin of the Gettysburg basin. I see no reason to doubt this to be a structurally continuous relationship.

I concur with Reinhardt that "the Frederick Limestone is not the Conestoga Limestone" (see remarks anent Conestoga Limestone and Henderson Marble below), although the lower half could be so called. While it will take some remapping to establish appropriate contacts for the units, I believe the Maryland slope to basinal clastic units are much more appropriate to the upper plate of the Stoner thrust. The "Chickies slate" with its occasional good quartzite beds is certainly much more like the Urbana Formation than anything in the Hellam Hills, and any one who wanders the hills southeast of Hanover with the Araby type description in hand might feel at home. We should avoid provincial chauvinism and accept the Maryland units, as we did on South Mountain. With these "grungey" rocks safely consigned to Poseidon, we may return to the real Chilhowee to ascertain if we can see what is happening at the Lower Cambrian shelf margin.

Chickies Formation (Hellam member excluded). I consider the Hellam "member" of the Hellam Hills with its conspicuous medial conglomerates to be Weaverton Formation of South Mountain affinity deposited well inboard of the Chickies of the Honey Brook upland. The Chickies formation does have a basal conglomerate (or nearly basal - a few feet to a few tens of feet of chloritic phyllite are evident below it in some places and could be quite common owing to the almost complete lack of exposure in the neighborhood of the basement contact). The conglomerate is fairly persistent to the north and thinner and more sporadic to the south, and is consistently overmapped everywhere by Jonas, Stose and company, presumably in keeping with the concept of a thick Hellam member from the Hellam Hills. Beds above the conglomerate are not consistently differentiable from other Chickies exposures in the same section, and the use of the term Hellam is clearly inappropriate. Whatever the origin of the rocks in one thrust slice within the Hellam Hills, the unit designation Hellam belongs uniquely to those rocks if it is stratigraphically useful at all.

Lesley (1892) devoted considerable space to the proposition that scolithus was characteristic of the Chickies Formation. Subsequent descriptions emphasized this characteristic while tending to supress the fact that in most areas less pure, usually very fine grained, quartzites of various shades gray, greenish gray, or moderate brownish gray, along with variable amounts of originally more pelitic rocks ("quartz schist" of Stose and Jonas) dominate the Chickies. The fact that such rocks do exist in most places north of the Gap fault where significant quantities of Chickies outcrop or rubble are encountered, however, is non trivial. Goodwin and Anderson (1974) characterize the Chickies (all studied localities north of the Gap fault through a longitudinal range essentially identical to that which I have examined much less intensively but more extensively) as a mosaic of migrating subtidal channels. That's good enough for me. That the scolithic quartzites are the product of shallow agitated water is quite apparent without such reminders as Gohn and Chacko (1974) though a recent study of a possible Holocene analog (Skoog and others, 1991) provides some insight on features of an infratidal to shallow subtidal depositional environment which are not apparent in the bioturbated rock. These rocks have a maximum strike-normal exposure of nearly 15 miles across the Honey Brook upland and everywhere underlie a stratigraphic section including units I consider diagnostic of the bank margin.

South of the Gap fault the situation is somewhat different. At the eastern end of the southern Chickies belt of the Downington and western Malvern quadrangles, where these rocks are in close proximity to those previously described, they are much saprolitized and support only slight relief. Little can be deduced of their initial lithology. Westward at the first adequate exposures some distinctions become apparent. While mostly still fairly decent quartzite, much of it is thin to medium bedded and Skolithos is notably absent. The whole formation is distinctly thinner than to the north, the basal conglomerate is restricted to a thin zone of fine pebbly granulite which may be absent in places, and I encountered the first occurrences I had seen of stretched tourmaline crystals which had been said to be characteristic of parts of the Chickies formation. A traverse across the Chilhowee in the vicinity of the West Branch of the Brandywine north of Coatesville, approximately at the western limit of bank margin facies exposed in the Chester Valley, reveals the following particulars. Except for a lens (channel?) of subangular blue quartz pebbles with a chloritic matrix enclosed in about 30 feet of dark, dense (high iron?) phyllitic chlorite, the basal conglomerate is not present. The basement contact is not exposed, but occasional gneissic fragments some distance to the east provide some constraint on the thickness of this unit. The overlying Chickies is somewhat reminiscent of the previously mentioned section up Chickies Ridge along PA Route 441. A lower member of fairly pure, but not scolithic, fine grained quartzite in beds to about 2 feet is succeeded by a thick interval of rocks possibly even more pelitic than those of the reference section. This is capped by an upper quartzite member, perhaps thinner bedded on the average, but generally similar to the lower member.

"It is all down hill from here" is just a sober statement of the depositional environment. At the Amtrack cut near Atglen the Chickies and Harpers Formations are converging in character. It is possible to see why Bascom and Stose (1932) placed the contact where they did, but any who claim it is obvious have an eye more schooled in the subtleties of such rocks than mine. At the westernmost series of fairly good exposures on the south flank of Mine Ridge along Nickle Mine Creek (central Gap quadrangle) the contact was more easily distinguished because the contrast was enhanced by weathering.

Harpers, Antietam, and Ibid. undifferentiated Formations. These two formations, though obviously related, should be differentiated where practicable owing to significant differences in geotechnical characteristics usually apparent between them. The third appellation is excessively used on the current State Geologic Map (Berg and others, 1980), though it is perhaps appropriate to such areas as Chickies Ridge where the distinction is discernible where there is optimal exposure but which becomes transitional to "zilch" (nonexistent) elsewhere. The change from the arenitic Chickies to the silty/sandy pelites of the Harpers occurs simultaneously, to the limit of resolution, between correlative strata throughout the Blue Ridge province, and it represents an abrupt increase in water depth. A major eustatic sea-level rise is possible, but in this apparently nonglacial epoch a critical stage in the recovery from the thermal rebound resulting from the Iapetan breakup is more probable. There is no evident difference in the kind, though the quantity may diminish, of mixed, predominantly fine grained, siliciclastic debris supplied to the evolving Lower Cambrian shelf during this interval. Following the initial rapid subsidence the shelf recovered sufficiently to permit winnowing of an upper sandy zone (e.g. Kauffman and Frey, 1979) which is continuous except toward the seaward margin.

There is no Antietam Formation on the south side of the Honey Brook upland or Mine Ridge. Bascom and Stose (1932) recognized this fact. They "waffled" a bit saying in effect that the Antietam ought to be there but they couldn't really quite see it, but they correctly mapped only Harpers. I considered substituting Araby Formation at this point, but, with still recognizable Chickies below it, just Harpers seems more appropriate. In 1938 they avoided further embarrassment by doing the same without comment. The appearances are that it was Stose who suggested to Jonas that the Antietam should be found in the Quarryville area (Knopf and Jonas, 1929). Being a good sport she did to the amazement or amusement, according to temperament, of subsequent observers. Barring this youthful aberration, and a questionable extension of this terminology to the Stoner overthrust plate, these units have been mapped by Jonas, Stose and Co. in most places where they occur in the Pennsylvania Piedmont with precision quite sufficient for 1:250,000-scale work. By the time these units are getting ready to slop over into the briny deep they are rather different than their remote types, but the association is more informative than misleading, and I perceive no need for stratigraphic revision.

Bank margin evolution during Chilhowee deposition. Bascom and Stose (1938) were clearly aware of the differences in the Chilhowee units on either side of the Gap fault. That they did not use this information as confirmatory evidence for the hypothecated Gap overthrust simply demonstrates they did not share my sensitivity that abrupt lateral variation in established units may be tectonically induced. If there is no substantial component of horizontal displacement here, then we may locate the top of the clinothem of a depositionally constructed Early Cambrian clastic prism fairly precisely between the northern and southern Chilhowee belts of the Downington quadrangle. This feature appears to be somewhat oblique to the present strike, as Skolithos in the Chickies reappears in the North Valley Hills in the central Malvern quadrangle to the east, while to west essentially all of the Chilhowee in Mine Ridge appears to belong to the clinothem facies. It is much more nearly parallel to the Gap fault than the lithic strike, though that coincidence is clearly significant only if there is a precursor fault which localized the depositional slope-break. Reactivation of an inboard rift related fault during post-rebound subsidence is a possible explanation. The depositional slope break is clearly located on rocks of the Laurentian platform, and it is not far removed from the bank to slope transition in overlying carbonates which also seem to be somewhat inboard of the actual breakup. The South Valley shear may be more directly influenced by proximity to that feature.

When the clinothem is followed down slope to the west, one arrives at the Quarryville area (Knopf and Jonas, 1929) where the down-slope residual of the Chilhowee Group was a significant factor in a bitter dispute (Mackin, 1935; Cloos and Heitanen, 1941). In retrospect it is apparent that Knopf and Jonas, in defining the Martic thrust and having some grasp of its implications, had more megatectonic insight than their critics, but they did make several mistakes in this area that left them vulnerable. Cloos and Heitanen were perfectly aware of Bascom and Stose's (1932, 1938) reluctant conclusion about the character of the upper Chilhowee; they cited both papers when convenient to their purposes. Their exaggeration of the previously noted foible--they reduced the Harpers to a member of the "Antietam", hardly a notable contribution to understanding of the stratigraphy of this area--was less than charitable but perhaps defensible. As noted in preliminary remarks on the Chilhowee, Knopf and Jonas had transported this nomenclature to Cloos's back yard for rather similar rocks.

Two other errors are apparent from my perspective on the rocks I looked at in this area. Their failure to recognize that the linear South Valley shear was not simply a manifestation of the Martic thrust led them to map Conestoga Formation in topographic embayments of the South Valley Hills which demonstrably contain Octoraro phyllite exposures in some cases. The second is related to this understanding, but more subtle and condonable. Southwest of Quarryville they mapped a thin tongue of "Wissahickon" projecting northward which was not inconsistent with their model.

Cloos and Heitanen landed on this tongue and traced it northward. They convincingly demonstrated that the "Wissahickon" became indistinguishable from the "Antietam" - hence the arbitrary cut-off noted previously. The perception that the shelf clastics must have some basinal pelitic derivative carried this well documented relationship onto two subsequent State Geologic Maps and the 1983 State Correlation Chart. What is less apparent to me on the basis of limited mesoscopic comparison is that this "Wissahickon" is identical to the Octoraro phyllite of the South Valley Hills. Ironically Valentino and others (1994) suggest that the latter is Precambrian as Knopf and Jonas insisted. Approaching these rocks from the platform yields a different perspective. The Chilhowee stratigraphy, which is obviously deteriorating down slope on the south side of Mine Ridge, simply becomes dysfunctional when extended to the even deeper outliers of the Quarryville area. I propose, unless I get too much static from Maryland, that these rocks should be assigned to the Araby Formation (Reinhardt, 1974) which, though more metamorphosed, they rather resemble. I believe that this terminology should be applied to the various outliers of "Antietam" near the Martic line which extend westward to Pequea Creek, but I bequeath validation to others.

Vintage Formation, Lower Cambrian. The Vintage is a carbonate ramp composed largely of detrital carbonate. East of the Susquehanna River it is almost entirely dolomite. Occasional limestone beds argue against any pervasive late dolomitization. The magnetically determined paleolatitude is consistent with the upslope development of a coastal sabkha, and it is reasonable that the original detritus may have been dolomitic. Considerably more calcareous rock appears in York County (Gohn, 1976), but this is associated with an earlier evolution of a carbonate bank margin in that area.

In the vicinity of the Susquehanna River, which was incised during the Upper Tertiary, the Vintage Fm. is fairly well exposed across several strike belts. It appears to be nearly 1500 feet thick in the belt north of the Chickies thrust, where it is exposed in riverbed and bank ledges and locally along a few tributaries. The basal contact is about a 10-foot-thick zone of thinly interbedded quartzite and dolomite; otherwise thick to very thick beds predominate. Medium dark gray dolomite with lighter gray patches, noted in almost all descriptions of this formation, is not conspicuous in these ledges but occurs with sufficient frequency in float to be reassuring; a differential weathering process may be indicated. Gohn (1976) attributes this characteristic to infaunal burrowing, which probably implies a maximum depositional depth of several hundred feet for these deposits. Other internal features of Vintage beds are not apparent to my eye at outcrop, but Gohn has demonstrated that Bouma cycle elements (usually considered diagnostic of turbidites) are often observed in peels from etched sections.

South of the Chickies thrust I calculate 660 feet of Vintage from the exposed basal contact to the first occurrence of dark basal Kinzers shale north of Columbia, and it is apparently about the same at Wrightsville. It is not well exposed in this area but appears to be lithically similar to the northern belt. This thickness is consistent with the thickness gradient of Gohn's estimate of about 1000 feet at York and 800$\pm$ feet near Kreutz Creek (village) on the same strike belt, and it is the upper limit of thicknesses reported elsewhere in Lancaster County. Lacking a reliable way to estimate the down dip thickness gradient, the amount of shortening on the Chickies thrust implied between this belt and the northern one is indefinite but must be substantial. The basal contact is exposed in a modest cut on the PA Route 441-US Route 30 interchange ramp, where a few feet of real quartzite (all that is apparent in this otherwise very fine grained and dirty Antietam section) grade through about 18 inches of calcareous sand with some doloclasts to fairly typical, but somewhat calcareous, Vintage much as near the New Holland quadrangle type section.

On the south side of the valley at Columbia the Vintage has limited exposure, but this belt is almost fully exposed in the County Line Quarries pit a half mile south of the center of Wrightsville. Gohn (1976) measured 256 feet in the quarry and estimated 279 feet for the complete section. South of the Antietam-cored anticline of the North Manor Hills the Vintage is best exposed in a series of small quarries along PA Route 441. The basal contact here shows a few feet of thinly interbedded Antietam and Vintage. The overlying unit is the basal Kinzers shale but the neighborhood of the contact is concealed. A reconnaissance "guestimate" of 200-250 feet of Vintage seems consistent with the previous measurement. The Vintage does not come to the river again in Lancaster County, but is present in the northeastern corner of the Red Lion quadrangle in York County. Here 100+ feet of Vintage lies below about 275 feet of Kinzers shale under the Conestoga Limestone. At the same latitude, starting about a half mile east of the river, a belt of Vintage of comparable thickness is beneath a possibly thinner black shale zone which has been mapped in the Conestoga. This situation is representative of a recurrent inconsistency in the mapping of this interval by Jonas, Stose and Co. which is addressed in the next section. The distal toe of the Vintage ramp appears in thrust slices near the Martic line in Lancaster County. The thickness here is estimated as about 20 feet maximum to complete pinch-out (Cloos and Heitanen, 1941). These beds are otherwise significant as localizing the ore at the Pequea mine.

In all occurrences noted in the preceding paragraph the Vintage has a markedly different aspect from the two northern belts. While there is some variation, the most characteristic exposure is dominated by fairly thin, typically about 3-10 cm, dolomite beds, often with nodular bedding surfaces, which may or not have conspicuous dark phyllitic partings. Some disseminated pyrite is not uncommon. While these beds usually appear internally massive at outcrop, Gohn (1976) discerned complete (a through e) Bouma sequences in such beds. Even without such laboratory confirmation the truly deep water character of these rocks is apparent. The abrupt change in thickness and character occurs within a mile across an interval in which there is no evidence for excessive tectonic shortening. What is clearly represented here is a slope break. Such a feature is not implicit in the definition of a carbonate ramp, but, as noted by Wilson (1975), it is more common than not in deposits so identified both in modern and ancient examples. This break occurs slightly seaward of the present contact between the Conestoga Limestone and rocks of the bank margin, but it is close enough in stratigraphic and geographic position and time that it probably profoundly influenced the later development.

In the moderate relief of central Lancaster County where colluviation is scant, the Vintage emerges in various localities, but the good exposures are artificial. To the east around the Honey Brook upland it joins it cousins, the Tomstown and Leithsville Formations, as a unit whose presence is much more commonly inferred than observed. In addition to the obvious problem of extensive colluviation these units are the first carbonates to be encountered by aggressive meteoric waters draining from adjacent silicic ridges, and they may be corroded to substantial depth. The few direct confirmations I have made of this unit from New Holland to the Malvern quadrangle, mostly in float, all appear to belong to the thick-bedded, shallower facies.

"Cash Smith Formation", Lower Cambrian. This name (Edwards, 1988) was introduced humorously into the south end of Figure 2. Judging by what I saw when taking a quick peek at the disputed "Antietam" and "Wissahickon" (Araby) west of Quarryville, the exposure was so poor it was necessary to walk some distance to confirm that the adjacent rock was Conestoga Limestone. There is certainly room for an interval of dark shale which is not uncommon where the Conestoga is (usually ill-) exposed above the Vintage Dolomite where the latter is found seaward of the bank margin. The Cash Smith Formation was described as dark shale unit above the Araby Formation in the Northern Frederick Valley, Maryland. It contains a trilobite fauna specifically identified with that of the lower Kinzers of Lancaster.

The serious point of this introduction is that we need a proper name for the starved basin shale facies of the lower Kinzers. The Cash Smith member of the Kinzers Formation would serve nicely and facilitate the understanding of interstate correlation. Stose and Jonas recognized the existence of this basal Conestoga shale in a number of reports. The most notable is in the Hanover Valley (Stose and Bascom, 1929) where they mapped a basal shale member of the Conestoga. This was subsequently removed to the Kinzers Formation (Stose and Stose, 1944) owing to discovery of Lower Cambrian fossils in it. This was a big deal when it was presumed that this meant moving these rocks to the other side of an unconformity believed to represent a hiatus of about 60 m.y. When it is recognized, the two units are in depositional continuity, as originally supposed, this is merely an appropriate adjustment for different basinal facies. It provides, moreover, positive faunal evidence for Lower Cambrian age of the Conestoga Limestone which is otherwise lacking. The Emigsville Member of Gohn (1976) in a sense has priority, but it is characterized as having the lower third (Taylor and Durika, 1990) composed of argillaceous dolomite transitional to the Vintage Dolomite, which could be assigned to the latter. The type area of the Cash Smith appears to be about as far inboard on the Laurentian Platform margin as the starved basin shale is recognized. The occurrence of transitional beds here is not astonishing, but, to the best of my knowledge, they do not occur much down slope from the bank margin, and certainly not at the Cash Smith type section which is beyond the down slope limit of the Vintage.

Gohn's Kreutz Creek member of the Conestoga Formation, which is described quite similarly to the rocks the Stoses (1944) flipped from the Conestoga to the Kinzers on faunal grounds in the Hanover area, are part of the same depositional package. I am not sufficiently familiar with the York County rocks to judge the merits of Gohn's (1976) lithic distinctions between his Emigsville (Kinzers) and Kreutz Creek (Conestoga) members. On the whole his observations in such matters seem to be reliable, but I am more prepared to believe that these are a function of several (present) miles down slope (plus indefinite tectonic shortening which includes the nontrivial Grantstown thrust) than they are distinct units of different age with the (undated) Kreutz Creek floating above an unconformity (? - query from Gohn, 1978) which has no reasonable place on the basinal side of an evolving passive margin. The capsule description he gives of the Emigsville member when attempting to establish this distinction in any case bears small resemblance to the equivalent rocks of the Lancaster area where the lower Kinzers is characterized by a dark brownish-gray, trilobite-rich, usually noncalcareous argillite of archetypal starved basin aspect. My basic premise is that the starved basin deposits, regardless of where found, are a discrete phase in the evolution of the Laurentian platform margin. They represent the early stage of development of a true constructional carbonate bank [Tomstown Dolomite and Thomasville member (Gohn, 1976) or (preferably as more typical) York member (Ganis and Hopkins, 1990) of the Kinzers Formation] that essentially cuts off all terrigenous platform detritus and had prograded sufficiently to shed much carbonate detritus to the platform margin area only in the York-Hanover district. This genetically similar and essentially isochronous package deserves a distinctive name that is not well served by two different members in two formations. Welcome aboard Cash Smith!

Kinzers Formation, Lower and Middle Cambrian. I take the Kinzers Formation to be everything so mapped by Jonas, Stose and Co. plus two types of rocks they mapped inconsistently where not so named. The first is obviously the "Cash Smith member" of starved-basin shale sometimes mapped as Kinzers, occasionally specified as lower Conestoga shale, and often ignored or vaguely alluded to in passing by Jonas, Stose and Co. As this is a rather extraordinary lithology to find at the base of a sequence presumed to overlie a major unconformity, perhaps they didn't want to think about it too much. I cannot certify that it is present everywhere it might be encountered, but it may be found by careful observation in many places, and shows up even where not sought, as in the insoluble residue data from Wise (1953) in his structural thickness determination of the Conestoga.

Up-slope, in the vicinity of the bank edge, the Kinzers comprises everything between the Vintage and Ledger Formations and is commonly divided into three members, although the upper two are not regionally uniform in age or lithology. The easternmost Kinzers in the Lancaster Valley, adjacent to the Honey Brook upland (Honey Brook quadrangle), is not clearly distinguished from "typical" Kinzers of the Lancaster Valley by Bascom and Stose (1938), but should be. At the best exposure, with a generous quantity of Vintage and Ledger float in the vicinity for control, the unit consists entirely of about 100 feet of dolomitic and calcareous sandstone or sandy carbonate, a lithology not unknown farther west as an upper element, though the silicarenite component there is usually finer. It was correctly mapped through this area as a low sinuous ridge with a conspicuously sandy soil. It is evident that Bascom and Stose (1932) considered the area of sand pits northeast of Coatesville to be the same horizon. A similar unit, but even more quartzose, is discernible in the north limb of the Marietta anticline in the Columbia West and York Haven quads. Stose and Jonas (1933) failed to recognize it among the alluvial gravel terraces and state the Vintage and Ledger are in contact in that area. [They also had not yet identified the upper sandy Kinzers (Stose and Jonas, 1939) in the eastern York Valley of these quadrangles, which might have alerted them.] One local occurrence (patch reef?) of 30 feet of white limestone at the base of this unit appears to be the York member (Ganis and Hopkins, 1990). The overlying beds are more like the upper sandy member in the eastern York Valley than the rock described above the thick York member in its type area, and most like the occurrences cited to the east. It is to be noted that these belts have no basal shale and generally little argillaceous matter. These winnowed deposits presumably formed above wave base and may represent a depositional interval quite as long as thicker, more variable Kinzers does elsewhere.

Campbell (1969) has examined the Kinzers Formation of Lancaster County more intensively than any other since Jonas and Stose. It is to their credit that he (Campbell and Kauffman, 1969) chose to quote their final synopsis of this unit (Stose and Stose, 1944) as fairly characterizing these rocks. He recognized that the uppermost rocks in the Longs Park area (southwestern Lancaster city, Stop 6 of the 55th Annual Field Conference of Pennsylvania Geologists, 1990) differed from the upper sandy member of Jonas and Stose, and he established (Campbell, 1971) that they extended into the early Middle Cambrian. Miller (1934) indicates that most of the quarries that were to provide the critical exposures were open when Adams (Stose, 1932) and York (Stose and Jonas, 1939) Counties were mapped. The Stoses (1944) made some corrections to the Kinzers Formation in the Bittinger, Adams County area from quarry development, but, in a sense, they mapped this area too soon. It is almost funny, or tragic, that the Hanover-York area has unique Kinzers unlike the description that satisfied Campbell for Lancaster County.

Ganis and Hopkins (1990) and Taylor and Durika (1990) describe the present exposure of these rocks in the West York area. They report up to 1200 feet of predominantly very pure limestone in their middle York member alone (which is almost twice the maximum the Stoses allocated to the whole formation). I believe the recent authors correctly relate these rocks to the marginal reef development near Austinville, Virginia described by Barnaby and Read (1990). The Kinzers of the eastern York Valley must be the prototype for the Kinzers the Stoses attempted to generalize in 1944, though the medial impure carbonates are poorly exposed. The best exposure of such rocks that I have seen are in the Kinzers type section on the north side of the Quarryville quadrangle (Knopf and Jonas, 1929) described by *Stose and* Jonas (1922) and Jonas and Stose (1926). [This cut is definitely not for student groups without prior approval and supervision of AMTRACK - the trains are fairly frequent, very fast, and almost silent. (Also not for government geologists they tell me, but I worked here and in the adjacent Vintage type for more than a day before the white hats ran me off. I didn't have the gall to ask them to send someone to watch my back while I fiddled around, though they were willing to make such arrangements - ain't contemporary liability law wunnerful?)]

Ledger Formation, Lower and Middle Cambrian. The Ledger is characteristically off-white to rarely medium light grey, sometimes with a yellow cast, very pure, finely to coarsely crystalline dolomite. It is almost always thick bedded, sometimes exceeding 15 feet, with the beds rarely revealing even a hint of primary structure, organic or physical. If this is the first time you have encountered it, after you have seen it at Stops 1 and 3, if you see it in another excavation in this part of the world (other exposure is almost nonexistent) you will know it. A local variant, the Willis Run member of Ganis and Hopkins (1990), darker bioturbated calcareous rock of lagoonal aspect appears in the West York area. The rock dividing the north and south quarries near Devault, Chester County (Malvern quadrangle) appears to be similar. Other major Ledger exposures indicate this lithology is not a persistent marker, but additional occurrences may be encountered as this economically attractive rock is exploited.

The Willis Run member is interesting for three reasons. As an intercalated lime member it proves that the pervasive dolomitization. and recrystallization of the Ledger is unlikely to be the product late dolomitization, a feature commonly ascribed to invasion by connate water tectonically expelled from basinal deposits. The striking difference between this member and the typical Ledger, which appears to result from a relatively small change in local mean sea-level, reflects the sensitivity of the depositional environment to this parameter. It also provides the only faunal confirmation for the traditional Lower Cambrian age of the

Ledger, which was temporarily in some doubt when Campbell (1971) established that it conformably overlay lower Middle Cambrian rocks at Lancaster.

Except where locally overlapped by Conestoga Limestone with a Middle Cambrian fauna (Taylor and Hopkins, 1990), the Ledger is the youngest Paleozoic-age rock exposed in York County. These authors take this exposure to imply that the top of the Ledger in this area probably is, but no younger than, low Middle Cambrian. However, the coarse clasts of conglomerates in this section imply proximity to contemporaneous bank-edge rocks. None of these clasts looks like "typical" Ledger, but this is also the case in other areas where the parent almost certainly is (e.g., Stop 2). The significant contrast is the clasts are not usually dolomitized. It appears that such clasts are in fact the best indication of the Ledger protolith we have, and that the Ledger dolomitization, if not late, was also not early diagenetic. If this is correct, this locality provides no constraint on the age of the youngest Ledger deposited in this area.

The Ledger is about 600 feet thick in the Marietta area. If my guess that the overlying sandy Zooks Corner Formation of this area represents distal upper Waynesboro sand is correct, the top is probably very low Middle Cambrian. It is over 1000 feet thick in the York area and it may get somewhat younger at the top. Thickness of the same magnitude is deduced in the Chester Valley and is apparent in the true Ledger of the White Marsh Valley (the quarry rock at Stop 3. There is something else underneath it that might be more like Leithsville than Vintage). Elsewhere it appears to be between these limits, possibly tending to the lower range near Lancaster where the formation is entirely Middle Cambrian. Meisler and Becher (1971) report the Ledger intertongues with Zooks Corner in the Lancaster area and the same is apparent at Stop 3, which implies possible Middle Cambrian age in this area also. Elsewhere and otherwise the age of the Ledger is poorly constrained. The base is clearly diachronous and the top probably is also.

If we add to the characteristics above a few other observations a plausible depositional model emerges. The seaward limit of the Ledger where exposed is always a slope facies limestone. It forms a fairly broad band landward, at least 15 miles across the Honey Brook upland, and probably much wider, as it appears on both sides of the Chickies thrust. This is considerably wider than a normal reef tract, and persistence of bedding throughout the unit militates against any significant reef development. Local lagoonal inclusions and the several necessarily partly contemporaneous formations all attest to the fact that fairly abundant pelitic detritus was available in the depositional system. Some quartz sand in the proximal Conestoga indicates coarser terrigenous detritus was also passing through. Both of these components are nearly absent in typical Ledger. The slope conglomerate clasts suggest well winnowed lime grainstones. It all adds up to a bank edge rimmed by a somewhat elevated marginal limestone shoal tract, a facies well able to shift in response to fluctuations in water depth and free to prograde to produce the diachronous contacts. A suitable modern analog is found in the Campeche bank of the Gulf of Mexico, and as in that area, the shoals may often have been emergent.

Conestoga Limestone (Restricted), Lower and Middle Cambrian. Stose and Jonas (1922, 1923) are a little vague about exactly what the type area of the Conestoga is. The narrowest option, "the valley of the Conestoga River", strictly interpreted as the slopes bordering this somewhat incised stream from the City of Lancaster (Stop 1 and 2 area) to the Martic line near Pequea, is sufficiently precise and provides the best overall exposure. All the facies of this distinctive slope limestone from proximal megabreccia to thin distal turbidites with substantial dark phyllitic partings are represented. It would be impossible to measure a section in the conventional manner here (or elsewhere), but Wise (1953) has computed a structurally corrected thickness aproximately down the axis of the synclinorium of 3000 ± 1000 feet. I calculate the thickness of the steeply dipping section in the western Chester Valley in the lower part of the error range. Ultimately neither figure is really significant of what may have been deposited because the upper boundary is a tectonic contact. All rocks except some of the coarser proximal beds have a distinctive fairly dark blue-grey color attributed to a modest content of reduced organic carbon (Tucker and Wright, 1990). Wise (1953) concluded the local value averaged about 0.75 percent from ignition of insoluble residues. Scattered cubic pyrite up to over a centimeter in size is often fairly common and reinforces the idea of a reducing environment.

These "Lancaster County Blues" of the local vernacular appear throughout the Hanover, York, Lancaster, and western Chester Valleys, and are known in attenuated form at least as far east as the Valley Forge quadrangle. The only known fossils in these rocks are in the previously noted Middle Cambrian occurrence near West York. I contend the basal Conestoga shale is properly assigned to the Cash Smith member of the Kinzers Formation. The great thickness of Lower Cambrian platform margin rocks overlying this unit in the York Valley inplies that much of the Conestoga is Lower Cambrian. The Frederic (Maryland) Limestone indicates rocks of this general type were deposited somewhere at least during all but the lowest Cambrian, but structural constraints suggest that little rock of this type younger than the West York locallity is preserved in Pennsylvania.

Other rocks assigned to the Conestoga Limestone by Stose and Jonas (1923) first appear near Coatesville and become increasingly prominent eastward until they may be the only rocks so named in and near the White Marsh Valley. These rocks differ in lithology and apparent depositional environment from the type Connestoga Limestone and may be entirely younger. A different formation is clearly indicated, and I suggest below the Henderson Marble. With the basal shale below assigned to the Kinzers Formation and the eastern dolomites to the Henderson, the Conestoga Limestone very properly deserves its own lithic identifier rather than the wimpy "Formation" with which it has often been latterly afflicted.

Zooks Corner Formation, Lower(?) and Middle Cambrian. I find this formation a rather enigmatic unit. East of York County, where the highest platform rock is Ledger, the Zooks Corner Formation appears immediately inboard of the Ledger with which it at least partialy intertongues. Near the Lancaster type section Meisler and Becher (1971) report the upper part interbeds with the Elbrook Formation. Rocks of this type are not associated with the basal Elbrook except in close association with the Ledger. I withdraw my implication in the State Correlation Chart (Berg and others, 1983) that the Zooks Corner is merely dolomitized Elbrook. That not only is probably inconsistent with the interbedding reported by Meisler and Becher, but the Zooks Corner has a somewhat different bedding aspect, and, although not especially pure, the regolith does not contain the shale chips that often characterize the Elbrook. It is apparently a distinctive inboard facies of the bank margin assemblage, but how it fits into a marginal shoal to lagoonal transition I do not know.

Lacking a clear understanding of its genesis, I can only assume that it may sometimes be adjacent to Lower Cambrian Ledger, and must be assigned in part a tentative Lower Cambrian age. The type section is quite clearly entirely Middle Cambrian.

Elbrook (Buffalo Springs) Formation, Middle Cambrian. The local name from the village of Buffalo Springs in the Richland quadrangle was established by Gray and others (1958) as the lowest member (base concealed) of the Conococheague Group following Stose and Ljungstedt (1931) who so mapped all of the Lebanon Valley south of the Beekmantown group. The earlier assignment was inevitable, as the "Buffalo Springs" of northern Lancaster County comprises most of the Conococheague mapped by Jonas and Stose (1926, 1930). It was raised to formational status and excluded from the Conococheague (Geyer and others, 1963) on rather speculative lithic corelation of the Snitz Creek Formation to the Big Spring Station member of Wilson (1952), which susequently has been faunally validated. If Meisler and Becher (1968) had been stratigraphers rather than hydrologists, they no doubt would have recognized that they had a complete package of Elbrook lithology and age. The original type Elbrook (Stose, 1908) is perfectly valid even though the name was misapplied to the Zooks Corner Formation of Lancaster Couty (Stose and Jonas, 1922). As it stands, the junior synonym (Buffalo Springs) is redundant and should be abandoned.

The Elbrook formation is a normal marine, predominantly subtidal, limestone deposited on a shelf which was recieving significant input of mainly argillaceous terrigenous clastics (10 to 30 percent based on a limited number of chemical analyses and insoluble residue determinations). Variable proportions of dolomite beds, perhaps representing shoaling up to supratidal, are present but rarely are abundant. The formation is rarely well exposed, but through most of its extent from Maryland to Berks and northern Lancaster Counties it is easily mapped by the abundant shale chips in the regolith. It overlies the the Waynesboro Formation from its type area to eastern Cumberland County where it disappears below the Yellow Breeches thrust. Near Marietta in western Lancaster County, between the Mechanicsville and Chickies thrusts, it is separated from the Ledger Formation by less than 100 feet of dolomitic, silty and sandy rock which expands eastward into the Zooks Corner formation. This quartzose unit may be the distal fringe of the upper Waynesboro sandy member. At the the type section of the Zooks Corner Formation, east of the convergence of the Chickies and Mechanicsville thrusts, Meisler and Becher (1971) report the lower Elbrook (Buffalo Springs) is laterally gradational into the the Zooks Corner Formation. As the Zooks Corner Formation is much more areally restricted and always associated with inner margin of the Ledger formation bank margin facies, the last locality may be the best we have to represent the transition from bank margin to shelf.

Henderson Marble (new name), Middle Cambrian to Lower Ordovician(?).
Rocks formally assigned to the Conestoga Formation appear conformably above the Zooks Corner Formation in the Downington area where the contact is placed below the first thick limestone bed. Much of the overlying sequence is is predominated by thick white to medium light grey marble beds which may be either apparently massive or laminated. They are variably interbedded with grey dolomite beds as thick as 12 feet. The marbles appear generally fairly pure, and they are notably so in some quarry analyses. Little of this stone is presently quarried, but it was much used for ornemental and structural marble as well as lime and possibly railway ballest. These are quite apparently shelf facies rocks, and their only known contacts with rocks of the Conestoga slope facies are tectonic.

The name comes from the Henderson Park area of Upper Merion Twp., Momtgomery County, with specific reference to the former Henderson Station on the Main Line near which the only fossils reported from this unit were found (Bascom and others, 1908). They were identified as silicified brachopods of Lower Ordovician age by apparently reputable authorities. They appeared in float of "vuggy silicious rock" in an area long since (Miller, 1934) covered by development. I have seen float or debris of what must be the same rock in the vicinity of the old marble quarries near Henderson Station Road, and with sufficient peseverance additional specimens might yet be recovered. Much of the area underlain by these rocks has not yet been even superficially reexamined and designation of a particular type section would be premature.

LITHOFACIES AND DEFORMATION HISTORY OF THE OCTORARO FORMATION AND THE RELATIONSHIP TO THE PLEASANT GROVE-HUNTINGDON VALLEY SHEAR ZONE

David W. Valentino
Conrad College
Athens, WV 24712

ABSTRACT

The Octoraro Formation (previously the Octoraro Phyllite) is a vast unit containing numerous schist lithologies in the western Piedmont of Pennsylvania. Recent bedrock geologic mapping in Lancaster County delineated numerous lithologically distinct members of the Octoraro Formation including units of pelitic schist, plagioclase-bearing schist, and units containing interlayered schist and metasandstone. The regional distribution of the members of the Octoraro Formation reveal structural complexity and truncation at the contact with the Conestoga Formation (the Martic Line). As well, members of the Octoraro Formation are structurally truncated at the northern margin of the Pleasant Grove-Huntingdon Valley shear zone.

The Octoraro Formation experienced two phases of deformation and metamorphism that pre-date the formation of the Tucquan antiform. The early event (D1nw and M1nw) is generally preserved in the hinge regions of later folds (F2nw), as inclusion trails within metamorphic porphyroblasts and discrete microscopic domains or micro-lithons. Detailed analysis of the timing of D2nw deformation and M2nw metamorphism revealed that metamorphism pre-dates the accompanying deformation in the region of the north limb of the Tucquan antiform, and the metamorphism post-dates deformation in the region of the southern limb. A complex deformation and metamorphic history is inferred from these relationships that involves northward expansion of D2nw deformation during the M2nw metamorphic episode.

INTRODUCTION

Most of the western Piedmont of southern Lancaster and York Counties and portions of central Chester County, Pennsylvania is underlain by schist. This schist has been referred to in the literature as the Octoraro Phyllite (Bascom and others, 1909), the chlorite-albite facies of the Wissahickon Formation (Knopf and Jonas, 1929), and more recently as the Prettyboy schist (Howard, 1993). The southwestern corner of the geologic map of the Newark $1^{o}x2^{o}$ quadrangle (Lyttle and Epstein, 1987) covered a small portion of this belt of schist, and the early name of Octoraro Phyllite (Bascom and others, 1909) was reinstated for this unit. The Octoraro Phyllite will be referred to as the Octoraro Formation throughout this paper because this unit actually contains a minor amount of phyllite as compared to various schist lithologies.

In York County, Stose and Jonas (1939) subdivided the Wissahickon chlorite-albite schist (Octoraro Formation) into numerous lithologies or "schist facies". In Lancaster County the Octoraro Formation also contains some distinct lithologies that occur at the map scale. Between 1988 and 1992, the Octoraro Formation in southern Lancaster County and western Chester County was mapped at the scale of 1:24,000 and numerous members were delineated (Figure 3A, 3B, and 3C). Although the Octoraro Formation was named for the exposures along the branches of Octoraro Creek, the best exposures occur primarily along the east shore of the Susquehanna River. The type locations for most of the members occur along the Susquehanna River section, with the exception of a few units with type localities along small tributaries to Octoraro Creek.

The Octoraro Formation resides exclusively within the Tucquan antiform structural block. The northern and eastern contact is with marble lithologies of the Conestoga Formation across the Martic Line. The Drumore tectonite of the Pleasant Grove-Huntingdon Valley shear zone (Valentino and others, 1994) forms the southern boundary of the Octoraro Formation (Figure 3C). This paper attempts to characterize the lithologic variability, the distribution of lithologies, and the structural and metamorphic history for Octoraro Formation in Lancaster and Chester Counties, Pennsylvania. The regional relationships concerning the Pleasant

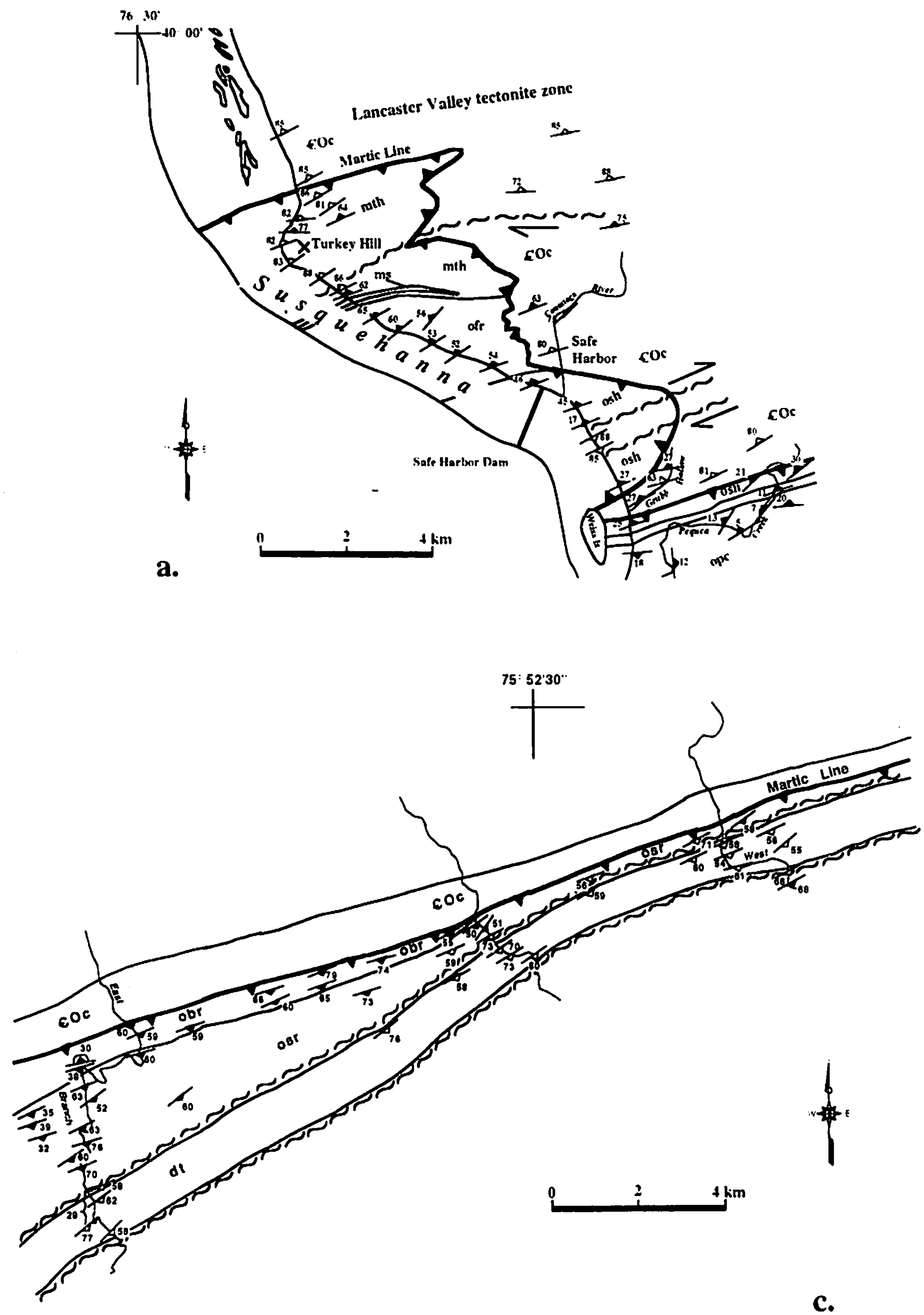
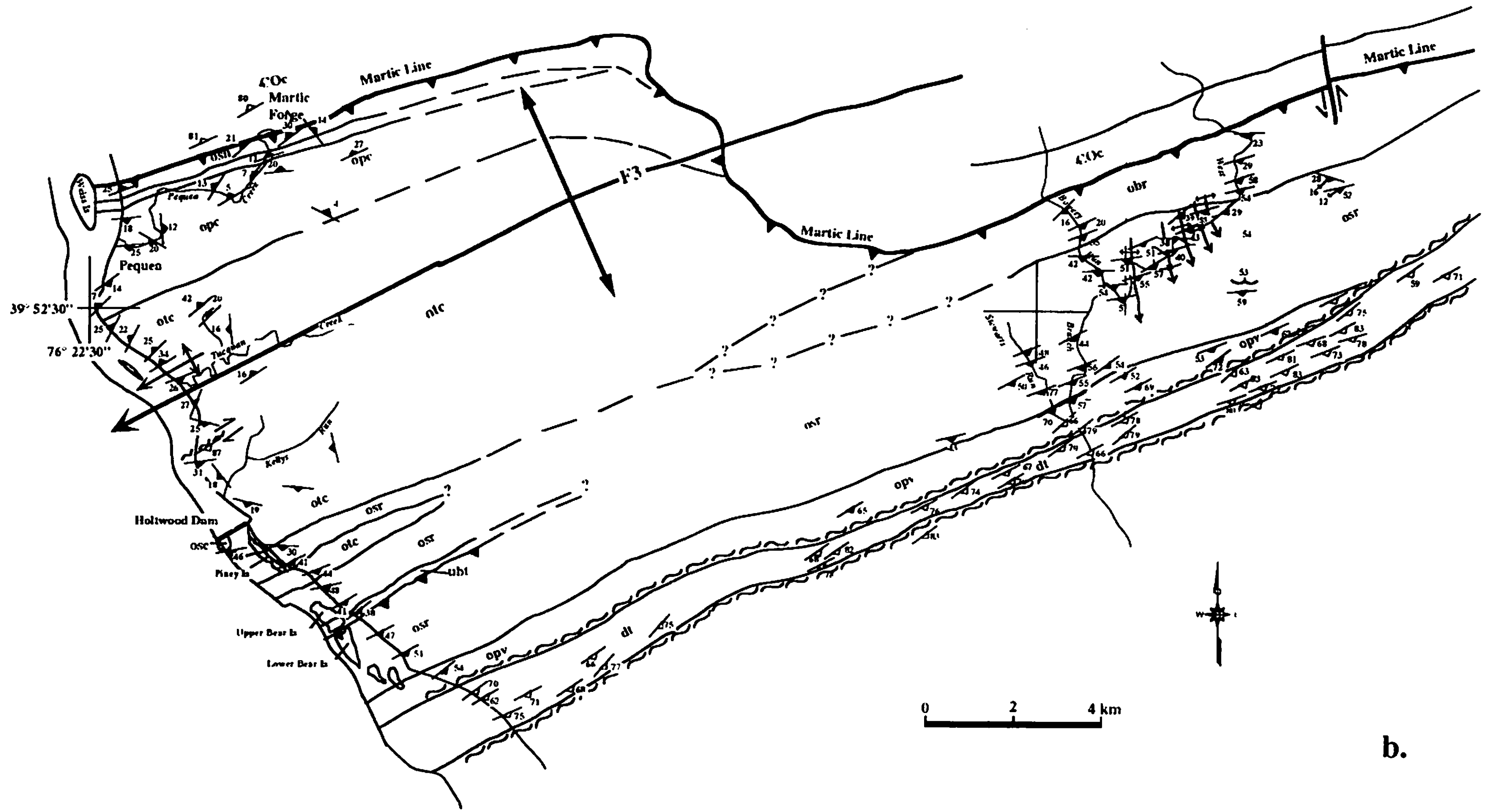


Figure 3. Bedrock geologic maps of the western Piedmont in Lancaster and Chester Counties, Pennsylvania. [A] Octoraro Formation on the northern limb of the Tucquan antiform; [C] Octoraro Formation on the southern limb of the Tucquan antiform east of Octoraro Creek. See text for explanation of unit and lithology descriptions. The solid strike and dip symbols with a half circle represent S1nw schistosity, the solid strike and dip symbols represent S2nw schistosity, the open strike and dip symbols represent S3 schistosity, and sinuous strike and dip symbols represent S4 kink bands.



b.

Figure 3. Bedrock geologic maps of the western Piedmont in Lancaster and Chester Counties, Pennsylvania. [B] Octoraro Formation near the crest of the Tucquan antiform. The solid strike and dip symbols with a half circle represent S1nw schistosity, the solid strike and dip symbols represent S2nw schistosity, the open strike and dip symbols represent S3 schistosity, and sinuous strike and dip symbols represent S4 kink bands.

Grove-Huntingdon Valley shear zone and the local relationships concerning the Martic Line will also be addressed.

THE OCTORARO FORMATION

This summary of the Octoraro Formation stems from a detailed investigation of the exposures along the Susquehanna River by the Pennsylvania Geological Survey. The Octoraro Formation occurs in the Tucquan antiform, and the Tucquan antiform is a deeply eroded broad open fold (F3) with more than 6 km of structural section exposed on each limb at the Susquehanna River. The following discussion includes descriptions of members of the Octoraro Formation starting with the structurally lowest member and proceeding structurally higher on the northern limb of the Tucquan antiform, and then the southern limb. Although the members are discussed in the order in which they are structurally stacked in the section, due to the complex deformation history for this region no stratigraphic relationships are implied.

Unit Descriptions

Core region of the Tucquan Antiform (Structurally lowest member)

Tucquan Creek member (otc): Muscovite-garnet-schist and muscovite-chloritoid schist. The Tucquan Creek member is composed of large scale (10-100's meters) interlayering of silver-gray, fine- to medium-grained mica schist containing the metamorphic mineral assemblages chloritoid-chlorite-muscovite or garnet-chlorite-muscovite. Rarely this unit contains abundant metamorphic plagioclase and biotite. The range of modal mineralogy is: muscovite (20-70%); chlorite (10-20%); quartz (10-30%) with accessory plagioclase, chloritoid, garnet, biotite, and magnetite. The dominant structure is the regional S2nw schistosity, with rare weakly developed S3. However, crenulations and meso-scale F3 folds (some times conjugate box folds) are prevalent throughout this member.

Northwestern Limb of the Tucquan Antiform

Pequea Creek member (opc): Muscovite-garnet-plagioclase schist. The Pequea Creek member is named for the excellent exposures along Pequea Creek west of Martic Forge. This member is a silver to gray medium- to coarse-grained schist. This schist is rich in muscovite, chlorite and plagioclase and generally does not contain biotite. The range of modal mineralogy is as follows: quartz (5-15%), chlorite (20-40%), muscovite (30-70%), plagioclase (5-20%), and accessory magnetite and garnet. The S2nw schistosity is dominant and defined by parallel alignment of muscovite and chlorite. Minute discrete shear surfaces associated with late northeast directed (Valentino, 1990) thrusting are defined by recrystallized chlorite and muscovite.

Martic Forge member (omf): Biotite-microcline-plagioclase metasandstone. The Martic Forge member is named for outcrops that occur along the road near Martic Forge. This unit is a light brown, massive metasandstone containing metamorphic plagioclase and biotite and primary quartz and microcline. Both the S1nw and S2nw foliations are present in this unit and are defined by recrystallized planes of quartz and microcline and muscovite-biotite partings. This member is characterized by 40-50% quartz, 10-15% biotite, 5-15% muscovite, and <10% microcline. The northern contact is with a chlorite-muscovite schist (osh) and the southern contact is with muscovite-garnet-plagioclase schist (opc). The areal width of this unit ranges between 50 and 200 meters along strike. The local Antietam Formation of the Martic Hills is lithologically similar to this member. There are also lithologic similarities with the Bowery Run member (obr) on the southern limb of the Tucquan antiform.

Safe Harbor member (osh): Muscovite-biotite-plagioclase schist. The Safe Harbor member is named for the exposures near the Safe Harbor Dam in Lancaster County. This unit is a silver to gray, medium- to coarse-grained schist with abundant grains of interlocking plagioclase. The rock contains muscovite (30-50%), biotite (5-20%), and plagioclase (20-40%), but is generally lacking in chlorite (<10%). Quartz percentage ranges from 10 to 40%, but averages 25%. The dominant structure is the regional S2nw schistosity and the S1 schistosity is preserved where F2nw isoclinal folds are present. The S2nw and S1nw schistosities are defined by the planar alignment of muscovite and biotite, and planar aggregates of interlocking plagioclase.

Fisherman Run member (ofr): Muscovite-plagioclase-quartz phyllitic-schist and schist. The Fisherman Run member is named for the exposures of this lithology along Fisherman Run which is a small tributary to the Susquehanna River. This member is a silver to pale green, fine-grained schist with the highest muscovite content in the Octoraro Formation, commonly in excess of 50% by volume. Average modal percentages are: muscovite (40-50%), chlorite (<10%), quartz (15-50%), plagioclase (5-40%), and biotite (0-15%). The Fisherman Run member can be subdivided into a structurally upper and lower unit based on the relative quartz and muscovite content. The lower unit is generally richer in quartz while the upper unit contains more abundant muscovite. Both the regional S2nw and S3 schistosities are present, however S3 is weak and only locally developed. The S2nw schistosity is defined by the parallel alignment of micas which is also parallel to quartz veins.

Marburg Formation (mth). The Marburg Formation occurs on the northern flank of the Tucquan antiform mostly in York County. The eastern projection of this unit enters Lancaster County at near the area of Turkey Hill. The Marburg Formation has a gradational contact with the structurally lower Fisherman Run member of the Octoraro Formation. Although the northern and eastern contact with the Conestoga Formation is not exposed, the lithology distribution suggests that the contact is complexly folded (Wise, 1970). In the area of Turkey Hill the Marburg Formation comprises two distinct lithologies that are broken out as members: (1) Chlorite-muscovite phyllonite; and (2) Massive plagioclase-chlorite schist (Figure 1A).

At Turkey Hill the dominant lithology is chlorite-muscovite phyllonite. This unit contains pale green to silver fine- to medium-grained phyllite and schist. These rocks are generally rich in muscovite, chlorite, and quartz and low in plagioclase. The range of rock modal mineralogy is as follows: quartz (25-45%), chlorite (5-45%), muscovite (5-40%), plagioclase (0-30%), and accessory magnetite, pyrite, and tourmaline. This unit is dominated by the regional S3 cleavage and contains frequent discontinuous small quartz veins. There are rare, 10 cm to 1 m thick, metasandstone layers composed of recrystallized quartz and calcite. This unit is named the Turkey Hill member of the Marburg Formation.

Within the Turkey Hill member there is a single layer of massive plagioclase-chlorite schist that contains planar aggregates of metamorphic plagioclase (M2nw). The planar aggregates of plagioclase give this rock a gneissic texture, although the metamorphic mineral assemblage is only chlorite-muscovite-plagioclase. Plagioclase porphyroblasts range from sub-millimeter to as large as 10 mm in diameter. This unit is approximately 70 to 85 m thick and has sharp contacts with the local fine grained chlorite-muscovite schist to the north and south. The arcuate trace of a small ridge east of the Susquehanna River possibly defines the eastern extent of this unit, but there is no exposure away from the river.

Southeastern Limb of the Tucquan Antiform

Upper Bear Island tectonite (ubt): Muscovite-sericite phyllonite. A black to gray, very fine-grained micaceous rock with a slaty cleavage that is parallel to the regional S2nw schistosity traces through Upper Bear Island at the Susquehanna River, and is also exposed along the railroad cut on the east shore of the river. This rock is composed of muscovite/sericite (20-35%), chlorite (10-15%), quartz (35-40%), biotite (5-10%), plagioclase (10-20%), and accessory pyrite and ilmenite. The upper and lower contacts of this tectonite are abrupt with plagioclase-mica schist lithologies. The approximate structural thickness is 190 m.

Sams Creek metabasalt (osc). A sequence of metabasalt 100-150 m thick is exposed in the fish ladder on the west side of the Holdwood Dam. Regionally the only other metabasalt that occurs within the Octoraro Formation is the Sams Creek metabasalt of central York County and northern Harford County, Maryland. Three metabasalt lithologies are exposed in the fish ladder: (1) light-green, fine-grained granular foliated rock, (2) thin (mm-scale) alternating chlorite-magnetite-rich and epidote-plagioclase-rich layers, and (3) silver-green color chlorite-epidote-rich rock with minor muscovite. These three lithologies are in order from structurally lowest to highest. The lower contact is not exposed at the base of the Holtwood Dam, but the upper contact is gradational with the local mica-plagioclase schist (over a thickness of about 10 m. A large block of mica schist outcrops within the metabasalt just to the east of the fish-ladder containing wall. The upper and lower contacts between this block and the metabasalt are concordant and parallel to the regional schistosity while the eastern contact is a subvertical fault. Foliation (S2nw) in the metabasalt is defined by planar mineral aggregates of epidote, amphibole, plagioclase, and phyllosilicates. Although this lens of metabasalt is not continuous with any of the layers of Sams Creek metabasalt to the west and northwest, the lithologic similarity and association with the Octoraro Formation places this unit in the Sams Creek metabasalt.

Bowery Run member (obr): Muscovite-quartz-garnet schist. The Bowery Run member is named for the exposures along Bowery Run in eastern Lancaster County. This unit contains silver to tan quartz-muscovite schist interlayered on the meter scale with tan feldspathic metasandstone. The metasandstone layers are composed of fine- to medium-grained detrital quartz and feldspar. The modal mineralogy for the schist portions of this unit is: quartz (10-40%), muscovite (30-60%), chlorite (5%), and accessory garnet, biotite, magnetite, and pyrite. The northern boundary with the Conestoga Formation and the southern boundary with plagioclase-mica schist (osr) of the Octoraro Formation are parallel to the regional S2nw schistosity. The contact with the Conestoga Formation is the Martic Line, interpreted to be a thrust fault by earlier workers. The Bowery Run member is dominated by S2 schistosity that is parallel to the compositional layering in most places. However, the S1nw schistosity is preserved in a few places, particularly along the East Branch of Octoraro Creek. The Bowery Run member is lithologically similar to the Antietam-Harpers Formations undivided (Knopf and Jonas, 1929) located on the southern flank of the Mine Ridge Grenvillian massif.

Stewarts Run member (osr): Muscovite-chlorite-biotite- plagioclase schist. The Stewarts Run member is named for exposures along Stewarts Run in eastern Lancaster County, but this unit is also well exposed at the Susquehanna River. This member of the Octoraro Formation is a green-gray muscovite-chlorite-biotite-plagioclase schist. Variation, such as the presence of very fine muscovite and chlorite in local abundance, causes the rock to have a locally phyllitic texture. Plagioclase porphyroblasts are abundant, ranging from 1 to 5 mm in diameter. Mineralogy is as follows: muscovite (15-50%, average approximately 30%), chlorite (15-20%), plagioclase (0-60%, average approximately 30%), ilmenite (1-5%), and accessory biotite, zoisite, calcite, and pyrite. Rare quartz-rich layers were observed at a few localities. Parallel alignment of phyllosilicates defines the regional S2nw schistosity, and abundant small isoclinal folds are usually defined by folded quartz veins. The lower contact is with the Bowery Run member (obr) in the east and (otc) in the west, and the upper contact is well exposed at the Susquehanna River and at Stewarts Run and appears to be very sharp with the Puseyville member (opv).

Puseyville member (opv): Muscovite-plagioclase schist with quartzite beds. The Puseyville member is named for exposures along the West Branch of Octoraro Creek near Puseyville in Lancaster County. This unit contains silver-gray to silver-green schist bearing abundant quartz and plagioclase porphyroblasts and layers of metasandstone that are tens of centimeters thick. There is quartzite-pelite interlayering on the centimeter to decimeter scale. The metasandstone layers range from a few centimeters to half a meter in thickness. Round blue quartz grains are present in the metasandstone layers (millimeters in diameter). Although these layers are quartz-rich, considerable muscovite is present (up to 25% in some samples). There are occasional phyllitic sections within the bedded metasandstone outcrops. Phyllosilicates and quartz aggregates define the internal S2nw schistosity within the metasandstone layers . The northern contact with the Stewarts Run member (osr) is gradational over tens of meters while the southern contact with the Drumore tectonite (dt) is abrupt.

Octoraro Formation map patterns

The members of the Octoraro Formation parallel the trends of the limbs of the Tucquan antiform. In general the strike of the units on the southern limb of the antiform ranges from 055^{o} near the Susquehanna River to about 070^{o} near the East Branch of Octoraro Creek and they consistently dip moderately toward the southeast (Figure 3C). The strike of the units on the northern limb of the antiform ranges from 250^{o} near the Pequea to 260^{o} near Safe Harbor and they dip moderately toward the northwest (Figure 3A). Although the contacts between members of the Octoraro Formation can not be traced continually eastward away from the Susquehanna River, mapping using the limited exposure in small creeks and roadcuts, and float made it possible to estimate the regional distribution of the lithologies. On the northern limb and crest of the Tucquan antiform it appears that the members of the Octoraro Formation that were observed at the Susquehanna River section continue eastward to the contact with the Conestoga Formation. The contact with the Conestoga Formation (Martic Line) truncates the various units in the Octoraro Formation (Figures 3A and 3B). On the southern limb of the Tucquan antiform it was possible to trace some members of the Octoraro Formation as far east as the West Branch of the Brandywine River (Figure 3C). The southernmost member, the Puseyville member, is structurally truncated against the Drumore tectonite between the West and East Branches of the Octoraro Creek. The Stewarts Run member is truncated farther east along the same tectonite zone.

Correlation of lithologies across the Tucquan antiform

The gross lithologies of the Octoraro Formation can be "correlated" across the crest of the Tucquan antiform. The general trend in lithology composition is an increase in the quartz content and decrease in mica content in the various members from the structurally lowest section in the core of the Tucquan antiform (Tucquan Creek member, otc) structurally upward on the limbs of the antiform to the north and south. The Marburg Formation, located on the northern limb of the antiform, contains chlorite-muscovite schist and rare metasandstone layers, and is compositionally similar to metasandstone-bearing units on the southern limb that occur at approximately the same structural level (the Bowery Run and Puseyville members). There are a few different types of plagioclase-bearing schist that were mapped, but the general distribution of plagioclase-bearing schist is restricted to the middle section on both limbs of the antiform. The Stewarts Run member resides structurally above the pelitic-schist of the Tucquan Creek member, and below the metasandstone-bearing Puseyville member at the Susquehanna River and the metasandstone-bearing Bowery Run member near the East Branch of Octoraro Creek. The plagioclase-bearing units on the northern limb (Safe Harbor, Martic Forge, and Pequea Creek members) occur structurally above the Tucquan Creek member and below the Marburg Formation and pelitic schist of the Fisherman Run member. Most likely this gross sequence of lithologies on the northern and southern limbs of the antiform are equivalent, and any detailed differences are probably the result of lateral variations in composition.

DEFORMATION HISTORY SUMMARY FOR THE OCTORARO FORMATION

There is evidence for two early phases of deformation and associated metamorphism in the Octoraro Formation that predate the development of the Tucquan antiform. Freedman and others (1964) identified two schistosities in the rocks of the Tucquan antiform structural block and called them as $S_{0.5}$ and S1, and they were re-classified as S1nw and S2nw by Valentino and others (1994), because they exist exclusively northwest (nw) of the Pleasant Grove-Huntingdon Valley shear zone (this new classification of structural nomenclature is discussed in detail in Valentino and others, 1994, and Valentino and Faill, this guidebook).

The S1nw schistosity consists of muscovite, chlorite, aggregates of quartz and plagioclase, and thin quartz veins. The S1nw schistosity generally occurs in the field in the hinges of meter-scale F2nw folds, and at rare outcrop-scale occurrences the S1nw is parallel to compositional layering. A common occurrence of S1nw is as planar to complex inclusion trails in M2nw plagioclase and garnet porphyroblasts, and micro-lithons. The S2nw schistosity overprints the S1nw schistosity, and the S2nw schistosity is axial planar to F2nw isoclinal folds defined by folded S1nw schistosity. The Tucquan antiform is defined by a regional arch of the S2nw schistosity.

The most common metamorphic mineral assemblage associated with S1nw is chlorite-muscovite-epidote$\pm$plagioclase, but on the southern limb of the Tucquan antiform M1nw garnet was observed as inclusions in M2nw plagioclase. The M2nw metamorphism ranges from chlorite-muscovite-plagioclase assemblage on the limbs of the Tucquan antiform through garnet-chlorite-plagioclase assemblage in the core of the antiform and structurally lowest region of S2nw schistosity (Valentino and Faill, this guidebook). A distinct metamorphic discordance exists between M1nw to M2nw assemblages across the Tucquan antiform at the Susquehanna River. In the core of the Tucquan antiform, M1nw is defined by chlorite-muscovite-epidote assemblage where the M2nw assemblages are chlorite-garnet or biotite-chlorite-chloritoid (Figure 4). On

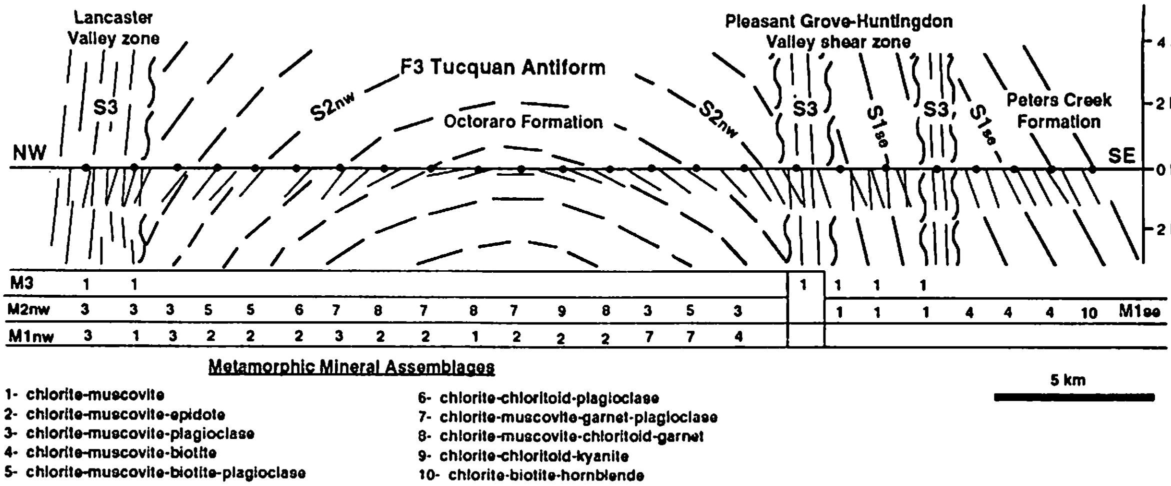


Figure 4. Simplified cross section of the Tucquan antiform along the Susquehanna River with metamorphic mineral assemblages shown (after Valentino and others, 1994).

the northern limb of the antiform, structurally above the M2nw biotite zone, both M1nw and M2nw are defined by the assemblage chlorite-muscovite-plagioclase and are difficult to separate based on metamorphic mineral assemblage. However, the M1nw assemblage in the some regions of the southern limb contains chlorite-muscovite-garnet in areas where the M2nw assemblage is chlorite-muscovite-plagioclase. These regional relationships clearly demonstrate discordance between M1nw and M2nw and suggest they are two discrete episodes of metamorphism and associated structures.

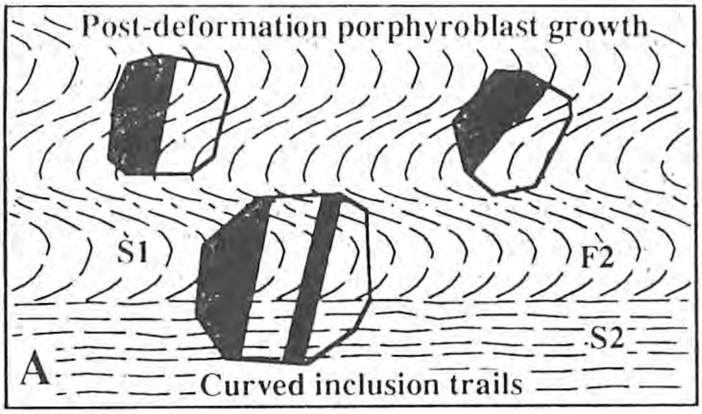
Discordance between D2 and M2 relative timing

In this area where regional deformation (D2nw) and metamorphism (M2nw) are grossly synchronous detailed analysis revealed that systematic differences in the relative timing between deformation and metamorphism actually exist. There is evidence that the M2nw metamorphic episode that occurred pre-, post-, and syn-deformation (D2nw). The diachronous deformation and metamorphism influenced (1) the type of schistosity that developed locally in the Octoraro Formation, (2) the mechanisms of deformation, and (3) the geometry of inclusion trails inside metamorphic porphyroblasts of plagioclase and garnet.

On the southern limb of the Tucquan antiform the S2nw schistosity is defined by rotated micas with minor recrystallization of muscovite and chlorite that occurs in randomly oriented fibrous patches. M2nw metamorphic plagioclase porphyroblasts contain spiral-shaped and fold-shaped inclusion trails, the dominant matrix schistosity (S1nw) is continuous with the inclusion trails, and the shape of the inclusion trails is identical to folds in the matrix (Figure 5A). These textures suggest that M2nw metamorphism and porphyroblast growth succeeded D2nw deformation on the southern limb of the Tucquan antiform. On the northern limb of the Tucquan antiform plagioclase porphyroblasts contain primarily straight inclusion trails of the S1nw schistosity, the external S2nw schistosity wraps the porphyroblasts, new (M2nw) micas are often randomly oriented, the S2nw schistosity is defined by rotated old (M1nw) and new (M2nw) micas, M2nw plagioclase porphyroblasts are often broken with jagged edges, and isoclinal folds (F2nw) in the matrix do not continue as inclusion trails through the porphyroblasts (Figure 5B). These textures suggest that M2nw metamorphism and porphyroblast growth preceded D2nw deformation.

In the region transitional to the northern and southern structural domains, near the crest of the Tucquan antiform, the relationship between metamorphism and deformation is complex. The M2nw schistosity is defined by recrystallized and rotated primary micas, dynamically recrystallized quartz, and planar aggregates of M2nw porphyroblasts. Inclusion trails in M2nw plagioclase within the same specimens are straight, straight with abrupt internal truncations, curved, and combinations of both. These complex relationships in this region suggest that M2nw metamorphism was synchronous with D2nw deformation locally.

The local relative timing relationships between M2nw metamorphism and D2nw deformation to



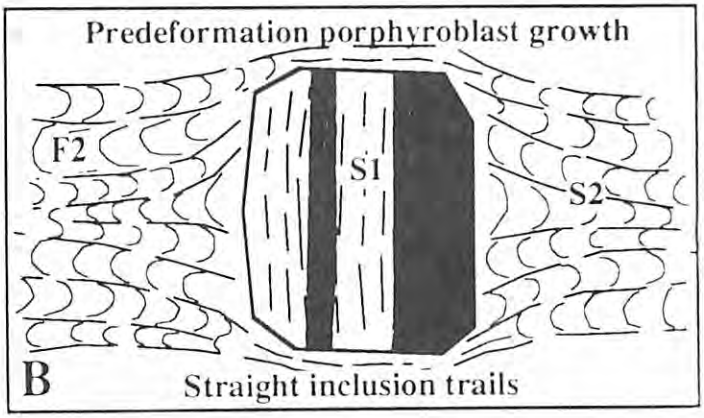


Figure 5. Sketches of the relationships between deformation features and growth of metamorphic porphyroblasts in the Octoraro Formation. [A] Plagioclase porphyroblasts that have overgrown microfolds suggesting metamorphism post-dates deformation; and [B] plagioclase porphyroblasts with straight inclusion trails that preserve the undeformed S1nw, and the S2nw wraps the porphyroblast suggesting metamorphism that pre-dates deformation.

result in early metamorphism in the north followed by deformation and metamorphism in the south can be interpreted as (1) a migrating thermal front from north to south across the Octoraro Formation prior to development of the Tucquan antiform, (2) D2nw deformation initiating in the south with northward expansion of the deformation front, or (3) a combination of the two models. Since at common structural levels in the Tucquan antiform the degree of metamorphism is virtually the same with increasing grade deeper in the structure (Valentino and Faill, this guidebook), the north-south migration of a thermal front during deformation is unlikely. From detailed petrology, geothermometry, and the distribution of metamorphic mineral assemblages Valentino and Faill (1993) and Valentino and Faill (this guidebook) concluded that the Octoraro Formation was exposed to a thermal source from below. Deformation probably initiated in the southern area and expanded toward the north during a metamorphic episode with a thermal source from below. This deformation history is consistent with north-northwestward emplacement of thrust slices and nappes as suggested by earlier workers (Cloos and Heitanen, 1941; Freedman and others, 1964; Wise, 1970).

CONCLUSIONS

The following conclusions concerning the Octoraro Formation in Lancaster County can be made based on regional mapping integrated with detailed structural and metamorphic petrology:

1. The name Octoraro Phyllite for the vast expanse of schist in the western Piedmont of Pennsylvania is misleading because most of the region is underlain by schist lithologies, therefore the unit is referred to as the Octoraro Formation.
2. Numerous lithologically distinct members of the Octoraro Formation were delineated

during bedrock mapping.

3. The regional distribution of the members of the Octoraro Formation reveal structural truncations at the Martic Line and at the northern margin of the Pleasant Grove-Huntingdon Valley shear zone.

4. The Octoraro Formation experienced two phases of deformation and metamorphism that pre-date the formation of the Tucquan antiform.

5. Detailed analysis of the timing of D2nw deformation and M2nw metamorphism revealed that metamorphism pre-dates and post-dates deformation in the north and south respectively suggesting that propagation of D2nw deformation proceeded from south to north.

REMNANTS OF AN IAPETAN RIFT BASIN IN THE WESTERN PIEDMONT OF PENNSYLVANIA

David W. Valentino[1] and Alexander E. Gates[2]
(1)Concord College
Athens, WV 24712
(2)Department of Geology
Rutgers University, Newark, NJ 07102

ABSTRACT

Rift basins filled with metamorphosed siliciclastic sediments presently occur in part of the crystalline core of the Appalachians. These rift basins developed as the result of Late Proterozoic-Early Cambrian extension of Grenvillian lithosphere. The Peters Creek Formation of southeastern Pennsylvania was deposited in one of these rift basins, and was subsequently deformed and metamorphosed during Paleozoic orogenic events. For most of the Pennsylvania Piedmont the regional deformation and metamorphism precludes stratigraphic and sedimentologic investigations in the metasiliciclastic rocks; however, portions of the Peters Creek Formation contains enough primary sedimentary features to warrant a detailed study. There are three Peters Creek Formation lithofacies: (1) metamorphosed graded sandstone beds, (2) quartzose metapelite, and (3) discrete massive metasandstone lenses within the graded bedded sequences. Rift-related depositional tectonic setting was inferred by the occurrence of greenstone interlayered with feldspathic metasandstone in the uppermost part of the Peters Creek Formation. Detailed lithofacies and limited stratigraphic study revealed that the Peters Creek Formation consists of at least two submarine turbidite-fan systems represented by thick sequences of interlayered feldspathic metasandstone and schist, separated by a region underlain by quartzose schist. The Peters Creek Formation contains some lithofacies similar to Upper Proterozoic rift sequences in the southern and northern Appalachians, and the discovery of potential Iapetan rift clastics in the Piedmont of Pennsylvania narrows the gap between the southern and northern rift basins.

INTRODUCTION

Late Proterozoic-Early Cambrian extension of Grenvillian lithosphere produced rift basins. The Iapetan rift-related metasediments and volcanics occur primarily in two belts, one in the Blue Ridge anticlinorium of the southern Appalachians and the other in western New England extending into Maritime Canada (Rankin, 1975). Based on the lack of rift clastics recognized in the Pennsylvania-Maryland Piedmont (Thomas, 1977), the two belts were connected by an interpreted Iapetan transform fault. The interpreted Iapetan transform is located in the core of the Pennsylvania reentrant in the Piedmont. Much of the metasedimentary rock in the Pennsylvania Piedmont is considered accretionary melange associated with Taconian convergence of the Chopawamsic-Wilmington Complex magmatic arc with the Laurentian margin (Drake and Morgan, 1981; Horton and others, 1989). Rift-related metasediments also occur in the Pennsylvania Piedmont, particularly the Peters Creek Formation (Gates and Valentino, 1991), and these rift-related metasediments may provide a tectonic link between rift assemblages of the Blue Ridge and New England.

THE PETERS CREEK FORMATION

During the early mapping of the Piedmont in Pennsylvania, the Peters Creek Formation was considered a schist unit that conformably overlies the Wissahickon schist and is overlain by the Cardiff conglomerate and Peach Bottom slate in Pennsylvania (Knopf and Jonas, 1929). More recently the Peters Creek Formation was interpreted as a metamorphosed sequence of accretionary melange associated with the closing of the Iapetus ocean during the Taconic orogeny (Horton and others, 1989). Detailed mapping of the Peters Creek Formation (Figure 6) for this study resolved some of the uncertainties in the regional stratigraphy. The stratigraphy internal to the Peters Creek Formation, lithofacies distribution, and tectonic setting of deposition are the focus of this paper and will be discussed at stops on the field trip.

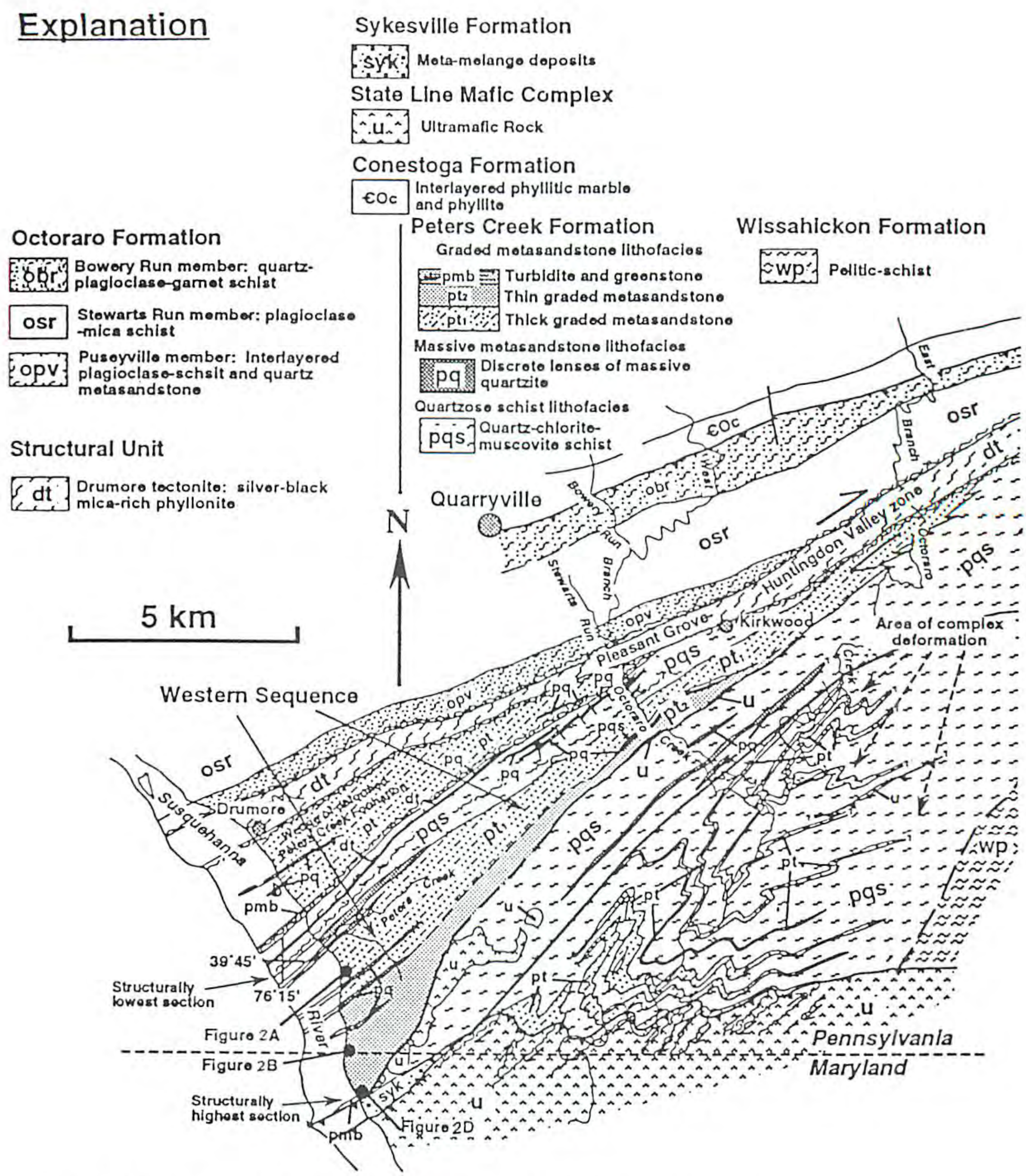


Figure 6. Bedrock geologic map of the Peters Creek Formation and adjacent units, Lancaster and Chester Counties, Pennsylvania and Cecil County, Maryland.

The Peters Creek Formation is a moderately southeast-dipping sequence of siliciclastic metasedimentary rocks in the central and western Piedmont of Pennsylvania (Figure 6). The exposed structural thickness is approximately 6 km near the Susquehanna River and thins eastward to approximately 1.5 km at the East Branch of the Brandywine River. The Peters Creek Formation is largely bounded by ductile shear zones within the Philadelphia-Baltimore structural block and was juxtaposed with the Octoraro Formation during post-Taconian dextral displacement on the Pleasant Grove-Huntingdon Valley shear system (Gates and Valentino, 1991; Valentino and others, 1994). The State Line mafic complex and Sykesville melange are thrust over the Peters Creek Formation on the south and southwest (Gates, Muller, and Valentino, 1991). A strike-parallel gradational contact is interpreted between the Peters Creek and pelitic schist of the Wissahickon Formations in the southeast, but farther east this contact is coincident with the Cream Valley thrust zone (Wiswall, 1990).

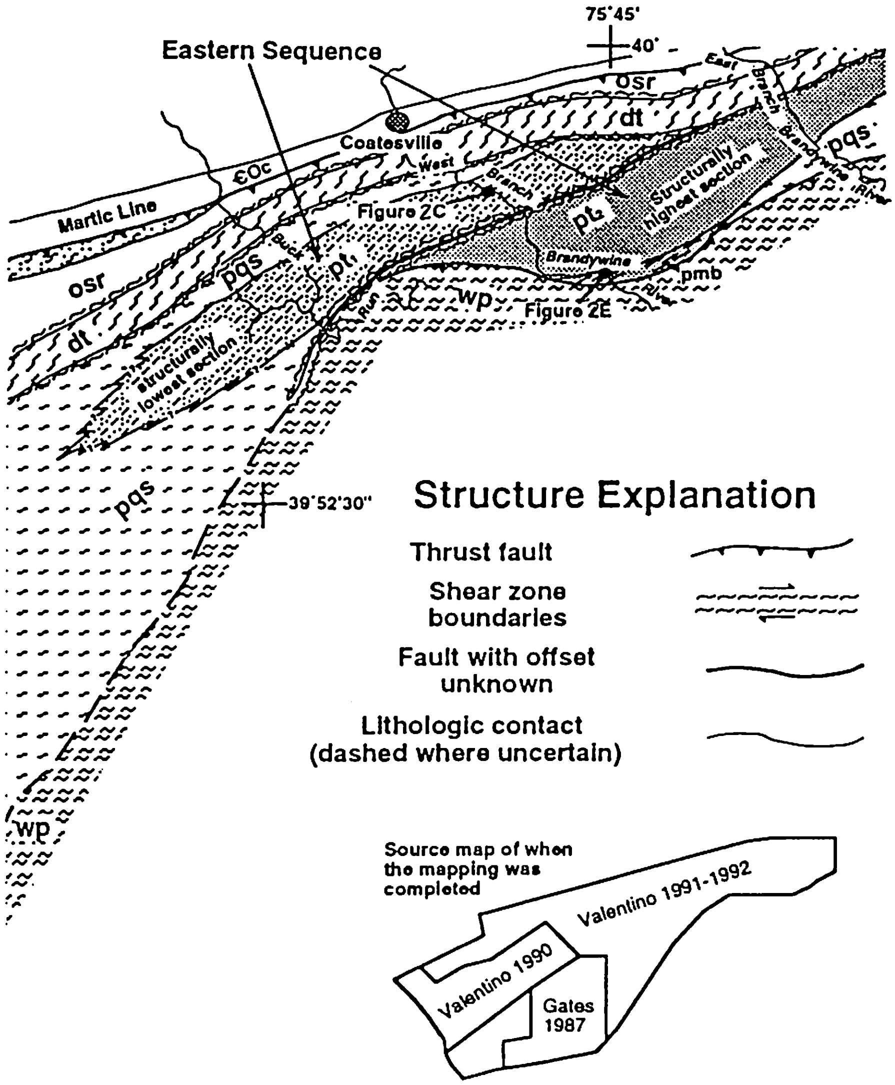


Figure 6. Continued.

The structurally lowest 3.5 km of the Peters Creek Formation between the Susquehanna River and East Branch of the Brandywine River forms a monoclinal structure defined by southeast-dipping bedding-parallel schistosity (S1se: see Table 7, p. 89) with rare small intrafolial isoclinal folds (F1se). A younger weak phase of extensional deformation produced folds and cleavages at a high angle to the compositional layering (Freedman and others, 1964; Valentino, 1993). Although deformed and metamorphosed, younging criteria such as compositional and grain-size grading indicate that the structurally lowest section is right-side-up with only minor reversal where intrafolial isoclinal folding exists.

Strike-slip repetition of part of the lower sequence also occurred in the area of the Susquehanna River where a wedge shaped body of intensely deformed Peters Creek is located between two anastomosing strike-slip shear zones (between the Drumore tectonite and the Peach Bottom structure: see the chapter on the Peach Bottom structure in this volume for more details) related to the Pleasant Grove-Huntingdon Valley system (Valentino, 1993; Valentino and others, in press). The rocks located adjacent to the State Line complex (structurally highest part) were intensely and multiply deformed and metamorphosed to garnet grade (Gates and others, 1991; Gates and Valentino, 1991), and stratigraphic analysis would not be beneficial in that region. Discussions in this chapter on sedimentary geology of the Peters Creek Formation focuses on the structurally lowest 3 to 4 km of the section that are least deformed and least metamorphosed.

Lithofacies descriptions

Three metasedimentary lithofacies occur in the Peters Creek Formation: (1) quartzose schist, (2) metamorphosed graded sandstone beds, and (3) discrete massive metasandstone lenses within the graded bedded sequences.

Quartzose schist lithofacies. The quartzose schist lithofacies occurs in numerous regions of the Peters Creek exposures: (1) immediately south of the belt of the Peach Bottom structure in Lancaster County, (2) the southern part of the West Branch of Octoraro Creek drainage basin, (3) the region between Buck Run and the West Branch of Brandywine River, and (4) the region between Buck Run and the East Branch of Octoraro Creek. The structurally lowest sections of quartzose schist (numbers 1 and 3 above) range from 400 to 1200 m in thickness. This lithofacies is generally characterized by silver-green, fine- to medium-grained schist bearing quartz (20-40%), chlorite (10-30%), muscovite (20-30%), and accessory tourmaline, magnetite, and ilmenite, and detrital grains of potassium feldspar, plagioclase and rutilated quartz. Rare metasandstone layers 0.1-1.0 m thick occur within the quartzose schist lithofacies. Contacts of quartzose schist lithofacies with the graded metasandstone lithofacies are gradational both laterally and across the strike of the Peters Creek Formation.

Graded metasandstone lithofacies. Metasandstone and metapelite interlayered at meter scale make up much of the Peters Creek Formation. Metasandstone-metapelite couplets ranging from less than a meter to 5 m in thickness are composed of graded metasandstone layers (decimeter- to meter-thick) separated by quartzose schist (decimeter-thick) and are commonly capped by a thin layer of mica-schist (centimeter- to decimeter-thick). The metasandstone layers are composed of rounded blue and gray quartz (55-75%), perthitic potassium feldspar (15-45%), plagioclase (< 10%), and accessory zircon, muscovite, chlorite, biotite, epidote, and opaques.

The variability of the interlayered rock types is represented by detailed measured sections from exposures of graded metasandstone (Figure 7). Individual metasandstone beds commonly have a tabular or sheetlike geometry, but individual layers vary in thickness locally. The thickest graded metasandstone layers are in the lower parts of the sequence, and they range in thickness downward from as much as 5 m. The thinnest metasandstones (less than a meter thick) are most abundant in the middle and upper portions of the sequence.

Greenstone is interlayered with the metasandstone and schist in the upper part of the sequence of least deformed Peters Creek Formation (Figures 7D and 7E). The greenstone layers are medium- to fine-grained, pale to deep green, and contain abundant epidote, zoisite, and chlorite with minor plagioclase, tremolite, hornblende, biotite, sphene, magnetite, ilmenite, and quartz. Discrete greenstone layers range from 0.1 to 3 m in thickness, and the contacts between the metasediments and the greenstones are generally abrupt, although some gradational contacts were observed. The greenstone layers are interpreted as metamorphosed mafic tuffs or volcaniclastic sediments.

Massive metasandstone lithofacies. The massive metasandstone lithofacies occurs in lens-shaped, white-gray and tan-brown feldspathic quartz arenites ranging from 10 m to as much as 100 m in thickness, and hundreds to thousands of meters long. The most prominent of the feldspathic quartz arenite bodies was traced along strike for more than 6 km. The smallest ones occur as bodies tens of meters thick and hundreds of meters long (Figure 6). The feldspathic quartz arenite bodies contain abundant detrital grains of rounded blue and clear quartz, rutilated quartz, perthite, and microcline with minor detrital zircon. Cryptic bedding is suggested by internal lamination or minor compositional layering, but, these laminations are generally parallel to metamorphic foliation (S1se) defined by parallel alignment of micas and flattened quartz grains. The primary modal mineralogy ranges as follows: quartz (55-95%), K-feldspar (20-45%), plagioclase (5-10%), muscovite, chlorite, and minor biotite. Most commonly the grain size is gradational between sand and granules. The massive metasandstone lenses occur throughout the graded metasandstone lithofacies but do not occur in the regions of quartzose schist. The lateral terminations of these bodies are abrupt, as are the sharp upper and lower contacts with the graded metasandstone lithofacies. Although the composition and grain size of the massive metasandstone bodies is variable, there

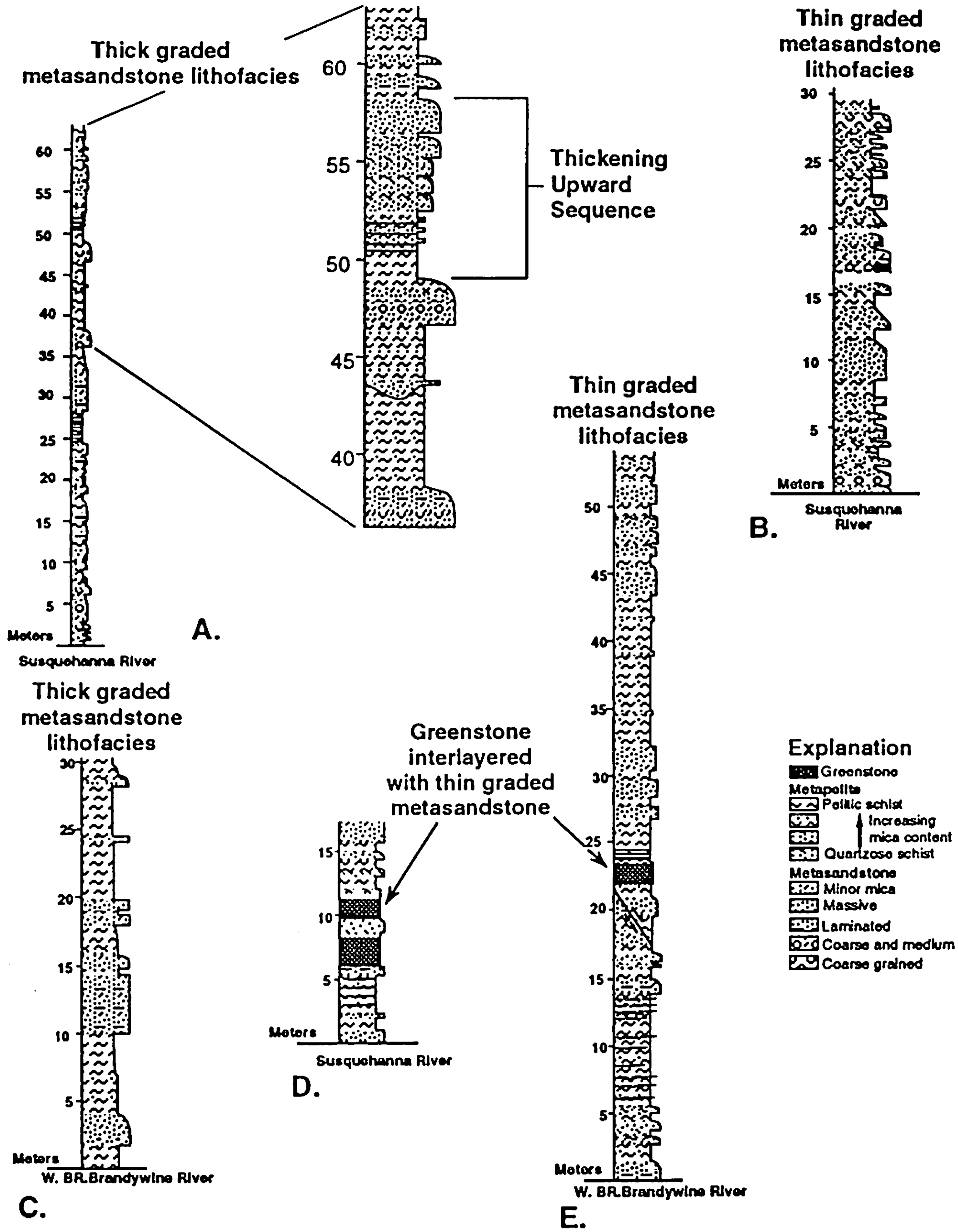


Figure 7. Sample reference columns for the various graded metasandstone lithofacies in the Peters Creek Formation. **[A]** Thick-bedded graded metasandstone lithofacies from the Susquehanna River section, **[B]** Thin-bedded graded metasandstone lithofacies from the Susquehanna River section, **[C]** Thick-bedded graded metasandstone lithofacies from the West Branch of Brandywine River section, **[D]** and **[E]** Thin-bedded graded metasandstone lithofacies with interlayered greenstone from the Susquehanna and West Branch of Brandywine Rivers, respectively. Field locations for these reference columns are shown on Figure 6.

is a systematic geographic distribution of the varied types. Quartzite bodies in the structurally lowest part of the Peters Creek Formation are quartz vein pebble and cobble conglomerates [the Cardiff conglomerate of Knopf and Jonas (1929) but here included within the Peters Creek Formation]. In the middle part of the sequence the bodies are dominated by quartz with minor feldspar of granule- to sand-size grains. The bodies in the upper part of the sequence contain as much as 45 percent feldspar and mica, and sand-size grains are dominant.

Protoliths and depositional setting

The graded metasandstone lithofacies represents turbidite deposits, judged by the grading within individual beds, the scale on which this grading occurs, and the systematic compositional variation of the lithologic couplets. Although exposures of individual turbidites are limited, sand layers in the turbidite deposits have a sheet-like geometry and are continuous in outcrop across tens of meters. Thickening-upward turbidite sequences are portrayed in the generalized turbidite-fan lobe model of Mutti (1977) as a transition from lobe-fringe facies to lobe facies. Outcrop scale sequences commonly show systematic thickening-upward trends of turbidites 1 to 5 m in thickness. Because the massive metasandstone lenses occur primarily within the turbidite deposits, therefore there is probably a genetic link between these two lithofacies. The sandstone lenses generally lack internal sedimentary structures except for some gradation in grain size, and compositions are generally more mature than the surrounding turbidites, particularly in the structurally lowest and middle part of the sequence. Thick coarse-grained sandstone bodies within a turbidite succession suggest channel deposits composed of amalgamated beds, however the exclusive occurrence of mature clastics in many of these deposits suggest they may represent gravity-flow deposits of material that was reworked. The quartzose schist and other metapelitic rocks with rare metasandstone layers are interpreted as interbedded siltstone and shale, and the rare metasandstone layers probably reflect coarser grained turbidite deposition.

Stratigraphic analysis and tectonic provenance

Lithologic columns were constructed for the Peters Creek Formation (Figure 8) at the following locations: (1) Susquehanna River, (2) West Branch Octoraro Creek, (3) East Branch Octoraro Creek, (4) Buck Run, (5) West Branch Brandywine River, and (6) East Branch Brandywine River (Figure 8). The stratigraphic and structural base of each column is the Drumore tectonite that developed in the Pleasant Grove-Huntingdon Valley shear system (Valentino, 1990; Valentino and others, 1994). The Peters Creek Formation at the Susquehanna River is in thrust contact with the Sykesville Formation (Gates and others, 1991). Due to structural complexities in the southern exposures of the Peters Creek Formation, the top of the Octoraro Creek and Buck Run sections were chosen at the upward limit of simple structure to avoid

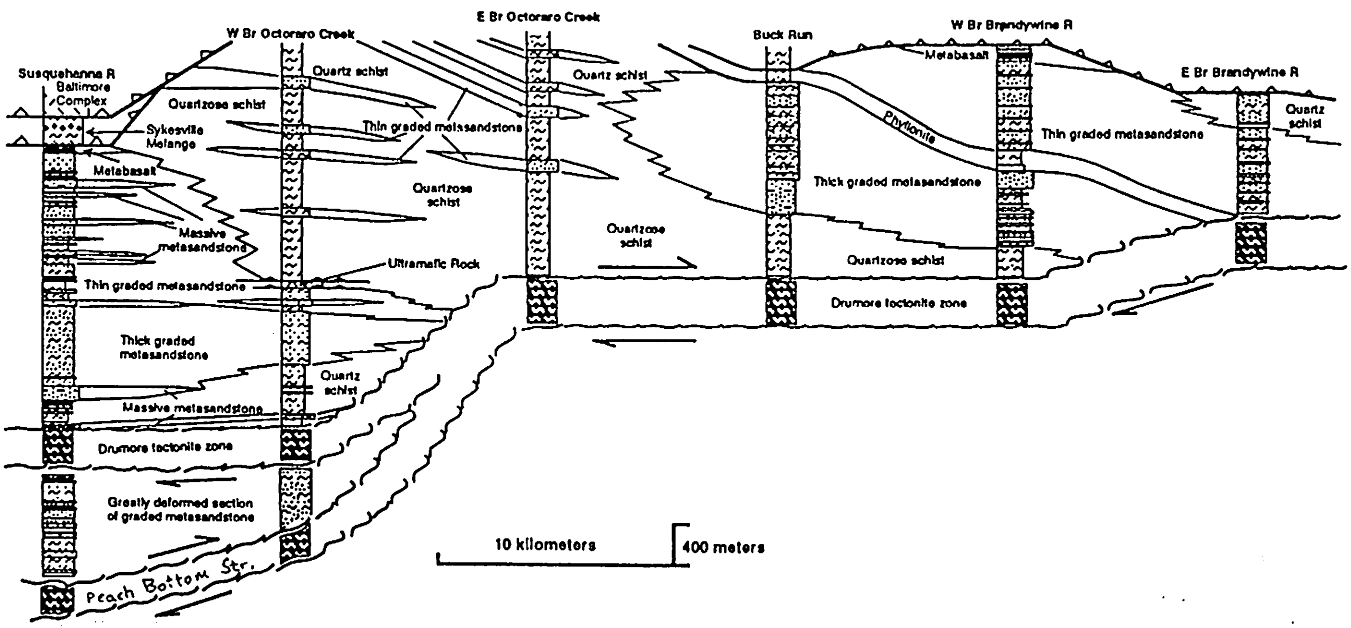


Figure 8. Lithostratigraphic correlation diagram for Peters Creek Formation, Lancaster and Chester Counties, Pennsylvania.

stratigraphic analysis in the complexly folded southern region. Full sections are probably much thicker than represented by the columns in Figure 8, but the structurally highest part is so multiply folded and so poorly exposed that they are not included in our stratigraphic analysis. The uppermost Brandywine River sections are truncated by the Cream Valley thrust zone (Wiswall, 1990).

Much of the Peters Creek Formation is composed of the graded metasandstone lithofacies that occurs in two sequences separated by a thick belt of quartzose schist: (1) a western sequence located between the Susquehanna River and the West Branch of Octoraro Creek, and (2) an eastern sequence located in the region of Buck Run and the branches of the Brandywine River. The western sequence has a lower chlorite-muscovite schist unit (1.5 km thick), overlain by a thick section of turbidites (3.2 km thick). The western sequence grades laterally into quartzose schist between the Susquehanna River and the East Branch of Octoraro Creek (Figures 6 and 8). The eastern sequence is composed of a lower quartzose schist unit (1.2 km thick) overlain by 4 km of interlayered feldspathic metasandstone and schist that also represent a graded metasandstone lithofacies. There are no contiguous units common to both the eastern and western Peters Creek sequences due to separation by the belt of quartzose schist. It is interesting that only the stratigraphically highest sections of the eastern and western sequence contain thin layers of greenstone. The two separate clastic sequences, each more than 3 km thick, separated laterally by quartzose schist lacking abundant metasandstone, suggests that the two regions dominated by graded metasandstone lithofacies represent separate turbidite-fan systems.

Modal mineralogy of the metasandstones was plotted on a QFL ternary diagram for comparison with sandstone from various tectonic settings (Dickinson and Suczek, 1979; Dickinson and others, 1983). Peters Creek metasandstone turbidite deposits plot in the transitional continental field (Figure 9). The ratio of potassium feldspar to plagioclase ranges from about 2:1 to 5:1. Data from the massive metasandstone lenses plot in the craton interior field. There are two types of lithic fragments: (1) polycrystalline aggregates of interlocking feldspar and quartz, and (2) rare fine-grained quartzite fragments. Grains containing interlocking feldspar and quartz are granite or granitic gneiss fragments. These lithic fragments coupled with the overall feldspar-rich composition of the metasandstone suggests a source region dominated by granitoid rocks, such as the Grenvillian basement. Granitic lithic fragments indicate unroofing of continental crustal material. The presence of greenstone interlayered with feldspathic metasandstone in the Peters Creek Formation suggests rift-related deposition, although other origins for the volcaniclastic rocks are possible.

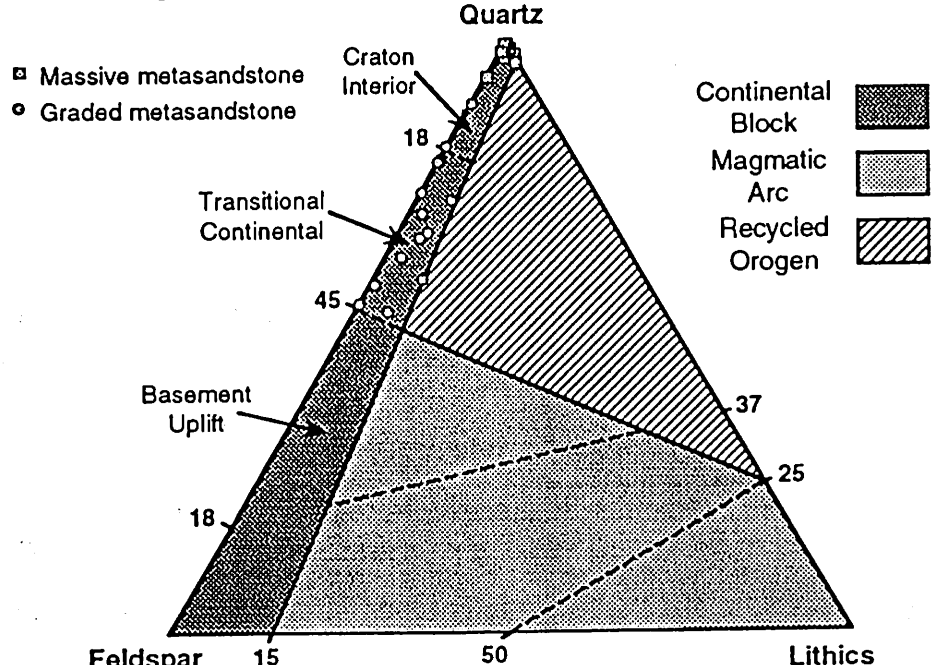


Figure 9. Quartz-feldspar-lithic fragment plot (after Dickinson and Suczek, 1979, and Dickinson and others, 1983) for the Peters Creek Formation metasandstone-bearing lithofacies.

The occurrence of greenstone and an exhumed granitic basement sedimentary source suggests extensional tectonics during the deposition of the Peters Creek Formation. The lack of lithic fragments derived from the Laurentian passive margin, such as carbonate detritus, suggests the Peters Creek Formation was deposited prior to the developement of the Cambrian passive margin, and most likely during late Proterozoic-early Cambrian Iapetan rifting (Gates and Valentino, 1991). The fine-grained quartzite fragments probably originated as intraformational sandstone clasts, but quartzite does occur in the Grenvillian massifs and that is another possible source.

INTERPRETATIONS AND REGIONAL RELATIONSHIPS

The physical elements of the Peters Creek Formation submarine turbidite-fan deposits are consistent with the type D submarine basin of Mutti and Normark (1987). Type D basins develop on tectonically active continental lithosphere and have complex basin dynamics, such as frequent changes in basin shape and sediment source. The spatial distribution of clastics in the Peters Creek Formation is complex, defining multiple depocenters (the eastern and western turbidite sequences), and evidence points toward a rift-related depositional setting. Such basins are filled with sediment dominated by gravity-flow deposits comprising coarse-grained channel and lobe sequences, and the time span of deposition is relatively short (10^4-10^5 yrs.).

The lateral distribution of lithofacies clearly defines two separate sequences of turbidite-dominated rocks (Figures 6 and 8), with an intervening region dominated by shale and siltstone. The lack of substantial topographic relief in the Piedmont and the lack of a basal depositional unconformity makes it nearly impossible to place constraints on the three-dimensional geometry of the turbidite-fan sequences. A down dip reconstructed cross section of the Peters Creek Formation provides a two-dimensional view of the sequence geometry (Figure 10A). The western turbidite sequence is wedge-shaped, with the thickest sections at the Susquehanna River thinning progressively eastward where the sequence is truncated at its base by a strike-slip fault. The eastern sequence of turbidite deposits has a lens-shaped geometry. These cross sectional geometries are consistent with turbidite-fan systems aligned subparallel or oblique to the depositional margin of the basin, with transport direction to the north or south relative to the present position (Figure 10B).

In the central and southern Appalachians, the metasediments of the Lynchburg Group form a rift sequence in the Blue Ridge anticlinorium (Wehr and Glover, 1985). In southwestern and central Virginia, the Lynchburg Group is dominated by submarine turbidite-fan deposits that grade along strike to the north into metasediments characterized by an overall fining upward succession of alluvial and fluvial to marine deposits (Wehr and Glover, 1985). The Lynchburg Group was flooded by rift basalts of the Catoctin Formation, extruded at approximately 570 Ma (Badger and Sinha, 1988). The lateral transition of Lynchburg Group lithofacies represent the transition in the environments of deposition across the depositional margin of the Lynchburg rift basin, and this relationship was interpreted to represent the ancient hinge zone (Wehr and Glover, 1985; Glover and others, 1992). The local hinge zone geometry across the Blue Ridge block was constructed slightly oblique to the structural grain of the orogen (Wehr and Glover, 1985). Shallow-water rift facies in the northern Blue Ridge suggests that the Lynchburg rift basin terminates toward the north, but the basin probably continues in the subsurface toward the east and northeast in northern Virginia and Maryland. Although the Peters Creek Formation contains lithofacies similar to the Blue Ridge rift sequence, it is structurally separate from the Blue Ridge sequence, particularly across a Paleozoic strike-slip fault (Valentino, 1993; Valentino and others, 1994); therefore any direct correlation is speculative at best at this time.

The Late Proterozoic-Early Cambrian stratigraphy in southern New England is generally defined by the arkose- and quartzite-rich sequence of the Lowerre Formation, which rests unconformably on Grenvillian basement of the Fordham gneiss, and the Lowerre Formation is overlain by the stable platform carbonates of the Inwood Marble (Hall, 1968; Prucha, Scotford, and Seider, 1968; Hall, 1979). Although the Lowerre Formation is similar in composition to that of the Peters Creek Formation, it lacks the complex distribution of turbidite deposits. Therefore, direct correlation of the Peters Creek and Lowerre Formations is unlikely.

An Iapetan rift-related transform fault was hypothesized to cross-cut structure of the Maryland-Pennsylvania Piedmont (Thomas, 1977; Fisher and others, 1979; Thomas, 1983) to link the southern rift basin in the Blue Ridge with a comparable rift basin in southern New England. The Peters Creek Formation might represent deposits associated with the proposed transform rift margin, but the Paleozoic structural history for the central Appalachians includes a component of dextral strike-slip faulting, suggesting the Peters Creek Formation originated 150 to 200 km east of its present location (Valentino, 1993; Valentino and others, 1994). Although there are no obvious correlations between the southern and northern rift sequences with that of the Peters Creek Formation at this time, the occurrence of rift deposits in the Pennsylvania Piedmont partially closes the gap between the southern and northern Upper Proterozoic Iapetan rift basins.

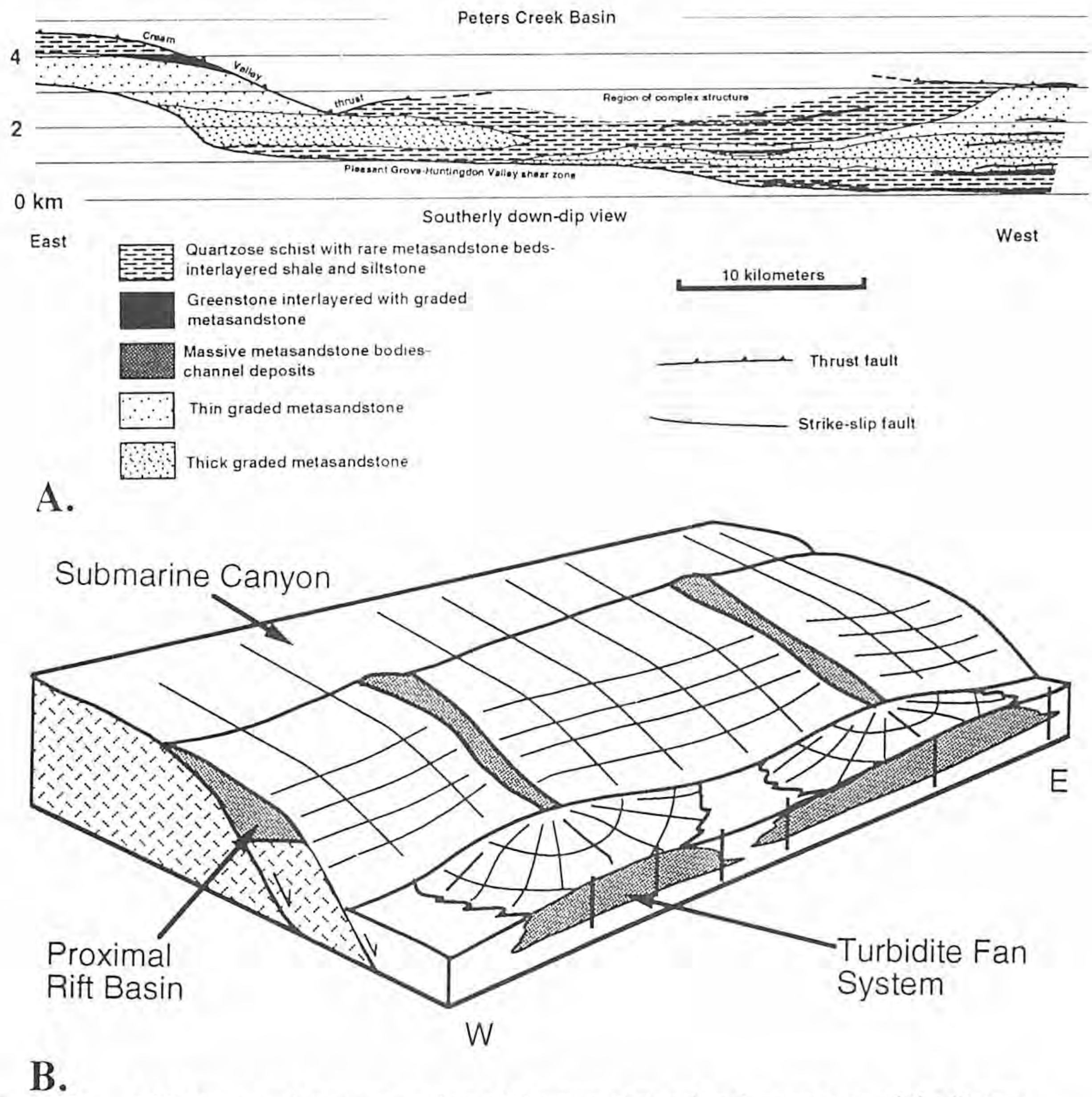


Figure 10. **[A]** Structural reconstruction of the two dimensional geometry of the Peters Creek Formation rift deposits. The view of this cross section is toward the south and into the earth. **[B]** Schematic depositional model for the Peters Creek Formation turbiditefan systems. The six vertical black lines represent the relative positions of the six stratigraphic columns of Figure 3.

GEOCHEMISTRY AND GEOLOGY OF METABASALT IN SOUTHEASTERN PENNSYLVANIA AND ADJACENT MARYLAND

Robert C. Smith, II, and John H. Barnes
Pennsylvania Geological Survey
P. O. Box 8453, Harrisburg, PA 17105-8453

ABSTRACT

Field observations and analyses of metabasalts in southeastern Pennsylvania suggest that widespread emplacement of latest Precambrian Catoctin Metabasalt in the Laurentian continent was independent of emplacement of a variety of Iapetan metabasalts, many of which appear to be associated with the Baltimore Mafic Complex (BMC).

The Catoctin Metabasalt is interpreted as within-plate initial-rifting continental tholeiite. Metabasalts having Catoctin affinity include the Catoctin Metabasalt *sensu stricto* of the South Mountain section of the Blue Ridge physiographic province and the Accomac-area metabasalts, including chemically associated metadiabase dikes in Grenville terranes other than Mine Ridge and the Brandywine massifs. However, some related suites, such as the Pigeon Hills-area metabasalt, exhibit marked chemical evolution toward oceanic basalts as rifting progressed into the drifting phase of ocean-floor generation. The "Holtwood Metabasalt" appears to represent a transition from the Accomac-area type to the Pigeon Hills-area type.

Basalts associated with the BMC include the Bald Friar Metabasalt from a back-arc spreading center, the well-known James Run island arc or back-arc volcanics, and various basalts of boninitic affinity from the forearc. Some of these are incorporated into ophiolitic mélanges in the so-called Peters Creek Formation that were later thrust-faulted and folded.

Presumed Iapetan ocean-floor basalts that are associated with the drifting phase of ocean-floor generation but that are not directly associated with the BMC include the Kennett Square Amphibolites from Chadds Ford to West Grove, Chester County, and the type Sams Creek Metabasalt from Maryland, both of which lack any inherited geochemical memory of Catoctin rifting.

INTRODUCTION

Metabasalts, although volumetrically almost insignificant in the Piedmont province (Figure 11), were, on the whole, well mapped by Florence Bascom, A. I. Jonas, G. W. Stose, and others in the 1920's and 1930's. Perhaps this was, in part, due to the thrill of finding something other than muscovite-chlorite-quartz schist. Beginning with the availability of reliable trace element analyses in the 1960's, some geochemists began correlating basalt chemistry with known tectonic environments. Systematic variations related to mantle inhomogeneity were noted and empirical diagrams relating chemistry to emplacement environment in plate-tectonic terminology were prepared (Pearce and Cann, 1973). Attempts to apply the diagrams to metabasalts from poorly known tectonic environments met with variable success. A fair degree of success was achieved by those who incorporated information on the field relations, utilized the relatively immobile trace elements, maintained sampling and analytical quality control, and who actually read the initial papers describing the geochemical techniques and utilized the diagrams in the recommended sequence and manner.

Further refinement was achieved by scientists such as Poul E. Holm (1982) who, in his study of the non-recognition of continental tholeiites using the Ti-Y-Zr diagram, noted that the York Haven and Rossville Diabases of Pennsylvania, among others, appeared to be incorrectly classified. Not content to be merely a critic, Holm went on (1985) to create a diagram that could be used to compare basaltic-rock data that were normalized to primordial mantle (Figure 12D). He proposed that, through the use of this diagram, continental initial-rifting basalts (IRT) and many other basalt types could be correctly recognized. Thus, the eastern North America early Mesozoic diabase trace-element enigma (apparent continental basalt that showed calc-alkaline tendencies) was largely solved by 1985. Despite this and Holm's (1985, p. 305) noting that "Truly continental basalts, of course, should be

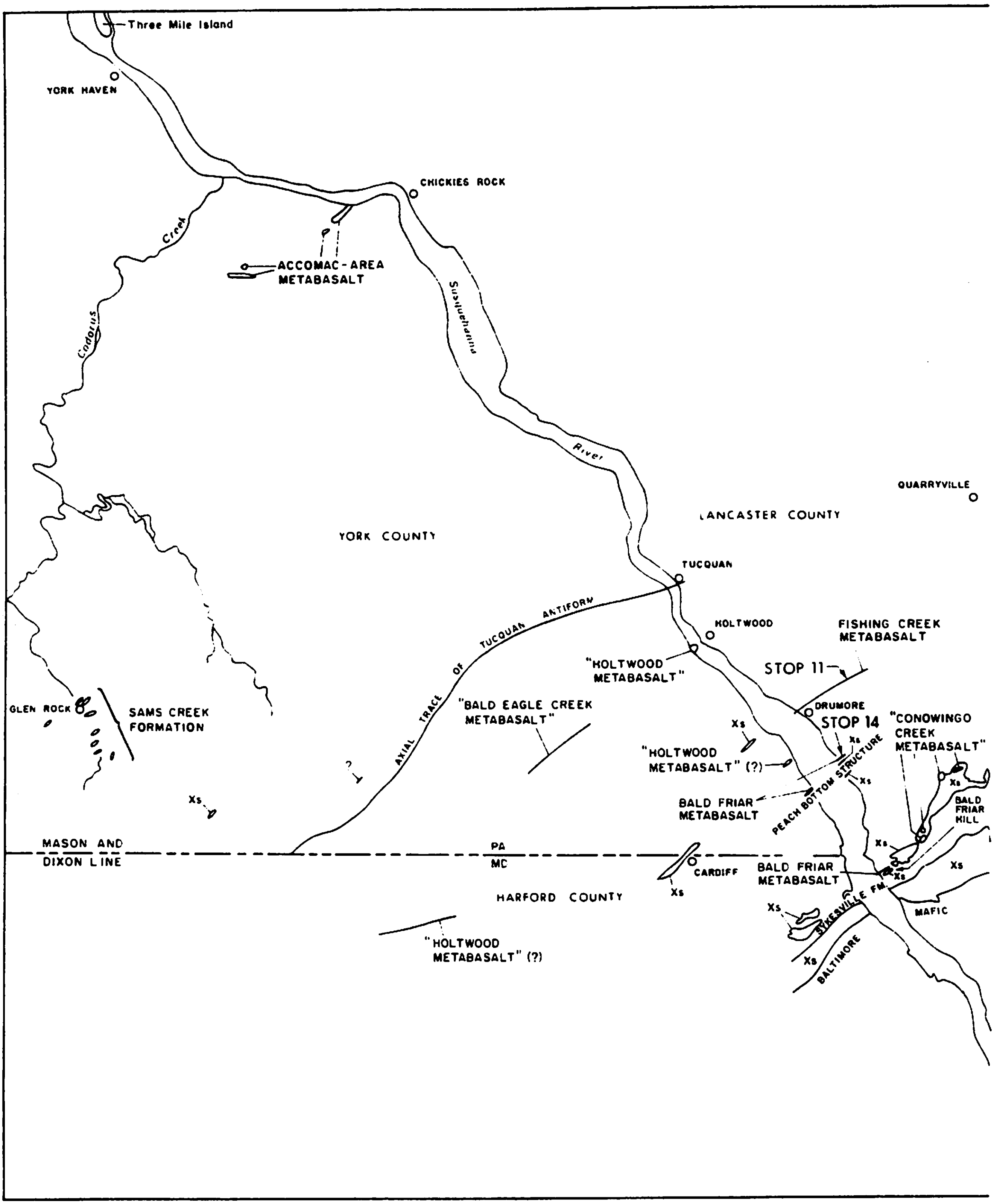


Figure 11. Location of selected metabasalt and amphibolite populations on a base map modified from Berg and others (1980), Cleaves and others (1968), and Higgins and Conant (1986).

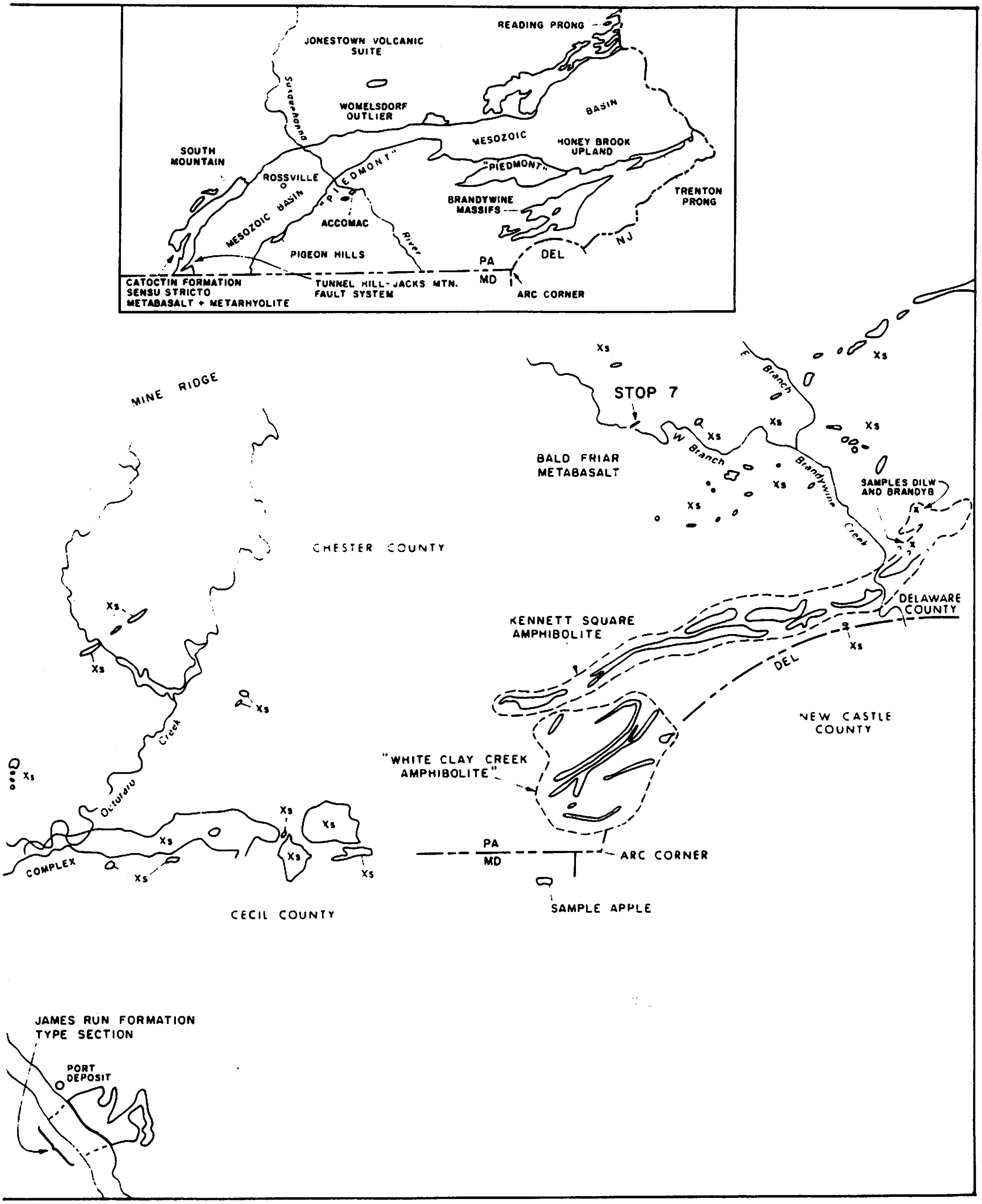


Figure 11. Continued.

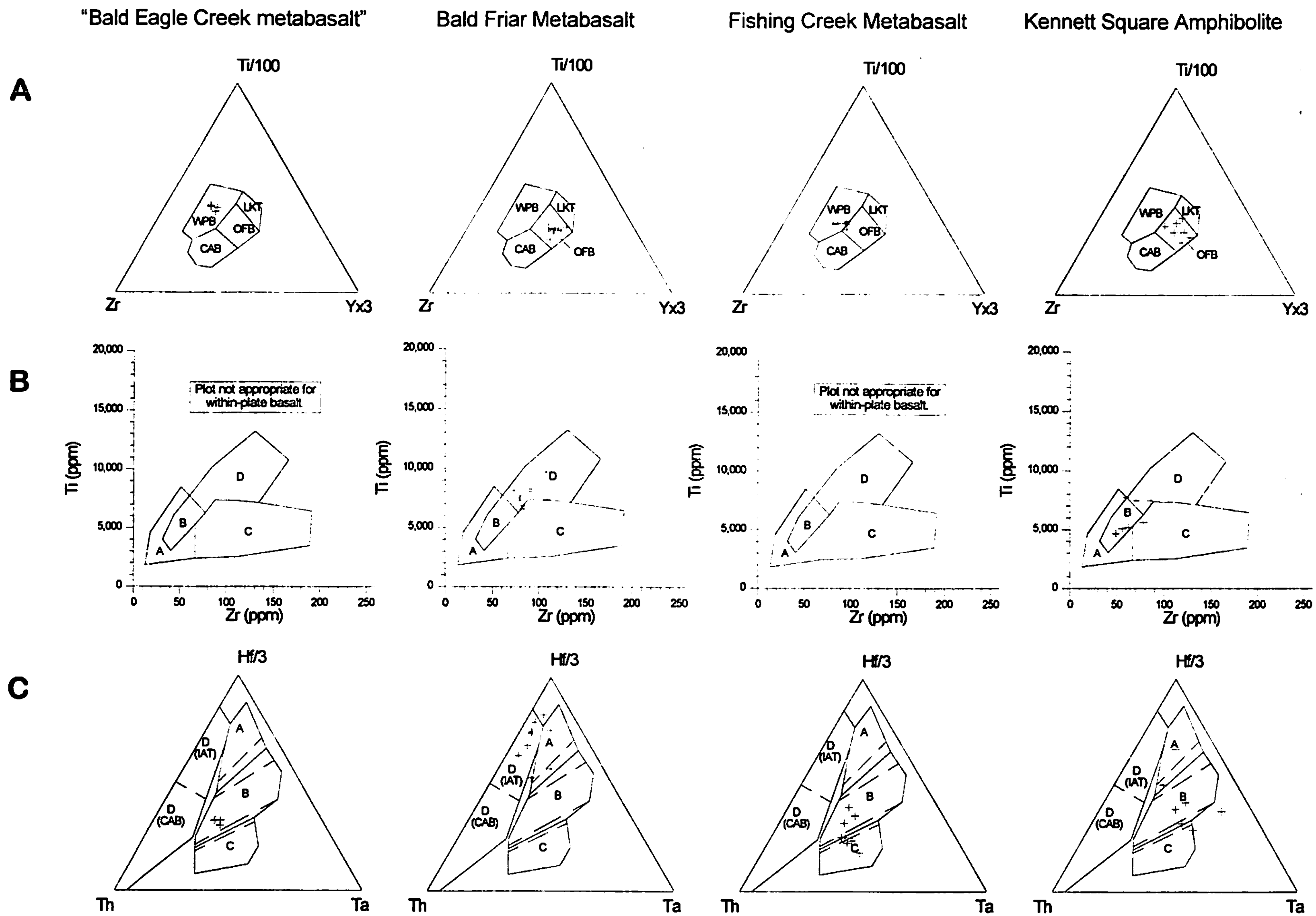


Figure 12. (A) Plots of Ti, Zr, and Y (Pearce and Cann, 1973). WPB, within-plate basalt; LKT, low-K tholeiite; CAB, calc-alkaline basalt; OFB, ocean-floor basalt. (B) Plots of Ti vs ZR (Pearce and Cann, 1973). A, low-K tholeiite; B, low-K tholeiite, ocean-floor basalt, and calc-alkaline basalt; C, calc-alkaline basalt; D, ocean-floor basalt. (C) Plots of Hf, Th, and Ta (Wood, 1980; Wood and others, 1979). A, normal mid-ocean ridge basalt (N-MORB); B, enriched mid-ocean basalt (E-MORB) and tholeiitic within-plate basalt; C, alkaline within-plate basalt; D, destructive plate-margin basalts. As used herein, the term "MORB" can also apply to spreading centers that are not located at mid-ocean ridges.

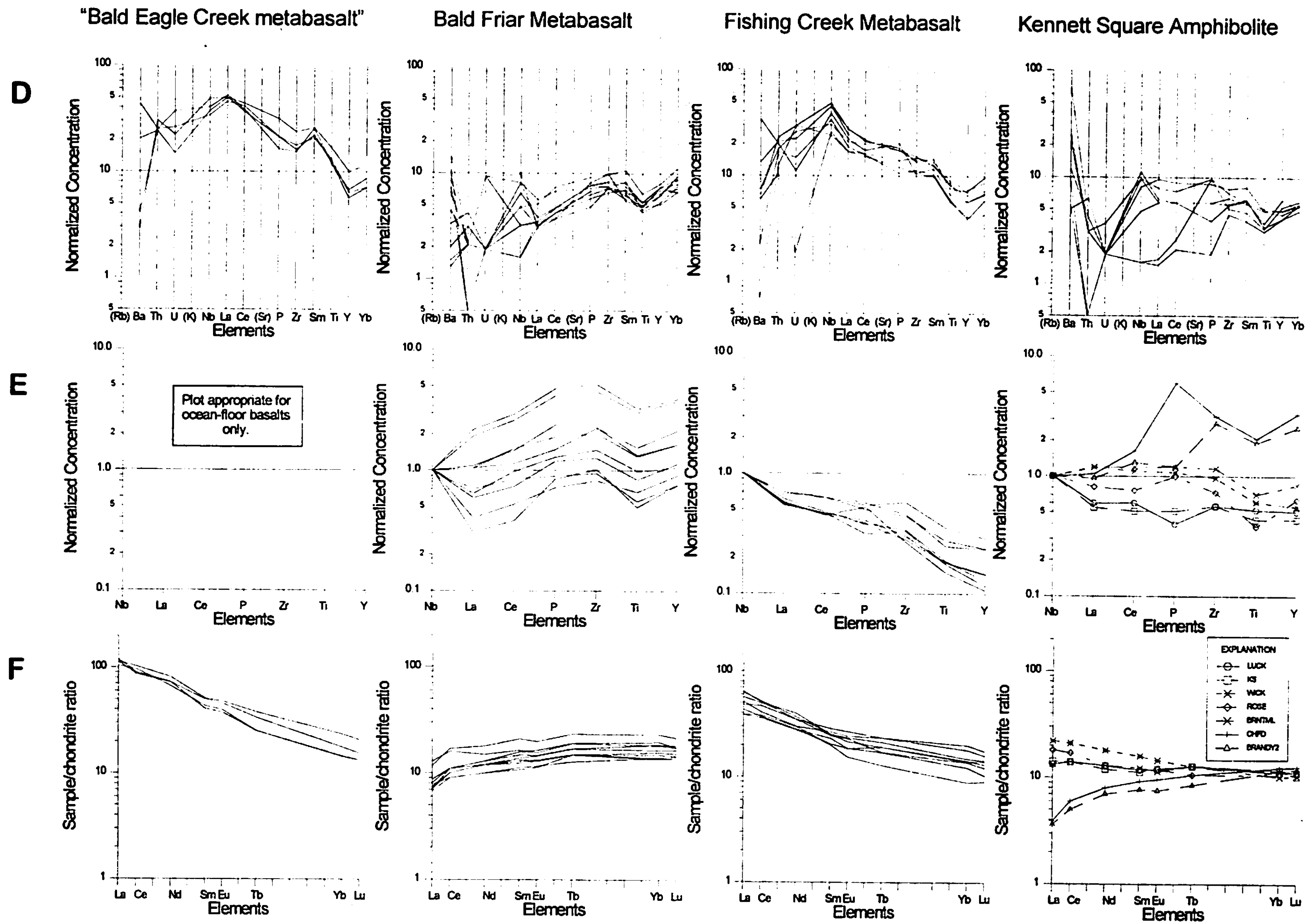


Figure 12. (D) Plots of selected elements normalized to primordial mantle (Holm, 1985). (E) Spidergram plots (Myers and Breitkopf, 1989; Thompson and others, 1983). Data for each element are plotted as the ratio of the element to Nb, normalized to the ratio of the element in primordial mantle to Nb in primordial mantle. (F) Plots of chondrite-normalized rare-earth elements. Chondrite values from Anders and Ebihara (1982).

unrelated to any initiating plate boundary, otherwise the designation *within-plate continental* has no meaning," some misuse of discriminant diagrams has continued. Julian S. Marsh (1987) echoed Holm's comments on continental basalts and emphasized the variety of tectonic environments that some geologists have lumped under "continental."

The misuse of discriminant diagrams has often revolved around the misconception that tectonomagmatic environments could be neatly categorized and do not evolve through time. Russell E. Myers and Jörg H. Breitkopf (1989) were among the first to suggest the need for dynamic rather than static reasoning:

> "The realization that tectonic environments are not static but evolve continuously is probably the key to understanding the relationships between tectonics and basalt geochemistry. The problem with working in an evolving tectonic environment is that just as the tectonic processes are transitional so are the basalt compositions. For this reason we believe that there is a fundamental philosophical error in attempts to develop *discrete* tectonomagmatic chemical classifications. This is probably the reason why tectonomagmatic classification diagrams so often fail to produce con-clusive results. It is our contention that by means of careful stratigraphic geochemical studies it may be possible to document temporal and spatial variations in basalt compositions which may then be related to the evolutionary patterns of specific tectonomagmatic cycles." (Myers and Breitkopf, 1989, p. 53-54)

Myers and Breitkopf's Nb-normalized diagrams (Figure 12E) seem to work well for normal ocean floor basalts (N-OFB) (see the explanation of abbreviations and selected terms, Table 1) through transitional (T-OFB) to plume=enriched (P=E-OFB) ones. Indeed, within two of the ocean-floor-basalt populations to be discussed below, the Sams Creek Metabasalt and Kennett Square Amphibolites, there appear to be correlations of these types with geographic direction.

Distinction between P=E-OFB, a modern example being the 45°N Mid-Atlantic Ridge, and within-plate ocean-island tholeiites (OIT), such as Hawaii, was difficult during the present study. Without other types of data, only the Y-Yb slope of the primordial-mantle-normalized diagrams of Holm (1985) seemed to help. Based on the observation by Sun (1980) that P-OFB and OIT have similar isotopic trends, it may be that the geochemical character of oceanic mantle plumes or hotspots overwhelms differences that are dependent on whether the volcanism is located in a plate interior or plate margin.

When they began their study of the Catoctin Metabasalt, Smith and others (1991) were largely occupied with primary igneous features, facing directions, and map corrections. However, they eventually began to think of the Catoctin event as a dynamic process in terms of progressive dilation of the crustal cover. This put them not far behind their sedimentological co-workers studying the rift-to-drift facies of the Chilhowee Group. Further, they briefly addressed the trace-element depletion in the nearby early Mesozoic diabases in Pennsylvania and suggested a connection between this depletion and the Catoctin event, 400 Ma prior, as discussed below.

A problem that remains is that many of us received our formal training prior to the general acceptance of plate-tectonic theory and have not yet adjusted our thinking to the fact that it is the mantle, via its radiogenic heat engine, that controls crustal processes and not vice versa. *Mantle diapirs control plate tectonics.* Basaltic chemistry, especially in areas of thick continental crust, frequently anticipates crustal processes such as sedimentation. Not unlike the plans of mice and men, tectonomagmatic processes, such as continental rifting, frequently are uncompleted. In such cases, the basalt chemistry may truly reflect the mantle situation at a given time, but be preserved with incongruous, laggard sediments.

"Although it has been popular in recent years to bash the use of such diagrams, we find that their discriminant (pun intended) use can be very powerful. At the minimum, they tend to group similar basaltic rocks. At best, they provide one tool to interpret tectonomagmatic environments" (Smith and others, 1991, p. 10). This is not to imply that all hurdles have been leapt. A plethora of basalt type abbreviations remain to fend off all but the most intrepid. The use of the explanation of abbreviations in Table 1 and chemical symbols in Table 2 may assist the reader over these hurdles.

Table 1. Explanation of abbreviations and selected terms used in this report.

BABB: Back-arc basin basalt. Presumably formed at a spreading center in the back-arc region of a destructive plate margin and, except for volatiles, may be indistinguishable from MORB.

E–OFB = P–OFB ≅ E–MORB = P–MORB: Enriched or plume ocean-floor basalts. As used herein, does not assume that all spreading centers are located at mid-ocean ridges. Our experience suggests that spreading centers are not readily divided on the basis of abundances of nonvolatile elements. Therefore, we prefer the more inclusive "OFB."

HREE: Heavy chondrite-normalized rare-earth elements. Generally Er, Tm, Yb, and Lu.

IAT: Island-arc tholeiite. Presumably associated with destructive plate margins. As used by some, would include the series that evolve from LKT to calc-alkaline to alkaline to shoshonite.

IREE: Intermediate chondrite-normalized rare-earth elements. Generally Sm, Gd, Tb, Dy, and Ho.

IRT: Continental initial-rifting tholeiites. In our experience, may be transitional from continental to ocean-floor (OFB). Also a part of the New York subway system.

LKT: Low-K (potassium) tholeiite. Presumably an early stage of island-arc development associated with a destructive plate margin.

LREE: Light chondrite-normalized rare-earth elements. Generally La, Ce, Nd, and Pr.

N–OFB ≅ N–MORB: Normal ocean-floor basalt. The former abbreviation does not, unlike the latter, assume that all spreading centers are located at mid-ocean ridges.

OFB: Ocean-floor basalt. Does not assume that all spreading centers are located at mid-ocean ridges. Back-arc and possibly other environments could yield similar basalts.

OFT: Ocean-floor tholeiite. This term appears to be intended as an equivalent to N–MORB, but without the implied unique genesis at a mid-ocean ridge spreading center.

OIT = OIB: Ocean-island tholeiite = ocean-island basalt. Presumably formed within oceanic plates.

OTL: Out-to-lunch basalts. Presumably includes those subjected to intense hydrothermal alteration.

"Steerhorn": A chondrite-normalized rare-earth plot in which the intermediate-atomic-number rare-earth elements are depleted relative to light and heavy rare-earths. For volcanic rocks, generally considered to be diagnostic of boninites.

T–MORB ≅ T–OFB: Transitional mid-ocean ridge basalt = transitional ocean-floor basalt. The transition presumably is from N (normal) to P=E (plume = enriched) basalt.

VAB: Volcanic-arc basalts associated with a destructive plate margin. Presumably includes a range of K_2O contents.

WPB: Within-plate basalts. From either continental or oceanic plates. "WP" is used as an adjective to refer to the same environment.

METABASALT POPULATIONS AND INTERPRETATIONS

The intent of the research reported here is to present data on numerous suites of basaltic rocks in southeastern Pennsylvania and to present interpretations of affinities, correlations, and tectonic environments that may be helpful in deciphering the geologic history and relations in that region. The suites are discussed in alphabetical order.

Analytical accuracy may be estimated from the data in Table 3 for USGS standard reference sample BCR-1, which was submitted to the laboratory as a blind unknown. Precision of sampling and analyses may be estimated from the analyses in Table 3 of samples FSHCKFG, FSHCKFGII, and FSHCKFGIII, which are from a magnetite-bearing metabasalt block that was cut perpendicular to foliation. Each slab was prepared separately and analyzed in a separate batch in order to obtain an estimate of sampling error plus analytical error. Except for Ni and V, which were

Table 2. Explanation of geochemical significance of elements listed in Tables 3, 5, and 6.

Symbol	Substance	Significance
TiO_2	Titania	Titania is the oxide of a first-row (of the periodic table) transition metal, titanium. In reduced rocks titanium tends to be incorporated in silicates; in rocks having an intermediate oxidation state it tends to be associated with iron oxides; and in rocks having a high oxidation state it occurs as a straight oxide. Titanium is typically enriched in alkali basalts but is depleted in within-plate continental basalts and especially in boninites.
Zr Hf	Zirconium Hafnium	Zirconium, a second-row transition metal, and its lanthanide-contraction-related third-row sibling hafnium tend to be concentrated in the extremely stable silicate mineral zircon. Zircon is so resistant to weathering and abrasion that crystals survive multiple geochemical cycles, the cores retaining U–Th–Pb ages from previous cycles. Hafnium can be relatively abundant in normal ocean-floor and island-arc basalts.
Nb Ta	Niobium Tantalum	Niobium, a second-row transition metal, and its lanthanide-contraction-related third-row sibling tantalum tend to be concentrated in the last phases to crystallize out of a magma or the first to melt. Both elements are strongly depleted in island-arc basalts and somewhat depleted in continental tholeiites. However, they are moderately enriched in initial-rifting continental tholeiites and in plume, relative to normal, ocean-floor basalts. They are enriched in ocean-island within-plate basalts, especially if the basalts are alkali.
Th	Thorium	Thorium, a radioactive actinide-series element, tends to be concentrated in continental rocks and calc-alkaline basalts relative to normal oceanic basalts.
U	Uranium	Uranium, a radioactive actinide-series element, is a bit too mobile during weathering and metamorphism to be interpreted alone. It tends to be very low in abundance in most normal oceanic basalts but more abundant in alkali basalts.
Ni	Nickel	Nickel is the first-row transition metal that comprises 25 percent of a U.S. 5-cent copper coin. Like uranium, it is too mobile to be reliable itself, but it tends to be a good measure of how mafic a rock is. Normally, the more mafic a rock, the higher the nickel content. However, in slowly cooled, sulfur-rich magmas, Fe–Ni sulfides may separate from the melt. Scandium (Sc) can sometimes also be used to measure how mafic a rock is because of its substitution for magnesium (Mg) in silicates such as olivine and pyroxene. Nickel and chromium (Cr) tend to be of high abundance in oceanic basalts and especially in boninites.
V	Vanadium	Vanadium, a first-row transition metal, has different natural oxidation states and tends to be concentrated in different minerals, depending on the oxidation state of the magma. Shervais (1982) organized vanadium data against titanium and showed that alkali basalts tended to be reduced; island-arc basalts, including boninites, tended to be oxidized; and back-arc basalts tend to be low in vanadium but have variable amounts of titanium. Boninites are reported to typically be low in vanadium.
Y	Yttrium	Yttrium, a second-row transition metal, is a rare-earth element that typically behaves like a heavy lanthanide element because of the similar ionic size. Yttrium tends to be of somewhat low abundance in within-plate oceanic or continental basalts and island-arc basalts and of extremely low abundance in boninites.
La Ce	Lanthanum Cerium	Lanthanum and cerium are the two lightest of the lanthanide elements. Lanthanum is typically depleted relative to cerium in normal ocean-floor basalts. Both are highly enriched in alkali basalts.

determined by inductively coupled plasma (ICP) analysis, the results suggest that precision is typically within 10 percent.

To better understand the discussions that follow, the attention of the reader is directed to Figure 11, a map showing the general locations of the 16 populations; Table 4, interpretations of plots of analytical data on discriminate diagrams and field observations; and Table 5, raw analytical data and latitude and longitude of each sample.

Table 3. Tests of accuracy of analytical data. As a test of accuracy, USGS standard reference sample BCR-1 was submitted as a blind unknown. The analytical data obtained are compared with values reported in a compilation of "usable" values by Abbey (1983) and values obtained by Smith (1973). As a test of precision, three slabs of metabasalt sample FSHCKFG from Lancaster County were submitted separately for analysis. All analyses are in parts per million except TiO_2 which is in percent.

	TiO_2	Zr	Hf	Nb	Ta	Th	·U	Ni	V	Y	La	Ce	Lat. N	Long. W
U.S.G.S. STANDARD REFERENCE SAMPLE BCR–1														
This study	2.32	194	4.1	11	0.6	5.2	1.3	12	449	36	23.6	49	~45°34'33"	~122°09'14"
Abbey (1983)	2.26	185	5	19?	0.8?	6.1	1.7	10	420	40	27	53		
Smith (1973)	2.22	145	4.1	20.4	0.79	5.9	1.5	7	420	29	20			
FISHING CREEK METABASALT														
FSHCKFG	2.24	150	3.6	30	1.4	2.2	0.8	50	240	34	19.5	42	39°47'56"	76°15'23
FSHCKFGII	2.09	150	3.3	28	1.6	2.2	0.6	90	310	32	17.5	40	39°47'56"	76°15'23
FSHCKFGIII	2.15	168	3.4	28	1.6	2.0	0.6	63	230	28	19.7	41	39°47'56"	76°15'23

Accomac-area Metabasalt

Accomac is in York County on the southwest side of the Susquehanna River across from Chickies Rock. The suite from Accomac consists of 5 samples from one long roadcut that contains cryptic high-angle faults, some of which contain trace amounts of Cu mineralization. As mapped by Stose and Jonas (1939), the metabasalt is associated with metarhyolite and occurs beneath the Lower Cambrian Chickies Formation of the Chilhowee Group. Based on this association, it has a *presumed* age of latest Precambrian. Some of the metabasalt is highly amygdaloidal and contains pillows, especially near the creek.

The Accomac-area suite has strong chemical affinities with the Catoctin Metabasalt *sensu stricto* of South Mountain in Adams and Franklin Counties. The range of rock compositions at Accomac may perhaps best be considered to be a vignette of the rifting phase of the Catoctin event. Plots of TiO_2-Nb (Figure 13), Th-Ta, and Hf-Y suggest one major pulse of magma which differentiated with time. According to most of the diagrams utilized, the basalt should be classified as within-plate. The exception is the diagram of primordial-mantle-normalized elements, by which one can recognize the aspect of continental initial rifting. A Catoctin "basement" should be considered in this area. Catoctin metadiabase dikes in the Grenville terranes, discussed elsewhere in this report, seem to have formed from the same magma as the Accomac-area metabasalt!

"Bald Eagle Creek Metabasalt"

This informally defined suite consists of only 4 samples of garnet-grade metabasalt from a limited area in York County SW of Holtwood Dam. Definitive primary igneous textures were not observed and only their parallelism to regional strike suggests that they are extrusive. All samples are similar, somewhat alkali within-plate basalts or alkali initial-rifting tholeiites (Figure 12). Their regional significance is uncertain, but a relationship to Catoctin rifting seems plausible.

Because of the rather distinctive alkali affinity, as suggested by the high TiO_2 content (Table 5), correlation of these rocks to possible metabasalt in the NW limb of the Tucquan anticline (Figure 11) may be possible. Rock from the NW limb of the Tucquan anticline that would be a suitable candidate for testing is the sparse, weathered, dense, garnet-bearing metabasalt float observed at 39^{o} 45' 32" N, 76^{o} 33' 07" W.

Table 4. Summary of observations and interpretations resulting from plots of analytical data on geochemical diagrams and summary of field observations.

POPULATION	Ti–Zr–Y (Pearce and Cann, 1973)	Ti–Zr (Pearce and Cann, 1973)	Hf–Th–Ta (Wood, 1980; Dawson and Jacobson, 1989)	Primordial-mantle-normalized hygromagmatophile elements (Holm, 1985)	Nb-normalized spidergram (Myers and Breitkopf, 1989)	Chondrite-normalized rare-earth elements	Comments[1]	Conclusions
Accomac-area metabasalt	All in WPB field.	Diagram not appropriate for WPB.	All cluster in tholeiitic WPB plus E=P–OFB field.	Nb slightly (+); Sm to Y slope (–). Continental initial-rifting tholeiite.	Diagram appropriate for OFB only.	All slightly to strongly LREE (+).	Pillows in section. Rhyolite nearby. Ti–Nb plot suggests a single differentiation series.	Continental initial-rifting tholeiite. A vignette of the Catoctin Metabasalt.
"Bald Eagle Creek metabasalt"	All in WPB field.	Diagram not appropriate for WPB.	All cluster in tholeiitic WPB plus E=P–OFB field.	Sm to Y slope (–). Continental initial-rifting tholeiite.	Diagram appropriate for OFB only.	All strongly LREE (+), sample:chondrite >100.	At garnet-grade. Possibly on strike with Holtwood (HLTW) series. Ti/V of Shervais (1982) suggests alkali.	Very uniform alkali, continental initial-rifting tholeiite. Might be part of the Catoctin event.
Bald Friar Metabasalt	All cluster in OFB field.	Seven in OFB field and 3 on OFB–calc-alkaline boundary.	Three N-OFB. Six are below detection limit for Ta.	All (+) slope; therefore OFT=N–OFB. Uniform La to Yb. Nine Nb (+). Ti (–) in all.	All (+) slope; therefore N–OFB.	All LREE (–) and very similar; therefore N–OFB.	Pillows preserved at 2 locations. Ultramafic rocks within 100 m at 2 locations. Maryland and STOP 14 samples in possible melange.	Very uniform N–OFB from both sides of Peach Bottom structure. Probably from a back-arc spreading center. Probably related to BMC.
Catoctin metadiabase dikes in Grenville terrains	All cluster in WPB field.	Diagram not appropriate for WPB.	All cluster in or near tholeiitic WPB plus E–OFB.	Th (–), Nb (+), Sm to Y slope (–): Continental initial-rifting tholeiite.	Not appropriate for WPB. but a consistent group.	All LREE (+) and linear.	Well-preserved chilled margins. Occur only in Grenville terranes.	Part of the extremely widespread Catoctin event. Very similar to Accomac-area metabasalt.
Catoctin metabasalt *sensu stricto*	On WPB–OFB boundary, more samples in former.	Diagram not appropriate for WPB.	Most samples in field for tholeiitic WPB plus E=P–OFB.	Open ∩ shape, Nb slightly (+), Sm to Y slope (–). Continental initial-rifting tholeiite.	Diagram appropriate for OFB only.	Linear. LREE enrichment increases with Ti. KNSC, a possible xenolith, "steerhorn."	Associated with slightly younger rhyolite. Textures preserved locally.	Continental initial-rifting tholeiite, i.e., the igneous equivalent of rift to drift. Multiple pulses, the later ones more alkali.

Table 4. Summary of observations, interpretations, and field observations (continued).

POPULATION	Ti–Zr–Y (Pearce and Cann, 1973)	Ti–Zr (Pearce and Cann, 1973)	Hf–Th–Ta (Wood, 1980; Dawson and Jacobson, 1989)	Primordial-mantle-normalized hygromagmatophile elements (Holm, 1985)	Nb-normalized spidergram (Myers and Breitkopf, 1989)	Chondrite-normalized rare-earth elements	Comments[1]	Conclusions
"Conowingo Creek" metabasalt	"The extremely low contents of Ti, Zr, and Y [among others] of boninites invalidates the use of *standard* geochemical diagrams . . ." (Coish, 1993, p. 9).			Open ∩ shape with scatter. Th and Y (–). Y to Yb slope (+).	Diagram not appropriate, but shows extreme Y (–) for 3 samples.	CONJSE and CONJSEII IREE (–) (equals "steerhorn"). These plus 3 show Sm (–).	Occurs within BMC. CONJSEIII suffers pillow-rim enrichment. APPLE included for convenience.	Boninitic affinity based on "steerhorn" REE, high Mg, Cr, and Ni, and low Ti, Zr, and Y.
Early Mesozoic diabase	Not WPB.	"Calc-alkaline" and LKT.	"Calc-alkaline."	K and Nb (+). Slope, Ti, and Sr (–). Therefore, continental to initial-rifting tholeiites.	Diagram appropriate for OFB only.	All LREE (+), but QUARRY and possibly ROSS are "steerhorn," boninitic affinity.	Occurs with older continental sediments. Smith (1991, p. 16) suggested derivation from depleted mantle. Ti / V for QUARRY is boninitic.	Continental to initial-rifting tholeiites, but from depleted mantle.
Fishing Creek Metabasalt	All in WPB field or nearby in OFB.	Diagram not appropriate for WPB.	In tholeiitic WPB plus E=P–OFB field and some into alkaline WPB.	Plot is ∩ shaped. Lack (+) U and Th; Nb (+); but anomalous Y to Yb (+) slope.	Diagram not appropriate for WPB, but high Nb suggests plume influence.	All LREE (+) and very similar.	Too sheared to determine textures at most outcrops, but no offset across proposed Drumore tectonite zone. Ultramafic along strike to the southwest.	Zr–Nb–Y suggests WP. Ti–V suggests OIT variety of WP. Probably WP, somewhat alkali, possible plume influence.
"Holtwood metabasalt"	Pa. HLTW samples WPB. COYLK and HLTWMD OFB. HLTWBASE on boundary	COYLK, HLTWMD, and HLTWBASE OFB.	Four samples below Ta detection limit.	Pa. samples rather flat, Nb (+), Th and Sm to Y slope (–); therefore initial-rifting tholeiite. HLTWMD E=P–OFB.	HLTWMD P=E– or transitional OFB. Diagram not appropriate for others.	Pa. samples linear, very similar, and slight LREE (+). HLTWMD L and IREE (–).	Pennsylvania HLTW samples from single location; no proof that they are extrusive. COYLK and HLTW included for convenience.	Pennsylvania samples might be continental initial-rifting tholeiite. Possibly related to Pigeon Hills, but earlier, and to Accomac, but later.
James Run Formation	All various active margin.	OTL	Four calc-alkaline and one IAT-destructive plate margin.	Nb (–) for all but JRS and JRFMC. Ti (–) for all. Therefore, LKT from island arc or continental margin.	Diagram appropriate for OFB only.	Four are LREE (+) including JRS, which is slightly "steerhorn." JRFMC LREE (–)	Not OFB based on low Cr. Large Ti–V spread fits back-arc. Nb–Zr–Y of Meschede (1986) fits VAB or N–OFB.	Island arc above a subduction zone near a continental margin. The "steerhorn" LREE in dike JRS suggests boninitic affinity.

Table 4. Summary of observations, interpretations, and field observations (continued).

POPULATION	Ti–Zr–Y (Pearce and Cann, 1973)	Ti–Zr (Pearce and Cann, 1973)	Hf–Th–Ta (Wood, 1980; Dawson and Jacobson, 1989)	Primordial-mantle-normalized hygromagmatophile elements (Holm, 1985)	Nb-normalized spidergram (Myers and Breitkopf, 1989)	Chondrite-normalized rare-earth elements	Comments[1]	Conclusions
Jonestown Volcanic Suite	OFB.	Two OFB, 1 indeterminate.	Two E–OFB plus WPB, 1 calc-alkaline.	Nb (+), Sm to Y slope (–). Possibly E–OFB or IRT?	PA72 T–OFB. Other 2 P=E–OFB.	Slightly LREE (+).	Chromite octahedra in olivine. Abundant pillows. Andesite reported in area.	Based on very limited data, probably P=E–OFB.
Kennett Square Amphibolite	Scattered in OFB field.	Three eastern samples OFB, middle calc-alkaline, others indeterminate.	Easternmost N–OFB. The rest plot in and around E=P–OFB[2].	Two eastern (+) slope, therefore OFT=N–OFB. Others slightly (–) with Nb (+); 4 of these Y to Yb (+), therefore E=P–OFB.	Two eastern (+), N–OFB. Two middle T–OFB. Western P=E–OFB.	Three eastern samples LREE (–), therefore N–OFB. Others approximately flat, could be transitional or P=E–OFB.	Amphibolite grade except for WICK. Zr increases from east to west.	OFB from N-type in east to transitional to E=P in west. Spreading center was probably oriented north-south.
"Older diabase" dikes of Bascom and Stose (1932).	BALTP, BATLTP2, and BEAU2 are WPB. The rest are OFB or CAB.	ARMK2, BEAU, CPSUN, RADNOR, and possibly CRMCK plot as "calc-alkaline."	Low Ta in ARMK2, BEAU, CPSUN, and CRUMCK (as well as low Nb in RADNOR) suggest subduction or N–OFB.	High-Ti samples plot as open $\cap$ shape, Nb (+), Sm to Y slope (–), therefore initial-rifting tholeiite. Low-Ti samples depleted and flat, therefore T–OFB?.	Diagram not appropriate for BALTP, BALTP2, and BEAU2. Rest are P=E– or T–OFB.	High-Ti BALTP series plus BEAU2 LREE (+). Low-Ti BEAU and CRUMCK L- and IREE (–). RADNOR and CPSUN are IREE (–).	Narrow, planar-contact dikes, having well-developed chilled margins, which intruded the Brandywine massifs. Nb–Zr–Y fit N–OFB or VAB for lower-Ti. Cr–Y of Pearce (1982) fit OFB for most.	Transition from high-Ti continental initial-rifting tholeiite to depleted OFB? Boninitic (?) affinity for low-Ti group.
Pigeon Hills-area Metabasalt	PIGHL2, 3, 4, and 5 are OFB. PIGHL and PIGHL888 plot as WPB but Y may be mobile in these 2 highly altered samples.	All but PIGHL plot as OFB.	PIGHL, and PIGHL3, 4, and 888 plot as OFB[2]. Ta below detection limit in PIGHL2 and 5.	PIGHL4 and 5 slope (+), therefore OFT= N–OFB. In others Nb (+), Th and Sm to Y slope (–), therefore continental initial-rifting tholeiite.	PIGHL4 and 5 are N–OFB. Other four are transitional or P=E–OFB.	Essentially flat with La<Ce for PIGHL, PIGHL4, and 5; therefore N–OFB.	Metarhyolite was not observed at Pigeon Hills, in contrast to observations at Accomac and South Mountain.	The igneous equivalent of rift-to-drift facies is observed in samples PIGHL4 and 5. Early Iapetus seafloor? C.f. with the HLTW series, which is less mafic.

Table 4. Summary of observations, interpretations, and field observations (continued).

POPULATION	Ti–Zr–Y (Pearce and Cann, 1973)	Ti–Zr (Pearce and Cann, 1973)	Hf–Th–Ta (Wood, 1980; Dawson and Jacobson, 1989)	Primordial-mantle-normalized hygromagmatophile elements (Holm, 1985)	Nb-normalized spidergram (Myers and Breitkopf, 1989)	Chondrite-normalized rare-earth elements	Comments[1]	Conclusions
Sams Creek Metabasalt	Twelve samples plot as OFB, 5 plot near that field.	Ten OFB and 7 OTL.	Southern group E=P–OFB.[2] Northern group N–OFB.	Southern group Nb (+), therefore OIT or E=P–OFB. Some of northern group more like OFT=N–OFB.	Southern group P=E–OFB. Northern more T– and N–OFB.	Southern group LREE (+). Northern flatter and more scatter. La<Ce in GLENS and GLENT suggests N–OFB.	Southern group includes "type locality" and has carbonate caps. Metabasalt may thin to the northeast.	Southern group of P=E–OFB. Northern group of N–OFB. Therefore spreading center probably was to the southwest. Possibly related to Kennett Square?
"White Clay Creek Amphibolite"	All in WPB field. Eleven plot in one cluster.	Diagram not appropriate for WPB.	Scattered in and near E=P–OFB plus WPB.	Open $\cap$ shape, 9 have Nb (+), Sm to Y slope (–). Seven Y to Yb slope (+). E=P–OFB.	*If* OFB, then P=E–OFB.	All LREE (+); all but the low-Ti samples in a tight group. Four have La<Ce.	Field relations suggest thin, parallel flows, some having micaceous metatuff(?) beds at the top. Zr increases upsection in WCCM and LAN series. Marble and graphite in area.	Uncertain. Possibly a WPB OIT.

[1] ". . . chemical discriminant diagrams *along with geological evidence* . . . support tectonic interpretations" (Coish, 1993, p. 9).
[2] Assuming that WPB is ruled out.

Table 5. Chemical analyses of samples by population. The sites are listed in alphabetical order. All analyses are in parts per million except TiO_2, which is in percent. Detection limits vary with interferences.

NAME	TiO_2	Zr	Hf	Nb	Ta	Th	U	Ni	V	Y	La	Ce	Lat. N	Long. W
							ACCOMAC-AREA METABASALT							
ACC330E	4.06	324	9.4	37	1.8	2.4	0.8	<10	166	66	39.7	91	40°02'39"	76°33'51"
ACCOM	3.22	248	5.8	24	1.0	2.0	1.0	50	240	36	14.0	46	40°02'38"	76°33'56"
ACCPU	2.72	180	4.6	19	1.0	1.3	0.4	60	244	34	16.2	36	40°02'40"	76°33'50"
ACC48E	2.51	190	3.9	17	0.7	1.1	<0.5	60	324	32	16.0	37	40°02'38"	76°33'56"
ACC180E	1.73	94	2.0	11	0.4	0.4	<0.5	190	262	22	5.1	12	40°02'38"	76°33'55"
MEDIAN	2.72	190	4.6	19	1.0	1.3	0.4	60	244	34	16.0	37		
							"BALD EAGLE CREEK METABASALT"							
BLDEAGSW	4.27	261	6.5	30	1.7	2.9	0.6	84	310	49	35.1	83	~39°46'02"	~76°25'56"
BLDEAG2	3.32	195	5.2	23	1.3	2.3	0.4	54	268	31	35.1	70	39°46'25"	76°25'27"
BLDEAG	3.10	177	5.6	21	1.6	2.4	0.7	51	273	28	32.3	78	39°46'25"	76°25'27"
WOODB	3.03	169	5.2	25	1.6	2.4	1.0	77	227	33	36.9	73	39°47'04"	76°24'14"
MEDIAN	3.21	186	5.4	24	1.6	2.4	0.6	66	270	32	35.1	76		
							BALD FRIAR METABASALT							
BFMZ1	1.62	108	2.8	3	<0.1	0.2	<0.1	110	262	39	3.6	14	39°42'36"	76°12'20"
TLPPOP2	1.38	90	2.2	4	<0.1	0.2	<0.1	100	206	36	2.7	9	39°45'57"	76°14'27"
TLPPOP	1.36	73	2.1	5	<0.1	0.2	<0.5	85	310	36	2.3	8	39°45'56"	76°14'26"
BRANDYSE	1.34	92	2.1	<2	<0.1	<0.1	<0.1	124	239	31	2.5	9	39°56'14"	75°45'09"
SYKFM2	1.28	80	1.6	2	0.2	0.2	<0.5	80	300	34	2.4	7	39°42'28"	76°13'04"
BRANDY	1.25	80	1.8	<2	0.1	<0.1	<0.1	90	280	30	2.2	8	39°56'13"	75°45'10"
SYKFM	1.22	109	1.6	6	<0.1	0.2	<0.5	70	250	36	2.1	7	39°42'29"	76°13'04"
BRANDYN	1.22	79	1.6	3	<0.1	0.2	<0.1	84	193	25	2.2	9	39°56'14"	75°45'10"
BFMZ2	1.14	84	2.0	5	0.2	0.4	<0.1	111	262	30	4.0	13	39°42'35"	76°13'08"
PB1	1.10	81	1.8	2	<0.1	0.3	<0.1	129	210	26	2.5	9	39°45'02"	76°13'08"
MEDIAN	1.26	82	1.9	3	<0.1	0.2	<0.1	95	256	32	2.4	9		
							CATOCTIN METADIABASE DIKES IN GRENVILLE TERRANES							
PATPK	3.88	320	7.1	29	1.3	3.3	0.8	<20	372	52	30.0	64	40°05'46"	75°47'11"
ANTRESV	3.84	286	5.6	24	1.1	1.5	0.6	<50	310	42	24.0	52	40°21'16"	75°52'09"
BROOKM	3.71	230	6.1	25	1.2	1.7	0.6	20	338	50	23.0	50	~40°06'02"	~75°31'16"
HSTR	3.48	247	6.3	26	1.8	1.9	0.4	20	320	44	24.3	61	40°08'30"	75°44'10"
LS3	3.34	280	5.5	25	1.1	1.5	<0.5	40	420	44	24.0	51	40°26'29"	75°40'37"
TOPFND	3.29	178	4.1	18	1.0	1.2	0.6	53	370	36	19.0	46	40°27'55"	75°41'07"
LS2	3.29	230	5.2	21	1.1	1.3	<0.5	30	370	42	23.0	49	40°26'26"	75°14'14"
HUFFC	3.19	302	6.9	31	1.9	2.4	0.8	40	330	46	26.1	62	~40°27'10"	~75°37'46"
ISHMTN	3.02	190	3.8	17	0.7	1.1	<0.5	50	350	36	17.0	38	40°24'45"	75°52'43"
D266	2.78	225	5.2	20	0.8	1.1	0.6	<50	380	36	20.0	46	40°17'51"	76°08'21"
UWC	2.78	210	4.3	18	0.9	1.1	<0.5	50	420	38	15.0	34	40°03'53"	75°38'20"
DV	2.76	190	4.1	19	0.7	1.0	<0.5	40	420	38	13.0	31	40°05'14"	75°32'27"
LYDRY	2.74	210	4.6	19	0.8	1.2	<0.5	40	330	36	16.0	38	40°28'16"	75°45'13"
LUDCOR	2.45	153	3.7	19	0.8	1.6	0.5	<50	330	28	17.0	36	40°07'51"	75°42'51"
HUFFM	2.43	193	4.6	18	1.2	1.5	0.3	70	310	34	17.3	40	~40°27'10"	~75°37'46"
STPFA	2.31	174	3.7	17	0.7	1.1	<0.5	93	260	36	12.0	29	40°08'57"	75°01'39"
ELBNW	2.21	200	3.4	16	0.7	1.3	<0.5	50	320	30	13.0	30	40°05'26"	75°36'14"
HNYNW	1.97	130	2.8	15	0.5	0.7	<0.5	50	310	26	9.2	22	40°03'32"	75°56'47"

Table 5. Chemical analyses of samples by population (continued).

NAME	TiO_2	Zr	Hf	Nb	Ta	Th	U	Ni	V	Y	La	Ce	Lat. N	Long. W
CATOCTIN METADIABASE DIKES IN GRENVILLE TERRANES (continued)														
FURNCK	1.93	180	4.0	16	0.8	0.9	<0.5	40	250	32	17.0	38	40°19′22″	76°10′11″
HNY82	1.75	130	3.1	12	0.6	1.0	<0.5	60	290	28	11.0	24	40°04′10″	75°48′50″
MRCK	1.59	110	2.4	11	0.4	0.5	<0.5	70	290	22	8.5	19	40°04′01″	75°44′00″
MEDIAN	2.78	200	4.3	19	0.8	1.2	<0.5	40	330	36	17.0	38		
CATOCTIN METABASALT *SENSU STRICTO* (53 samples)														
MEDIAN	2.23	160	3.2	12	0.6	0.6	<0.1	87	330	36	10.8	26		
"CONOWINGO CREEK METABASALT" OF THE BALTIMORE MAFIC COMPLEX														
CONJSEIII	5.13	188	5.2	28	1.5	<0.1	<0.1	44	408	49	16.5	52	39°45′42″	76°10′25″
PLGRV	1.34	136	5.0	8	<0.1	<0.1	<0.1	114	267	12	14.3	32	39°43′51″	76°11′30″
CONJSEII	1.34	75	1.6	9	0.4	<0.1	<0.1	150	206	6	9.9	20	39°45′45″	76°10′16″
WAKES	0.98	86	2.7	8	0.2	<0.1	<0.1	139	205	18	16.4	38	39°45′32″	76°10′49″
CONJSE	0.94	41	0.8	5	0.3	<0.1	<0.1	221	554	5	4.8	10	39°45′47″	76°10′11″
APPLE	0.83	37	0.5	4	<0.1	<0.1	<0.1	39	339	3	3.8	10	39°42′04″	75°48′04″
PLGRVS	0.64	56	0.9	<2	<0.1	<0.1	<0.1	266	149	6	6.4	15	39°43′31″	76°11′41″
MEDIAN	0.98	75	1.6	8	0.2	<0.1	<0.1	139	267	6	9.9	20		
EARLY MESOZOIC DIABASE INCLUDING THE YORK HAVEN, ROSSVILLE, AND QUARRYVILLE DIABASES														
D-32	1.12	109	2.1	11	0.3	1.6	<0.5	73	220	24	8.7	18	40°06′52″	76°42′53″
ROSS	0.88	77	1.3	7	0.2	1.0	<0.5	70	254	24	5.2	12	40°04′04″	76°55′16″
QUARRY	0.41	59	0.9	4	<0.1	0.9	<0.5	320	160	20	4.0	9	39°54′02″	76°08′13″
FISHING CREEK METABASALT														
FSHCKOTC	2.33	163	3.4	16	0.7	1.0	<0.1	54	275	36	12.0	31	39°48′30″	76°14′09″
FSHCKFG	2.24	150	3.6	30	1.4	2.2	0.8	50	240	34	19.5	42	39°47′56″	76°15′23″
FSHCKRR	1.94	154	3.8	19	1.1	1.7	0.7	99	215	35	14.8	38	39°47′34″	76°16′04″
FSHCKQ	1.70	118	3.6	15	1.0	1.2	0.8	140	270	28	12.0	29	39°48′00″	76°15′15″
FSHCKINT	1.52	124	2.6	21	1.1	1.6	0.4	94	209	19	13.3	30	39°48′37″	76°13′38″
FSHCKMON	1.48	114	2.4	24	1.8	1.9	0.3	110	193	20	15.8	33	39°48′10″	76°14′50″
MEDIAN	1.82	137	3.5	20	1.1	1.6	0.6	96	228	31	14.0	32		
"HOLTWOOD METABASALT"														
HLTWCOL	2.26	110	2.6	10	<0.1	0.3	<0.5	100	410	30	7.5	18	39°49′28″	76°20′27″
HLTWM	2.24	100	2.5	9	0.3	0.4	<0.2	80	338	30	7.1	18	39°49′29″	76°20′26″
HLTWNE	1.92	97	2.1	8	0.4	0.4	<0.1	75	191	23	7.1	18	39°49′32″	76°20′26″
HLTWBASE	1.67	113	1.8	9	<0.1	0.3	<0.5	60	240	26	5.6	14	39°49′31″	76°20′26″

Table 5. Chemical analyses of samples by population (continued).

NAME	TiO_2	Zr	Hf	Nb	Ta	Th	U	Ni	V	Y	La	Ce	Lat. N	Long. W
"HOLTWOOD METABASALT" (continued)														
COYLK	1.58	98	2.1	10	<0.4	0.4	<0.5	<50	290	30	6.6	16	39°45'59"	76°16'28"
HLTWMD	1.36	86	2.8	6	<0.1	0.3	<0.1	60	317	35	4.8	13	39°40'54"	76°31'53"
MEDIAN	1.80	98	2.3	9	<0.1	0.4	<0.5	68	304	30	6.9	17		
JAMES RUN FORMATION INCLUDING THE FRENCHTOWN AND GILPINS FALLS MEMBERS														
JRFMC	2.16	97	2.4	6	0.2	0.6	<0.5	<10	76	52	6.5	18	39°35'03"	76°05'54"
JRGFAS	1.48	246	5.9	11	0.7	6.0	1.3	10	117	64	24.3	55	39°35'20"	76°06'09"
JRGFA	1.45	180	4.4	<2	0.5	3.6	0.8	<10	110	48	14.0	34	39°35'21"	76°06'10"
JRFMV	1.20	166	4.0	10	0.5	3.9	1.2	<20	120	40	17.0	37	39°35'03"	76°06'10"
JRS	0.84	59	1.3	5	<0.1	1.1	<0.5	<20	288	24	3.0	6	39°35'03"	76°05'54"
MEDIAN:	1.45	166	4.0	6	0.5	3.6	0.8	10	117	48	14.0	34		
JONESTOWN VOLCANIC SUITE														
BKHLM	1.90	150	2.7	12	0.4	0.6	<0.5	220	300	36	7.4	19	40°24'02"	76°29'17"
BKHL	1.48	95	2.0	14	0.5	0.6	<0.5	190	270	24	6.9	16	40°24'33"	76°27'50"
PA72	1.10	66	1.6	5	<0.1	0.8	<0.5	95	310	22	4.4	11	40°23'04"	76°28'33"
MEDIAN	1.48	95	2.0	12	0.4	0.6	<0.5	190	300	24	6.9	16		
KENNETT SQUARE AMPHIBOLITE														
LUCK	1.29	60	1.4	6	0.2	0.3	<0.2	90	282	24	4.1	11	39°49'50"	75°43'03"
KS	1.25	70	1.6	7	0.4	0.3	<0.1	90	266	24	4.4	11	39°50'21"	75°41'52"
WICK	1.25	86	1.9	5	0.7	0.6	<0.1	171	269	22	6.8	17	39°48'00"	75°50'05"
ROSE	0.94	78	1.6	6	0.4	0.4	<0.1	129	256	30	5.6	14	39°50'35"	75°39'32"
BRNTML	0.87	62	1.3	3	0.6	0.4	<0.1	126	234	20	4.1	11	39°50'16"	75°39'04"
CHFD	0.85	56	1.2	<2	0.1	<0.1	<0.1	80	290	26	1.2	5	39°51'44"	75°35'40"
BRANDY2	0.78	49	1.0	<2	0.4	<0.1	<0.1	90	280	20	1.1	4	39°51'12"	75°35'50"
MEDIAN	0.94	62	1.4	5	0.4	0.3	<0.1	90	269	24	4.1	11		
"OLDER DIABASE" DIKES OF BASCOM AND STOSE (1932)														
BALTP2	3.52	275	6.1	31	1.6	3.3	0.6	80	380	52	26.1	57	39°53'21"	75°30'50"
BALTP	3.42	275	6.8	28	1.7	0.8	0.4	70	370	50	13.5	38	39°53'21"	75°30'41"
BEAU2	3.19	220	5.0	27	1.2	1.5	0.5	20	332	46	20.5	45	40°01'22"	75°25'46"
BALTP3	2.18	144	3.7	5	0.6	0.6	<0.1	61	335	44	14.4	37	39°52'15"	75°36'07"
ARMK2	0.96	84	1.5	4	<0.1	0.4	<0.1	160	214	28	5.1	13	40°03'47"	75°18'24"
CPSUN	0.81	120	1.5	6	<0.1	0.6	<0.5	80	244	24	6.7	14	39°54'18"	75°31'27"
CRUMCK	0.74	50	0.7	3	<0.1	<0.1	<0.1	110	246	24	1.1	3	40°00'33"	75°28'02"
BEAU	0.59	68	0.8	4	<0.1	<0.1	<0.5	180	200	20	1.0	3	40°01'22"	75°25'46"
RADNOR	0.59	67	1.3	<2*	0.6	0.4	<0.1	234	227	20	5.2	13	40°02'06"	75°22'22"
MEDIAN	0.96	120	1.5	5	0.6	0.6	<0.5	80	246	28	6.7	14		

Table 5. Chemical analyses of samples by population (continued).

NAME	TiO_2	Zr	Hf	Nb	Ta	Th	U	Ni	V	Y	La	Ce	Lat. N	Long. W
							PIGEON HILLS-AREA METABASALT							
PIGHL1	2.48	155	3.8	11	1.0	0.7	<0.5	220	380	34	7.4	21	39°51'39"	76°58'15"
PIGHL2	1.85	110	2.5	7	<0.1	0.3	<0.5	130	306	36	7.0	18	39°52'10"	76°57'47"
PIGHL888	1.73	102	2.2	8	0.2	0.2	<0.5	130	250	24	7.4	19	39°52'40"	76°56'52"
PIGHL3	1.35	90	1.9	9	0.3	0.3	<0.5	120	248	28	5.6	14	39°51'33"	76°58'09"
PIGHL5	1.30	86	1.5	<2	<0.1	0.1	<0.1	151	317	28	3.2	9	39°52'09"	76°57'24"
PIGHL4	1.28	81	1.7	<2	0.2	<0.1	<0.1	138	324	20	3.3	11	39°52'04"	76°57'27"
MEDIAN	1.54	96	2.0	8	0.2	0.2	<0.5	134	312	28	6.7	16		
							SAMS CREEK METABASALT							
GLENRK	2.63	133	4.6	12	<0.2	0.7	0.7	190	460	44	13.0	32	39°46'14"	76°42'56"
DISEQ	2.32	170	3.9	21	1.1	1.8	<0.1	314	440	60	20.4	39	39°46'41"	76°43'23"
SC340NE	2.30	180	3.3	24	1.1	1.4	<0.5	80	324	36	14.0	31	39°30'28"	77°07'46"
PCSC4	2.14	120	3.3	5	0.5	0.5	<0.1	97	397	44	9.9	25	39°47'40"	76°44'00"
PA616	2.02	104	2.5	5	0.4	0.4	<0.1	36	454	64	6.8	17	39°46'13"	76°42'43"
PCSC3	2.02	145	3.5	21	1.4	1.6	<0.1	148	328	36	16.5	38	39°47'17"	76°43'24"
PCSC	1.97	160	3.0	22	0.8	1.2	<0.5	160	310	38	14.0	29	39°46'33"	76°43'22"
SC240NE	1.93	155	3.3	26	1.4	2.0	0.2	144	307	30	18.7	39	39°30'28"	77°07'47"
GLENM	1.90	110	2.8	10	0.4	0.6	<0.2	80	338	36	7.1	18	39°47'36"	76°43'49"
SCTYPE	1.88	151	3.1	28	1.5	2.1	0.3	78	304	32	16.6	36	39°30'00"	77°07'01"
SC40NE	1.88	141	3.0	21	1.0	1.4	<0.1	107	296	36	12.2	26	39°30'27"	77°07'50"
SC440NE	1.80	118	2.4	19	0.9	1.5	0.3	157	349	32	14.3	31	39°30'29"	77°07'45"
SCQ	1.74	120	2.8	11	0.5	0.8	<0.2	70	300	36	8.6	21	39°47'03"	76°45'09"
GLENT	1.70	93	2.3	5	0.3	0.2	0.1	112	417	36	4.9	15	39°47'38"	76°43'49"
SC144NE	1.50	141	2.6	20	0.9	1.2	<0.5	210	190	26	12.0	26	39°30'27"	77°07'49"
GLENN	1.47	60	1.8	4	0.2	0.2	<0.5	360	340	28	3.5	9	39°47'36"	76°43'49"
GLENS	1.39	70	1.6	6	0.1	<0.1	<0.1	140	284	30	2.3	7	39°47'36"	76°43'46"
MEDIAN	1.90	133	3.0	19	0.8	1.2	<0.2	140	328	36	12.2	26		
							"WHITE CLAY CREEK AMPHIBOLITE"							
WCCMBIV	4.75	261	6.7	28	1.2	1.9	<0.1	41	284	41	15.3	43	39°47'20"	75°48'09"
YRKLYN	4.64	264	7.1	19	1.6	1.8	0.6	25	442	50	21.5	55	39°48'32"	75°40'05"
LANS	4.50	280	6.6	27	1.1	1.5	<0.5	20	450	50	22.0	49	39°46'21"	75°46'03"
WCCW	3.44	131	4.7	13	0.6	1.4	0.8	131	635	22	14.9	40	39°44'38"	75°47'38"
LANM	3.04	160	4.3	18	0.9	1.1	0.7	80	444	38	17.3	37	39°46'36"	75°45'40"
WCCMBI	2.88	125	3.9	14	0.7	1.8	1.4	54	525	28	15.3	39	39°47'20"	75°48'09"
WCCMBIII	2.64	164	5.2	17	0.9	1.8	0.8	69	283	30	18.4	50	39°47'20"	75°48'09"
WCCN2	2.38	122	4.1	14	1.2	1.6	<0.1	66	423	23	12.8	36	39°45'38"	75°46'04"
WCCMBII	2.36	156	3.7	18	0.4	1.1	0.3	111	234	24	15.4	37	39°47'20"	75°48'09"
MP5	2.04	124	3.0	11	1.0	0.8	<0.1	209	245	28	10.6	27	39°46'55"	75°43'47"
LANN	1.94	110	2.7	9	0.3	0.7	<0.5	130	290	26	8.5	20	39°46'26"	75°46'11"
WCC	1.66	108	2.8	10	0.7	1.2	<0.1	85	273	24	15.6	26	39°44'24"	75°46'25"
MEDIAN	2.71	140	4.2	16	0.9	1.4	<0.5	75	356	28	15.4	38		

*Analytical error for Nb in sample RADNOR suspected based on Nb/Ta.

Bald Friar Metabasalt (new name)

This unit is herein named for exposures within a previously undescribed mélange (mapped by Higgins and Conant (1986) as part of their Sykesville Formation) on the north side (2 samples) and southwest side (2 samples) of the hill known as Bald Friar in Cecil County, Maryland. The six samples of this metabasalt from Pennsylvania occur in what was previously mapped as part of the Peters Creek Formation but is very likely not. As suggested by the histogram for Ta (Figure 14) using a traditional abscissa scale, the sampled population is extremely uniform despite sampling on both the north and south sides of the Peach Bottom structure along the Susquehanna River and 48 km ENE of the river. Pillows of variable quality are present at two of the locations, and sample BFMZ1 has linear extrusion-like features. The Bald Friar Metabasalt can be mapped in the field on the basis of its dark green color, laminated character, and especially its high density. Outcrops of ultramafic rock typically occur within a few tens of meters, but at one locality (Stop 14) they are within a few centimeters. At the Chester County locality (Stop 7), the closest recognized ultramafic rock is 1.4 km to the southwest.

Because of the extreme chemical uniformity of the samples (Table 5), it is very likely that the ophiolitic mélange exposed on the N side of the Peach Bottom structure is the same one exposed on the N side of Bald Friar, Maryland, well to the south of the Peach Bottom structure. This ophiolitic mélange may also account for the steatitized ultramafic clast at the SE side of the Peach Bottom Structure (Stop 14, Sample SETALC), but no diagnostic metabasalt fragments have been found with this steatitized clast.

The discriminant diagrams (Figure 12) suggest an ocean-floor basalt having a minimal range of compositions. The diagrams also consistently indicate normal ocean-floor basalt, i.e., one that has not been enriched by a plume. Normal ocean-floor basalts are thought to form at linear spreading centers. Such spreading centers, however, are not restricted to mid-ocean ridges. Further interpretation is somewhat limited by the fact that most of the known occurrences are at structural discontinuities and much of the metabasalt appears to occur as fragments within an ophiolitic mélange. Thus, the relationship to type Peters Creek Formation is not obvious. Based on proximity to ultramafic mélange fragments that are likely from the Baltimore Mafic Complex, the Bald Friar Meta-basalt also appears to be part of that island-arc ophiolite complex. If so, these basalts are likely from the back arc, i.e., SE side of the complex. For further discussion, see the comments for Stops 7 and 14.

Catoctin Metabasalt sensu stricto

This suite consists of 53 metabasalt samples from the South Mountain section of the Blue Ridge province of Adams and Franklin Counties, Pennsylvania (inset to Figure 11). Over 40 of the samples were collected from outcrop, the rest from float. Smith and others (1991) included photographs and/or descriptions of pillows, pipe vesicles, pyroclasts, ropy pahoehoe, pahoehoe toes, probable agglomerate-agglutinate, and thin flows having chilled tops. They also noted that much of the exposed metabasalt was overturned. More recently, a probable tuff bed having rhyolitic trace-element contents and occurring within metabasalts (39°48'17"N, 77°23'32"W) has been roughly correlated chemically with Catoctin Metarhyolite (39°49'11"N, 77°27'29"W).

Based on systematic changes within a given diagram that correlate with TiO_2 variation, the Catoctin magmas evolved appreciably with time; possibly as rifting progressed to drifting. Histograms, such as for TiO_2, suggest several pulses of basaltic magma, each of which also differentiated over time in shallow magma chambers. Some of the diagrams suggest a transition from within-plate to oceanic basalt, and the primordial mantle spidergram clearly suggests a continental initial-rifting tholeiite for the main pulse.

Badger and Sinha (1988) reported a Sr isochron age of 570 ± 36 Ma for Catoctin Metabasalt from Virginia. Aleinikoff and others (1991) reported an age of 597 ± 18 Ma for euhedral zircons in Catoctin Metarhyolite from Pennsylvania, but noted that the metarhyolite also contained partially resorbed Grenvillian zircon xenocrysts. Thus, the Catoctin Metarhyolite, if not the slightly older metabasalt, might contain an inherited Grenvillian component.

All-in-all, the Catoctin may vie as the most underrated tectonomagmatic event in eastern North America.

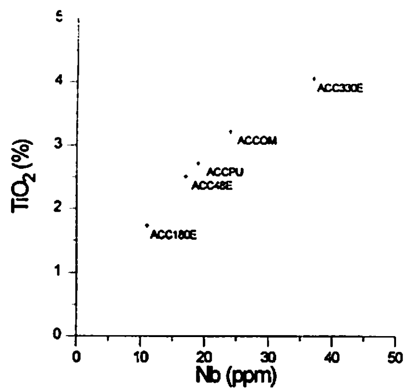


Figure 13. Plot of Nb vs. TiO_2 for the Accomac samples, which are considered to be a vignette of the rifting phase of the Catoctin event.

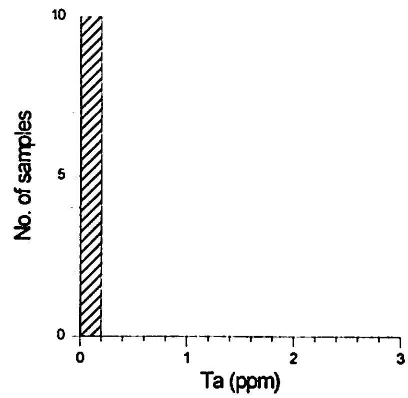


Figure 14. Histogram for Ta for the Bald Friar Metabasalt, Lancaster, York, and Chester Counties, Pennsylvania, and Cecil County, Maryland, using a conventional abscissa of 0 to 3 ppm. This suggests that the population represents one uniform pulse despite the geographic spread.

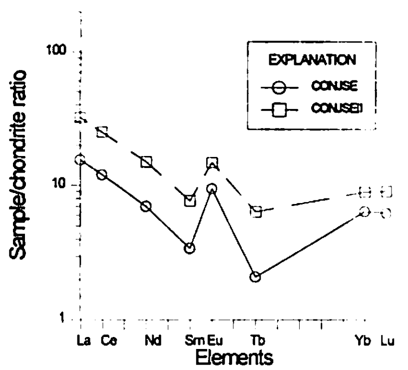


Figure 15. Chondrite-normalized rare-earth plot for samples CONJSE and CONJSEII from the informal "Conowingo Creek metabasalt" population. Both samples appear to be of volcanic rock from within the Baltimore Mafic Complex and have a "steerhorn" pattern of intermediate rare-earth depletion considered by many to be diagnostic of boninites (Coish, 1989, p. 275).

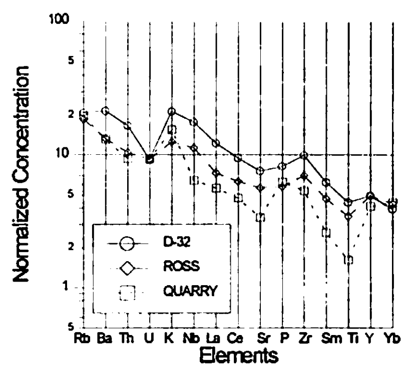


Figure 16. Plots of hygromagmatophile elements in early Mesozoic diabase normalized to primordial mantle, following procedure of Holm (1985). The plot shows no negative Nb anamoly.

Catoctin Metadiabase Dikes in Grenville Terrane

The suite consists of 21 samples from the Reading Prong, Womelsdorf outlier, Honey Brook Upland, and Trenton Prong (Figure 11). The samples were collected from dikes, most of which trend ENE, have well-preserved chilled margins, and are typically 1 to 5 m wide. All samples are from terranes that have undergone Grenvillian metamorphism, but the dikes themselves are relatively unmetamorphosed and undeformed. Because the metadiabase dikes occur in all of the Grenville terranes in Pennsylvania except Mine Ridge and the Brandywine massifs, these may have been distant at 570 Ma, the presumed time of intrusion. Samples LS3, MRCK, and HNY82 contain primary plagioclase phenocryst laths. Samples D266 and HSTR contain sulfide globules in the 2- to 5-mm range. Reports of metadiabase dikes cutting Lower Cambrian Hardyston quartzite in the Reading Prong were not confirmed upon revisiting the sites. Neither have such dikes been observed to cut the Wissahickon Schist of Berg and others (1980), further weakening any remaining arguments that the Wissahickon is Precambrian.

Despite their widely scattered occurrence, the dikes form a rather consistent group on most diagrams. Comparison of the dike medians with those of Catoctin Metabasalt *sensu stricto* suggests that the dikes are somewhat more fractionated or enriched in crustal components than the latter. By analogy with HUFFM (2.43% TiO_2) and the crosscutting HUFFC (3.19% TiO_2), the dikes are somewhat younger than the main stage of the Catoctin flows. Comparison of the dike medians with those for the Accomac-area metabasalt suggests that they are essentially one and the same.

The trace-element diagrams strongly suggest that the dikes are tholeiitic within-plate basalts related to initial continental rifting. This is consistent with their intrusion as dikes related to the Catoctin event. Because of the widespread occurrence of Catoctin dikes in Grenville terranes (Table 5), it is possible that remnants of Catoctin flows will someday be recognized in those terranes. Quadrangles that have been the subject of modern mapping, such as the Easton quadrangle mapped by Drake (1967), are obvious targets.

"Conowingo Creek Metabasalt"

This suite, which is informally named because it is presently ill-defined, consists of 7 samples, six of which occur intimately associated with serpentinite-bearing mélange on the northern margin of the Baltimore Mafic Complex (BMC) and overlook Conowingo Creek and its tributaries in southern Lancaster County (Figure 11). Of these six, the CONJSE-series samples contain variable evidence of pillow textures. CONJSEIII, in fact, suffered from pillow enrichment disease[1] (Winchester and Floyd, 1976, p. 461). Although unlikely, a metasomatic blackwall origin (Sanford, 1982) should be considered for sample PLGRVS. The seventh sample, APPLE, is from extreme northeastern Maryland in a somewhat similar terrane and was tentatively included in this group because of its location.

Plots of chondrite-normalized rare-earth elements (REE) for samples CONJSE and CONJSEII indicate an intermediate REE depletion, informally called a "steerhorn" pattern (Figure 15) which, in volcanic rocks, is frequently considered diagnostic of boninites (Coish, 1989, p. 275). Boninites and their boninitic associates and differentiates are also characterized by high SiO_2, MgO, Cr, and Ni, and low TiO_2, Y, Zr, and V[2] (Bloomer and Hawkins, 1987). Modern boninites are found only in association with island arcs above subduction zones. They are generally believed to result from products of melt from mantle metasomatized by fluids emanating from a subducted slab into a hot, shallow mantle region that has previously been the source of one or more melts (Crawford and others, 1989, p. 1-49).

Rather interestingly, the *de facto* type ophiolite of Troodos, Cyprus, contains boninites in the upper pillow lavas and is now widely regarded to be an island-arc complex, the boninites having formed in the forearc environment (Robertson and Xenophontos, 1993; Crawford and others, 1989). In a sense, Coleman (1977, p. 19) foresaw this when he stated, "Our present state of knowledge indicates that unaltered volcanic rock types from island arcs, marginal basins, small ocean basins, [and] mid-ocean ridges are present within ancient ophiolites." Hopefully the works of Crawford and Coleman will not be too disturbing to those who associate ophiolites exclusively with midocean ridge basalts (MORB), despite Shaw and Wasserburg's caution (1984, p. 342). Likewise, this is hardly the first report of boninites associated with ophiolites in the Appalachians. Coish (1989) and Tremblay (1992) have thoroughly documented boninitic basalts associated with the Thetford Mines ophiolite, Québec, and the Betts Cove ophiolite, Newfoundland, both of which mark continental (Humber) vs.

* * * * * * * * * * * * *

1 Enrichment of immobile elements in the glass rims, presumably in part by loss of SiO_2 during devitrification.

2 A 1.0-m-thick, highly sheared rock, sample ABVBIF, located above the "banded iron formation" (BIF)-bearing BAld Friar Metabasalt noted in the Stop 14 description meets these criteria.

oceanic (Dunnage) terrane boundaries.

As noted by Coish (1993), "The extremely low contents of Ti, Zr, and Y of boninites invalidates the use of standard geochemical diagrams to pinpoint their tectonic environment." Nevertheless, extremely low Y contents of 3 to 6 ppm in some of the "Conowingo Creek Metabasalt" samples (Table 5) are suggestive of at least a boninitic affinity. Primordial-mantle-normalized spidergrams for CONJSE and CONJSEII somewhat resemble those prepared for 21 well-documented boninites, but the sample-to-sample variation of the latter is substantial.

The occurrence of pillow basalts of apparent boninitic affinity adds further evidence to the ophiolitic nature of the Baltimore Mafic Complex (BMC). Thus, the "Conowingo Creek Metabasalt," long recognized serpentinites and gabbros, and the James Run Formation and Port Deposit Tonalite (Hanan and Sinha, 1989) appear to be part of an island-arc-derived ophiolite complex that has likely received a continental component, possibly through subduction of Laurentian detritus and metasomatism during the Taconic orogeny. Should any choose to cite the Nd and Sr isotopic data of Shaw and Wasserburg (1984) as evidence that the BMC is not an ophiolite, we encourage careful reading of their admirably forthright paper which implies that the BMC was not classified as an ophiolite because they were not aware of the pillow basalts. Also worth reading is Gunter Faure's discussion of the $^{143}Nd/^{144}Nd$-$^{87}Sr/^{86}Sr$ mantle array, which he notes "may be lapsing into *disarray* as additional rocks from the ocean basins are analyzed" (Faure, 1986, p. 217-218). More recently, Crawford and others (1989, p. 2) explained low ϵ_{Nd} and radiogenic Sr in boninites as the result of the subduction of a slab containing recycled ancient crust such as pelagic sediments. If crust had been subducted during formation of the BMC, it seems that it would have been easiest with a SE-dipping subduction zone, similar to the case in the northern Appalachian ophiolites (See "Conceptual Model of Selected Metabasalts in the Piedmont").

Given the boninitic affinity of the parental magmas of the Merensky reef of the Bushveld Complex and the Stillwater Complex (Crawford and others, 1989), two important sources of platinum-group elements, it is hoped that study of the BMC will continue. This hope is not diminished by the formerly commercial gold-platinum-group placers of the Rivière-des-Plantes ophiolite mélange of Québec.

Early Mesozoic Diabase

This suite consists of only 3 samples (Table 5), two from the type localities for the York Haven and Quarryville Diabases (Smith and others, 1975) and the third from the principal reference section for the Rossville Diabase. The York Haven and Rossville Diabases are younger than most of the associated continental rift-related sediments. Of the traditional diagrams, only the primordial-mantle normalization diagram of Holm (1985) "correctly" classifies them as continental to initial-rifting diabases (Figure 16). The Ti-Zr and Hf-Ta-Th diagrams tend to "incorrectly" classify these diabases as calc-alkaline (Holm, 1982, and present study). However, as noted by Bloomer and Hawkins (1987, p. 374), "In most discriminant diagrams the boninites plot as calc-alkaline arc basalts." Why, then, does examination of these typically reliable Ti-Zr and Hf-Ta-Th diagrams indicate that the early Mesozoic diabases are calcalkaline or, even more absurdly at first glance, as boninitic? Further, why does Quarryville Diabase have a "steerhorn" chondrite-normalized rare-earth pattern (Figure 17A)? As suggested by many workers, such depletion in the intermediate rare-earth elements in volcanic rocks occurs only in those formed from magmas that have been depleted by prior melting and then "mantle metasomatized" by the introduction of LREE, Si, Na, . . . -bearing fluids. In the case of boninites, such fluids are typically attributed to derivation from the subducting slab. For the early Mesozoic diabases, we are not aware of other evidence suggesting that the successful early Mesozoic proto-Atlantic rift supplied the failed NW rift (the Mesozoic basin in Pennsylvania) with a subducted slab. Thus the island-arc aspect of calc-alkaline and boninitic basalts does not seem applicable to the

Mesozoic diabases at this time.[3] This may not, however, eliminate the possibility that the source for the oldest early Mesozoic diabase, the Quarryville Diabase, underwent prior melting. Indeed, mantle depletion by the Catoctin event was proposed by Smith and others (1991), who postulated that " . . . the mantle beneath southeastern Pennsylvania has remained attached to the continental crust, as a keel is to a boat, from ~600 Ma to ~200 Ma" (Smith and others, 1991, p. 19). More recently, analysis of a xenolith (?) nodule from the upper part of the preserved Catoctin Metabasalt in Pennsylvania showed that it may represent rock depleted by the main stage of the Catoctin. This nodule has a rather "anti-boninitic" rare-earth pattern (Figure 17B)!

We are thus suggesting that consideration be given to three things: (1) A possible relationship between the Catoctin and early Mesozoic rifting events. (2) The magnitude and lateral extent of the Catoctin event. (3) The possible value of *constructive* autopsies of apparently *wrong* answers on discriminant diagrams.

Fishing Creek Metabasalt (new name)

The data set consists of 8 analyses of samples from six localities on the SE limb of the Tucquan anticline along or near Fishing Creek, Lancaster County. The Fishing Creek metabasalt was mapped by Knopf and Jonas (1929) as a 0.3-km-long body, but it was well camouflaged on their Plate I. Its mapped length was extended to 4 km during the present study.

Sample FSHCKINT has an apparent intrusive texture. Efforts to trace metabasalt farther NE have failed despite the fact that float of the phyllite that typically forms its hanging wall can be traced in that direction. Thus, the paleoslope of the land surface may somewhat resemble that of the present, or vice versa! Talc schist of definite ultramafic origin occurs along apparent strike 1.4 km SW of the SW shore of the Susquehanna River. This float suggests the presence of a zone that received a variety of igneous oceanic-floor debris. Stose and Jonas (1939) mapped a 2.7-km-long zone of serpentinite that includes the above-mentioned talc schist occurrence but, except for that rock, only a single piece of metabasalt float was found at approximately 39°45'39"N, 76°18'45"W. Because talc-anthophyllite-chlorite float was found on the NW limb of the Tucquan anticline at 39°44'15"N, 76°39'00"W, correlation of folded and thrust-faulted mélange zones yielding ultramafic and mafic float across the anticline should be attempted.

It is herein recommended that the Fishing Creek Metabasalt be established as a formal stratigraphic unit because it is field mappable, useful as a marker, and has a rather consistent chemistry. The type locality is the exposure along the SW side of Fishing Creek (39°47'58"N, 76°15'16"W) 1.3 km NE of the Susquehanna River, where some primary features are preserved. At this locality it is estimated to be approximately 30 m thick. As presently mapped, the hanging wall of the Fishing Creek Metabasalt is typically a muscovite-paragonite-quartz phyllite. The footwall is typically a quartz-chlorite-muscovite schist. For additional information, please see comments for Stop 11.

The Fishing Creek Metabasalt is relatively uniform chemically along strike (Figure 12) despite the variation in stratigraphic level sampled. It exhibits geochemical features of both within-plate ocean-island tholeiite, such as modern Hawaii, and P=E-OFB, such as the modern 45°N Mid-Atlantic Ridge. This is not too surprising in that both involve hot spots on the ocean floor. The typically reliable primordial-mantle-normalized diagram of Holm (1985) suggests that the Fishing Creek Metabasalt is P=E-OFB, but an intermediate environment is possible.

* * * * * * * * * * * *

3 As shown in Figure 16, the chemical analyses for the early Mesozoic samples (Table 2) do not indicate a negative Nb anomaly as plotted on the diagram of Holm (1985). Likewise, the Catoctin *sensu stricto* rocks do not have a negative Nb anomaly. Holm's (1985) diagram of initial-rifting continental tholeiites also lacks a negative Nb anomaly. This is in apparent contrast to the model of Pegram (1990), which seems to require a negative Nb anomaly and pre-Grenvillian subduction of an island arc. We do not, however, wish to dismiss Pegram's study, which is likely to yield additional interpretations.

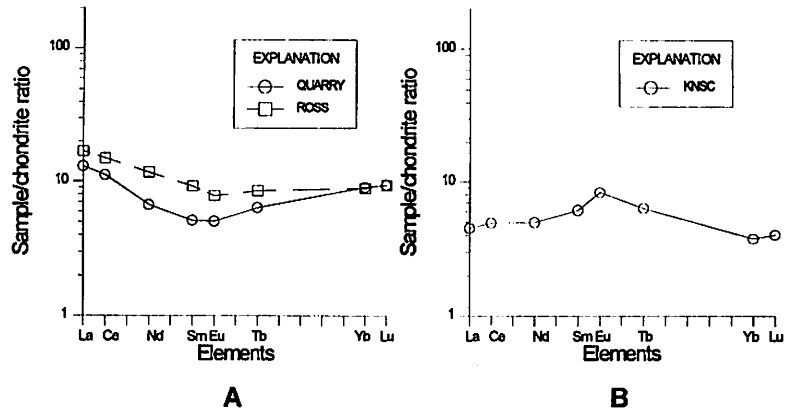


Figure 17. A. Chondrite-normalized rare-earth plots for chilled-margin samples from the Quarryville Diabase type locality and Rossville Diabase principal reference section. Based on the intermediate rare-earth depletion of sample QUARRY, which may also be the parental magma to sample ROSS, it may have been derived from a region of mantle that had produced an earlier magma.

B. Chondrite-normalized rare-earth plot for a xenolith (?) in Catoctin Metabasalt from high in the apparent metabasalt section. Its anomalous, "anti-boninitic" pattern suggests the possibility that it represents some sort of refractory residue.

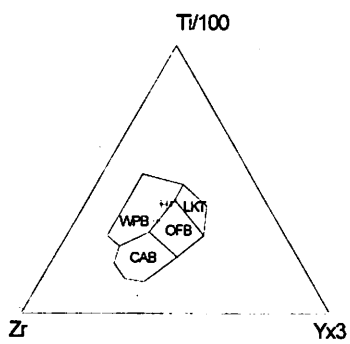


Figure 18. Ti–Zr–Y diagram for the four samples from the Holtwood Dam area. All four samples fall within the field for within-plate basalts (WPB) on this diagram but plot as initial-rifting continental tho-leiites on the primordial-mantle-normalized diagram of Holm (1985).

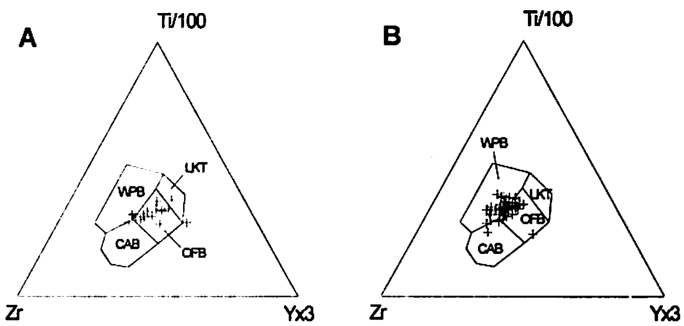


Figure 19. Ti–Zr–Y plots of the Sams Creek population (A) and Catoctin *sensu stricto* samples (B), suggesting that the Sams Creek samples are ocean-floor basalts whereas the Catoctin *sensu stricto* is transitional from within-plate to ocean-floor basalt.

"Holtwood Metabasalt"

The "Holtwood Metabasalt" suite consists of 4 samples of metabasalt from the structurally complex locality of E. B. Mathews (Stose and Jonas, 1939, p. 85) at the Holtwood Dam, York County, and, for convenience, two possibly vaguely related samples, one from northwest of the Peach Bottom nuclear power station (COYLK) and the other from south of Norrisville, Maryland

(HLTWMD) (Table 5). Primary igneous textures are poorly preserved and the metabasalt at Holtwood Dam could be intrusive. There are only vague suggestions of a few pillow structures SW of the fish ladder race. Chalcopyrite veinlets occur between HLTWB and HLTWNE.

The Holtwood Dam samples (Figure 18) plot as within-plate or initial-rifting continental tholeiites and seem to be related to the Catoctin event. The Holtwood Dam samples appear to represent a transition from initial continental rifting (represented by the bulk of the Catoctin *sensu stricto*, the Accomac-area samples, and the dikes in Grenville terranes) to the drifting of the continental plates and the formation of the Iapetus Ocean (represented by the continuation of the Catoctin from Maryland, exposed SE of the Tunnel Hill-Jacks Mountain fault system SW of Fairfield, Adams County, by the Pigeon Hills Metabasalt on strike with the Catoctin across the Mesozoic basin, and by the northern subgroup of the Sams Creek Metabasalt). The Holtwood Dam samples are more enriched in LREE and less mafic than the Pigeon Hills samples, which are believed to be more oceanic.

Based on its lower TiO_2 and Zr relative to Y, sample COYLK is a bit more OFB-like, indicative of drifting. HLTWMD is depleted in light and intermediate rare-earth elements (L and IREE) suggesting that it may be late, i.e., from a predepleted mantle region. On most diagrams it plots as P=E-OFB, but one sample does not an interpretable population make.

James Run Formation

This suite consists of 5 samples from the type section (Higgins, 1977) along the Susquehanna River in Cecil County, Maryland. Of two samples from the Frenchtown Member, one is from an amygdaloidal flow and one is from the chilled zone of a sheeted(?) dike that has only one apparent chilled contact. Both samples from the Gilpins Falls Member contain common amygdules. Sample JRD is from a dike cutting the transitional contact between the Port Deposit Tonalite and the Happy Valley Branch Member of the James Run Formation.

The trace element data suggest an evolving island arc (see range of lithologies and chemical compositions in Higgins and Conant (1990), as well as new data in Table 5) near a continental margin which, of course, implies subduction near an active margin. The spread of Ti-V contents suggests a back-arc environment. The elevated SiO_2 and low MgO, Cr, Co, and Ni suggest an andesitic affinity. The *slightly* depleted intermediate REE "steerhorn" pattern of JRS suggests that it may represent the second partial melt derived from the amphibolite that provided the Port Deposit Tonalite. Hanan and Sinha (1989) affirm the possible trondhjemitic affinity of felsic metavolcanic gneisses of the James Run Formation, which lack K-feldspar (Lesser, 1982). All of this is consistent with the interpretation of A. Krishna Sinha (personal communication, 1988) and Hanan and Sinha (1989, p. 13-15): "The association of the James Run tholeiitic and calc-alkaline volcanism in time and space with the BMC peridotitegabbro is suggestive of a continental margin volcanic arc on the oceanward side of a back arc basin." This was also discussed under the "Conowingo Creek Metabasalt," above.

Jonestown Volcanic Suite

Because of the availability of the holistic work of Lash (1986), the present study of this suite was limited to only 3 samples (Figure 11 and Table 5). The three samples are BKHL, from a hexagonal basalt column in a breccia; BKHLM, massive metabasalt from a section of pillows; and PA72, metadiabase from a sheet. One reconciliation of the limited data from the present study suggests an island arc that included a back-arc spreading center not directly related to subduction (BKHL and BKHLM) and hypabyssal intrusions related to subduction and to the main stage of arc development (PA72). Additional sampling and analyses are recommended.

Kennett Square Amphibolite

This suite consists of seven samples (Table 5) from outcrops of now-separate amphibolite bodies along the SSE side of the Brandywine massifs, roughly covering the belt from Chadds Ford to West Grove, Chester County. Except for sample WICK, which is from the west end of the belt and contains possible epidote-filled amygdules in an amphibole-bearing matrix, amphibolite-facies metamorphism has obscured primary features.

All 7 samples are of ocean-floor basalt (Table 4 and Figure 12), but are transitional from

N-OFB on the east to P=E-OFB on the west. Correspondingly, Zr generally increases from east to west. Based on the mapping of Bascom and Stose (1932), some of the P=E-OFB bodies might possibly be associated with as-yet undiscovered marble which would be interpreted as carbonate reef caps.

Metamorphosed mid-ocean-ridge basalts (MORB) *sensu stricto* cannot typically be distinguished from back-arc-basin spreading-center basalt (BABB) on the basis of geochemical data. Other types of data are lacking for this group. Whatever the type of spreading center, it seems likely that it was located toward one of the ends of the presently known belt.

The detectable Au and highly variable Cu contents of the Kennett Square Amphibolite suggest the potential for metallic ore minerals in the area, perhaps near contacts with carbonate rock.

As noted by M. E. Wagner (personal communication to R. T. Faill, circa 1990), the mafic body on the east side of Brandywine Creek (sample CHFD of the present study) is not continuous across U.S. Route 1 as mapped by Bascom and Stose (1932). This was confirmed by an examination of recently exposed outcrops of intensely deformed arkosic conglomerate(?) along the north side of U.S. Route 1. However, amphibolite samples BRANDYB and DILW from higher altitudes north of U.S. Route 1 (Table 6) do not, unlike many of the rocks of the Brandywine massifs, contain garnet. Might they be remnants of a P=E-OFB thrust over the Brandywine massifs? Based on their having a different slope on a TiO_2-Nb plot, they are not close relatives of the Kennett Square Amphibolites. Their TiO_2-Nb and Hf-Y plots do, however, have similar slopes to plots of samples from the Wilmington Complex east of the Rosemont fault.

Table 6. Chemical analyses of samples DILW and BRANDYB from an apparently garnet-free mafic body within the Brandywine massif north of U. S. Route 1. All analyses are in parts per million except TiO_2, which is in percent.

NAME	TiO_2	Zr	Hf	Nb	Ta	Th	U	Ni	V	Y	La	Ce	Lat. N	Long. W
DILW	2.55	185	4.6	16	1.4	1.5	0.4	89	364	44	13.4	32	39°53′44″	75°34′00″
BRANDYB	2.14	154	3.9	13	0.7	0.7	<0.1	77	444	40	12.9	32	39°52′33″	77°34′28″

"Older Diabase" Dikes of Bascom and Stose (1932)

This suite consists of 9 samples (Table 5) from typically 0.1- to 3-m-wide dikes that retain well-developed chilled margins despite their having been metamorphosed to garnet grade. All occur in the Brandy-wine massifs which, based on the chilled margins, appears to have been relatively cool at the time of intrusion.

Chemically, the suite consists of a higher TiO_2 ($\geq$2.0%) group that contains low amounts of CaO and MgO and a low-TiO_2 ($\leq$1.0%) group that contains higher amounts of CaO and MgO. Both occur in a small outcrop at 40°01'23"N, 75°25'47"W, where the relative ages could probably be determined. The higher TiO_2 group likely represents initial-rifting continental tholeiite but interpretation of the lower TiO_2 group is less certain. Based on the usually reliable LREE depletion indicating N-OFB, BEAU and CRUMCK may, indeed, be N-OFB and the others some sort of transitional OFB. Based on the TiO_2 vs. Nb plot, there is little if any evidence of fractionation between the two groups. However, the low-TiO_2 "older diabase" dikes may fit the Catoctin *sensu lato* trend for TiO_2 vs. Nb.

Pigeon Hills-area Metabasalt

The Pigeon Hills are on the SE side of the Mesozoic Basin (inset to Figure 11) along the York-Adams County line on the projection of the South Mountain of Maryland (MacLachlan, 1991). This suite consists of 6 rather mafic samples, mainly from natural outcrops (Table 5). Samples *IGHL1 and PIGHL888 could have been enriched in the most immobile elements by loss of SiO_2 and/or CaO by weathering. Stose and Stose (1944) mapped quartzite o the Chickies Formation and possible tuffaceous slate in the area, but not definitive metarhyolite. Amygdules are present in many samples; one piece contained sparse, disseminated 0.1-mm grains of native Cu.

On most diagrams, the fresher samples plot as ocean-floor basalt (OFB), such as typically forms at mid-ocean ridges and back-arc spreading centers. Based on chondrite-normalized rare-earth elements (REE) and the Nb-normalized spidergram, PIGHL4 and 5 are normal (N-OFB) and the others transitional (T-OFB) or even plume=enriched (P=E-OFB). The primitive-mantle-normalized spidergram suggests a hint of initial-rifting continental tholeiite in all samples but PIGHL4 and 5. An origin during the Iapetan drifting phase, which included the transition from Catoctin rifting, is strongly suggested. A few of the easternmost Catoctin *sensu stricto* samples on the Catoctin geographic trend from Maryland, such as sample PADOT (39°44'14"N, 77°22'45"W) from SE of the Tunnel Hill-Jacks Mountain fault system SW of Fairfield, Adams County, Pennsylvania, closely resemble the Pigeon Hills-area Metabasalt.

Sams Creek Metabasalt

This suite consists of one sample, SCTYPE, from the *de facto* type locality (Table 5), 5 samples, SC340NE, SC240NE, SC40NE, SC440NE, and SC144NE, from the *de facto* principal reference section along Maryland Route 31 (Table 5), and 11 samples from southwestern York County, Pennsylvania. Sampling of the known sections in Maryland was necessary for comparison because the name "Sams Creek" is herein being formally proposed for usage in Pennsylvania. Stose and Jonas (1939) mapped the continuation of the Sams Creek Metabasalt of Maryland into Pennsylvania without assigning a name to it. Their mapping showed the unit as being more extensive in Pennsylvania than appears to be the case. Except for the excellent section at Glen Rock, which begins with pillow basalt on the SE, many of the outcrops in Pennsylvania may be chaotic fragments in a mélange. Sample SCQ has a coarse, apparently intrusive texture.

Unlike the Catoctin metabasalt, the Sams Creek has strong OFB tendencies (Figure 19). Specifically, the type and reference sections in Maryland and some of the samples from Pennsylvania are P=E-OFB whereas most of the samples from the Glen Rock area are transitional or N-OFB and may be related to the "Holtwood Metabasalt". Assuming that the source is likely to be closer to the SW end of the belt where the Sams Creek Metabasalt is more abundant, this back-arc spreading center was located to the SW of Pennsylvania (present direction). In this terrane, a Nb content $\geq$18 ppm distinguishes P=E-OFB that are associated with probable carbonate reef caps. Such reef caps would have aquifer potential and might contain Cu-Pb-Zn-Ag-Au ore deposits of the type found in the Lingonore District, Maryland (Singewald, 1946). In this they differ from N-OFB, which have little potential for either.

"White Clay Creek Amphibolite"

This suite consists of 12 samples, most of which were collected along White Clay Creek north of the Arc Corner, the point at which the arc that forms the boundary of Pennsylvania and Delaware meets the Mason and Dixon Line, in Chester County. Primary features have been largely destroyed by amphibolite-facies metamorphism, but the shape of some of the bodies as mapped suggests that they are thin flows, a few of which may be overlain by highly micaceous tuffs. The amphibolite bodies may become more alkalic upsection, based on analyses of the LAN and WCCMB series samples, assuming that each is right side up.

The wide range of compositions is typified by TiO_2, which constitutes from 1.66 to 4.75% of the rock. This range is similar to that in the Catoctin Metabasalt *sensu stricto* and its outliers. TiO_2-Nb plots for White Clay Creek rocks (Figure 20), however, crosscut those for rocks formed during the Catoctin event, such as the Accomac samples (Figure 13). The "White Clay Creek Amphibolite" may have begun as a within-plate ocean-island basalt (OIB). Conceivably, the marble to the south of sample LANS, from the presumed youngest pulse, based on stratigraphic position, might have been a reef cap. Further interpretation will likely require accurate field mapping, which appears to be warranted based on the potential in this area for undiscovered carbonate aquifers.

CONCEPTUAL MODEL OF SELECTED METABASALTS IN THE PIEDMONT

The following is an attempt to present a conceptual model of the metabasalts of southeastern Pennsylvania. The reader is cautioned that portions of this section are highly conjectural.

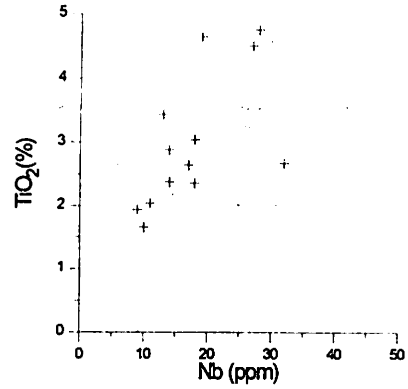


Figure 20. Plot of TiO_2 vs. Nb for White Clay Creek samples. Note that the slope is steeper than that for the Accomac-area metabasalt of the Catoctin event (Figure 13).

The principal visible indications of the latest Proterozoic-Z Catoctin event are the metabasalt and metarhyolite of South Mountain in Adams and Franklin Counties, Pennsylvania. This is the Catoctin Metabasalt *sensu stricto* that appears to be initial-rifting tholeiite. The Catoctin event, however, affected Proterozoic terranes. An extrusive rock that is a direct correlative is the Accomac-area metabasalt, which seems to be a good geochemical synopsis of the Catoctin Metabasalt *sensu stricto.* The Accomac-area metabasalt, in turn, has direct intrusive geochemical correlatives in the metadiabase dikes of the Reading Prong, Womeldorf outlier, Honey Brook Upland and Trenton Prong. Garnet-bearing metadiabase dikes of the Brandywine massifs also appear to be Proterozoic-Z in age but do not seem to be closely related in a geochemical sense.

In our model, the Pigeon Hills-area metabasalt represents the next stage, early Iapetan seafloor development. This stage may also be represented by a few closely related samples in the Catoctin Metabasalt *sensu stricto*. Specifically, these are from an area in the portion of South Mountain SE of the Tunnel Hill-Jacks Mountain fault system. Metabasalts collected at Holtwood and along Bald Eagle Creek appear to be transitions, less and more alkali, respectively, from the rifting stage of the Catoctin event, represented by the Accomac-area metabasalts, to the drifting stage, represented by the Pigeon Hills-area rocks and the northern subgroup of the Sams Creek Metabasalt. These are probably the last voluminous magmas related to subLaurentian mantle until the early Mesozoic.

The type Sams Creek Metabasalt and Kennett Square Amphibolite represent two sets of Iapetan (?) seafloor-generated N- and P=E-type basalts from spreading centers. They seem to represent the main stage of development of Iapetan sea-floor generation and apparently lack any inherited geochemical memory of Catoctin rifting. Samples DILW and BRANDYB from just north of the Kennett Square Amphibolite might represent erosional remnants of a different Iapetan(?) P=E-OFB related to the Wilmington Complex that was thrust over the Brandywine massifs. The "White Clay Creek Amphibolites" of Chester County remain enigmatic. They could be from ocean islands developed within the Iapetan plate.

Caught up in Iapetan closure are the Baltimore Mafic Complex (subarc?) and its 512 3-Ma kin, the James Run island-arc or back-arc volcanics; the Port Deposit Tonalite; the "Conowingo Creek metabasalts," some of which appear to be boninitic (forearc?); and the ultramafic mélange-associated Bald Friar Metabasalt from the back arc. These all appear to be remnants of an island-arc ophiolite complex. This and the associated mélange that contains fragments of the Bald Friar Metabasalt are likely to have become, in effect, the Taconian suture, i.e., a continuation of the northern Appalachian Baie Verte-Brompton belt between the oceanic Dunnage and the continental Humber terranes. In Pennsylvania the boundary may be a belt of ophiolitic mélange that has undergone later low-angle thrusting and folding (Smith 1993). However, in Pennsylvania the issue of the continental-oceanic boundary is likely to be a somewhat moot point, both because rocks of the mélange are exposed more than once by thrusting and folding and because some of the oceanic fragments have, by a combination of tectonic (mélange) and sedimentary (olistostrome) processes, been deposited in a matrix likely to be at least partially derived from the continental Laurentian or Brandywine massifs. In such a setting, the imagination is not stirred as it would be in the presence of a narrow, high-angle terrane boundary between purely continental and purely oceanic rocks. Perhaps the zone of the Fishing Creek Metabasalt, a probable P=E-OFB, and talc (Smith, 1993), as the closest known oceanic fragments to the axis of the Tuquan anticline, will have to suffice as a symbolic surface exposure of the Taconian suture. More likely, the suture may be a triparte belt of the type proposed by Rast and Horton (1989) for the northern Appalachians, in which the

Fishing Creek Metabasalt plus talc represent the inboard ophiolite mélange, the dismembered basalt-bearing part of the BMC the ophiolite, and the Sykesville Formation the outboard olistostrome.

Despite the above uncertainties, the uniformity of the Bald Friar Metabasalt fragments within the mélange indicates that it is the same mafic and ultramafic-bearing mélange that occurs immediately north of the Peach Bottom Slate along the NE shore of the Susquehanna River at Stop 14; south of the Peters Creek Formation at Bald Friar, Maryland; at Stop 7 in Chester County; and elsewhere. It is also likely that this same mélange is represented by the small talc-magnesite body just SE of the Cardiff Conglomerate on the NE shore of the Susquehanna River, but diagnostic metabasalts have not been observed.

The early Mesozoic diabases, like those of the Catoctin event, originated in sub-Laurentian mantle and may reflect mantle depletion that occurred during the Catoctin.

ACKNOWLEDGEMENTS

Samuel W. Berkheiser, Jr., assisted in the field sampling during the first half of the project. C. Gil Wiswall of West Chester University brought the Brandywine locality of the Bald Friar Metabasalt to the attention of the Pennsylvania Geological Survey. Rodger T. Faill found the York County locality of the Bald Friar Metabasalt. Localities for the Bald Friar type section and some of the "Conowingo Creek metabasalt" outcrops were found by use of the maps, notes, and collections of the late Davis M. Lapham. The Beaumont School outcrop of "older diabase" was found by Mary Emma Wagner of the University of Pennsylvania and called to our attention by David Valentino of Pecten International Company.

Reviews of the manuscript were provided by Raymond A. Coish of Middlebury College, Avery Ala Drake, Jr., of the U.S. Geological Survey, Arthur Montgomery, professor *emeritus* of Lafayette College, Arthur W. Rose of the Pennsylvania State University and A. Krishna Sinha of Virginia Polytechnic Institute and State University.

Finally, the cooperation of Dr. Eric L. Hoffman of Activation Laboratories, Ltd., which provided the analyses for this project, is greatly appreciated.

THE CREAM VALLEY FAULT: TRANSFORMATION FROM THRUST TO STRIKE-SLIP DISPLACEMENT

Rodger T. Faill[1] and C. Gil Wiswall[2]
(1) Pennsylvania Geological Survey
P. O. Box 8453, Harrisburg, PA 17105-8453
(2) West Chester University
West Chester, PA 19383

INTRODUCTION

The Cream Valley fault is an important geologic feature in the Piedmont of southeastern Pennsylvania. It separates the Taconian metamorphic core consisting of Mesoproterozoic Grenvillian gneisses of the West Chester massif and amphibolite facies metasediments ("Glenarm Wissahickon") on the south from shelf/rise metasediments (Octoraro) of Late Neoproterozoic(?) and Early Paleozoic age on the north (Figure 21). To the eastnortheast, it becomes the Huntingdon Valley fault separating the Wissahickon schists of the Philadelphia terrane on the south from the carbonates and siliciclastics of White Marsh Valley and the Trenton Prong gneisses on the north. To the west-southwest, the Cream Valley fault loses definition within the "Glenarm Wissahickon" phyllites and schists in the vicinity of the Woodville massif.

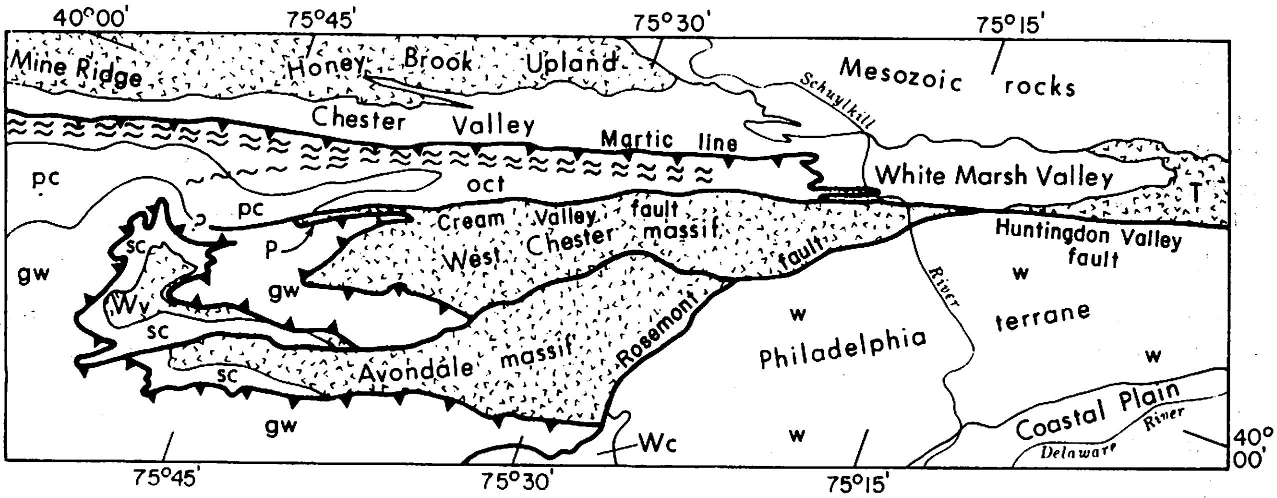


Figure 21. Geologic map of the Cream Valley fault and surrounding area (modified from Berg and others, 1980). P - Poorhouse prong; T - Trenton Prong; Wc - Wilmington complex; Wv - Woodville massif; gw - "Glenarm Wissahickon" of the White Clay nappe; oct - Octoraro Formation; pc - Peters Creek Formation; sc - Setters and Cockeysville Formations, undivided; w - Wissahickon Formation of the Philadelphia terrane.

Fabrics along the fault trace vary in character and record a complex movement history that implies multiple tectonic events. West-southwest of the Poorhouse massif where gneisses are absent, the fault trace corresponds to a zone of shallowly dipping, mutually cross-cutting foliations produced by shear. The older shear foliation crops out over a cross-strike distance of several hundred meters whereas the younger foliation is more-or-less restricted to the mapped fault trace. To the northeast where gneisses are present, the fault trace lies at the northwest edge of the Grenville-age gneisses of the West Chester massif, adjacent to a zone of steep foliation within the gneisses. Phyllites with multiple shear foliations and ultramafic bodies occur north of the fault trace in this area.

Over its entire length, the Cream Valley fault separates rocks metamorphosed to middle to upper amphibolite facies on the south from rocks in the middle to upper greenschist facies on the north. Several lines of evidence suggest that peak metamorphism and the initiation of faulting were synchronous (Wiswall, 1991), although most preserved structures formed after the metamorphic peak. Thus, initial faulting occurred at depth under amphibolite facies conditions. Retrograde mineral assemblages present along the fault trace indicate that subsequent movement occurred under greenschist facies conditions in the presence of fluids.

It is our thesis that the Cream Valley fault has a two stage history. The fault was initiated during the Taconian orogeny as a major throughgoing crustal thrust fault, herein named the Embreeville thrust. Motion on the Embreeville thrust raised the Grenvillian-age gneisses and cover northward over the siliciclastic metasediments of the continental slope, themselves having been already thrust over the carbonates of the upper continental slope and shelf by movement on the Martic thrust. The second phase involved late Alleghanian(?) transpression that produced folding and strike-parallel dextral displacement. It is this structure that was originally recognized as the Cream Valley fault. Dextral shear on the Cream Valley/Huntingdon Valley fault and other structures (Valentino and others, 1994) has shifted the Philadelphia terrane, the Brandywine terrane (composed of the West Chester, Avondale, and Woodville massifs and their cover sediments), and other Iapetan elements south of the fault, southwestward by an as yet undetermined distance relative to the Laurentian margin rocks on the north.

SYNOPSIS OF THE GEOLOGIC HISTORY OF THE CREAM VALLEY FAULT

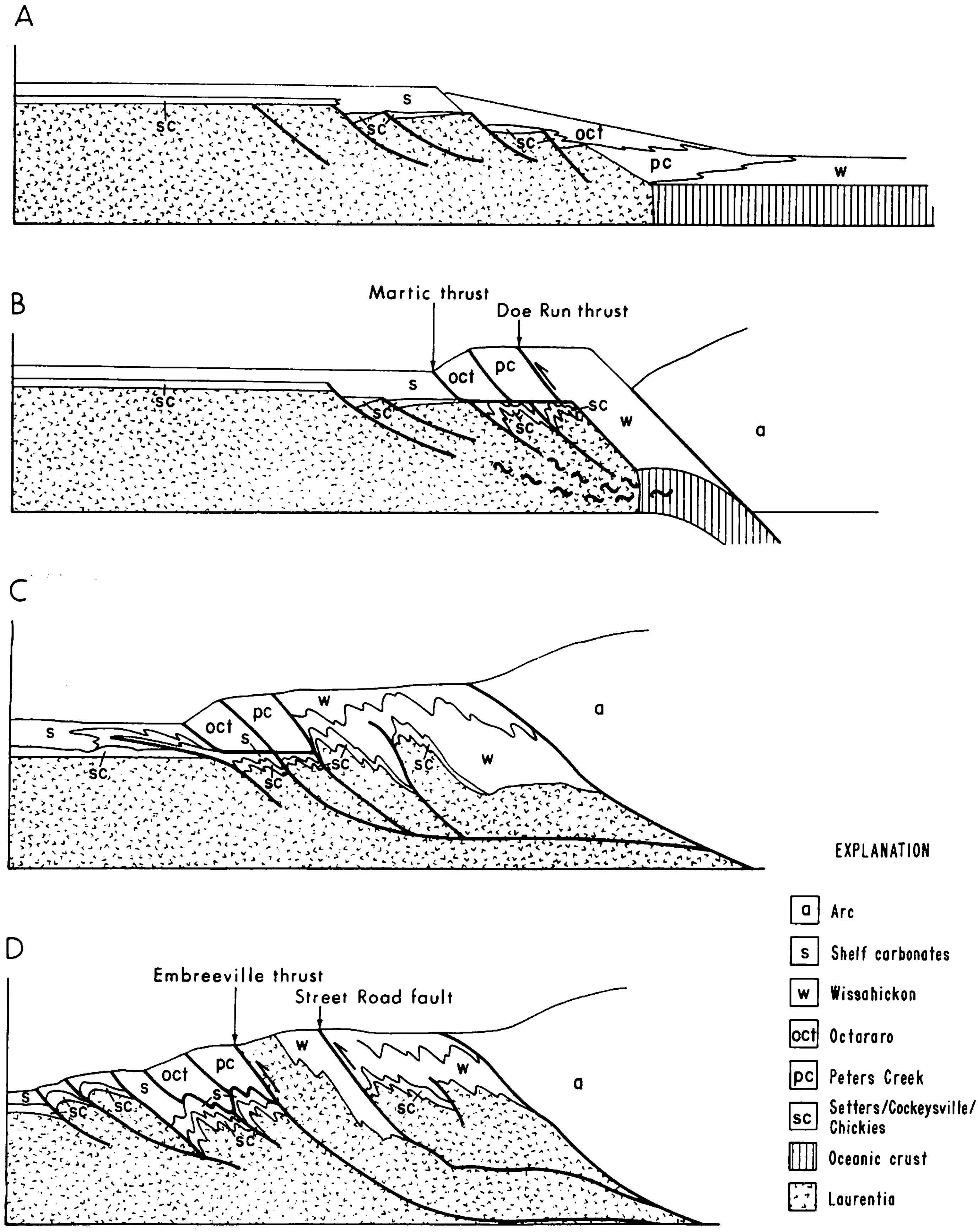
Early geologic setting

There is uncertainty surrounding the tectonic setting of the Grenville-age gneisses prior to the earliest stages of convergence. The gneisses either remained part of the Laurentian continent throughout the Neoproterozoic, or were microcontinental fragments within Iapetus prior to their collision with Laurentia. If the gneisses remained part of Laurentia, the Octoraro phyllite and Peters Creek schist represent slope/rise facies deposits (Figure 22A). On the other hand, if the gneisses represent a microcontinent, then these metasediments represent basin deposits of the Octoraro seaway (Figure 23A). Regardless, the history of the Embreeville and Cream Valley faults is largely independent of which view is preferred.

It seems clear that the Iapetan Ocean grew and widened during the Early Cambrian (and possibly latest Neoproterozoic) following the breakup of the Neoproterozoic supercontinent that included Laurentia and the Gondwana continents. Iapetus formed between the southwestern part of South America and eastern Laurentia (Bond and others, 1984; Dalla Salda and others, 1992). A passive continental margin developed on the eastern margin of Laurentia, on which a broad carbonate shelf gradually spread westward from the eastern margin for hundreds of kilometers over the craton. Eastward from the shelf edge, a carbonate/ siliciclastic continental slope descended and merged with an oceanic basin floor (Wagner and others, 1991).

Outboard of the shelf, three lithologies accumulated in either the Octoraro seaway or Iapetus (depending of which model is adopted). In their present distribution from north to south, the formations include the Octoraro phyllite, the Peters Creek schist, and the "Glenarm Wissahickon" schists (the last consisting of those rocks south of the Cream Valley fault and west of the Rosemont fault that were mapped in the past as Wissahickon, e.g., Berg and others, 1980). Telescoping during Taconian tectonism renders the original stratigraphic and geographic distribution of these rocks uncertain. The Octoraro phyllite was a pelitic deposit of probable great (>5 ? km) thickness. The Peters Creek schist was a pelitic/psammitic deposit that has been interpreted as a turbidite sequence (in part) with continental crust as a contributing source of sediment (Gates and Valentino, 1991). The presence of oceanic floor basalts in what has been mapped as the Peters Creek sediments (Smith, 1994, this volume), and of tectonically emplaced ultramafic bodies, indicates that the underlying crust had to be

Figure 22. **(A)** The Laurentian eastern margin in the Early Cambrian, with the Octoraro phyllite and Peter Creek schist deposited on the continental slope and rise. **(B)** Convergence of magmatic arc on the eastern Laurentian margin: obduction of the coalesced island arc (Wilmington complex) and accretionary prism (White Clay nappe) onto the continental slope. Movement on Martic and Doe Run thrusts. **(C)** Nappe formation in the Laurentian margin and early northwestward movement on the Embreeville thrust after peak Taconian metamorphism. **(D)** Continued northwestward movement on the Embreeville thrust of the coalesced Brandywine/White Clay/Wilmington complex onto the Octoraro/Peters Creek sediments.



A
s
sc
oct
pc
w
B
Martic thrust
Doe Run thrust
s
oct
pc
sc
w
a
C
s
oct
pc
w
sc
a
D
Embreeville thrust
Street Road fault
s
oct
pc
w
sc
a
EXPLANATION
a Arc
s Shelf carbonates
w Wissahickon
oct Octararo
pc Peters Creek
sc Setters/Cockeysville/Chickies
Oceanic crust
Laurentia

oceanic, at least in part. The "Glenarm Wissahickon" schists were originally *siliciclastic* sediments from Iapetus that may have included turbidites and intercalated basalts. These metasediments now comprise the White Clay nappe (herein informally named) which was thrust over the Brandywine microcontinent). Taconian metamorphism and deformation transformed these original deposits into their present state.

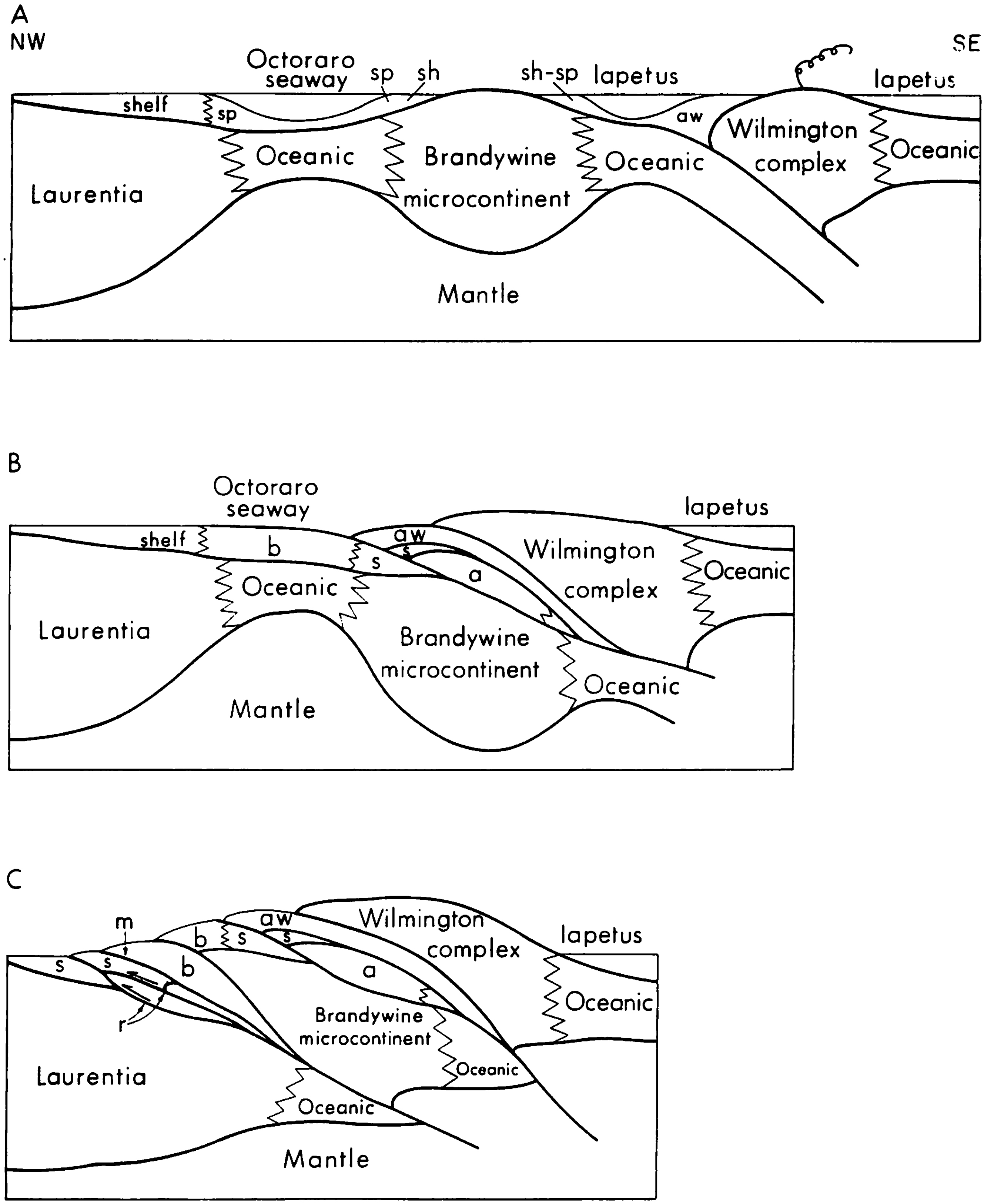
Convergence – the Embreeville thrust stage of the Cream Valley fault

The Cream Valley fault's history commenced with the formation of the Embreeville thrust in response to the Early Paleozoic closing of Iapetus. The Wilmington complex (Wagner and Srogi, 1987) and the Baltimore mafic complex (Hanan and Sinha, 1989) have been interpreted as magmatic arcs. It is possible that the Wilmington complex represents the roots (Wagner and Srogi, 1987), and the Baltimore mafic complex shallower levels of the same structure. Based on the 502 Ma crystallization age of the Arden pluton in the Wilmington complex (Foland and Muessig, 1978), and the 510 to 515 dates (Drake, 1987; Sinha, 1988) of the various parts of the Baltimore mafic complex, Iapetus had begun closing by the Middle Cambrian, if not earlier.

The beginning of convergence initiated a subduction zone over which an magmatic arc developed. From the apparent relative paucity of volcanic deposits in the Octoraro/Peters Creek basin sediments, and even in the "Glenarm Wissahickon" (although the numerous amphibolites there the suggest a greater abundance of basaltic layers), one can infer that the subduction zone dip was to the east. In this model, the White Clay nappe formed out of an accretionary prism that grew in the trench environment west of the advancing island arc (Figure 3a). An eastward dipping subduction zone under a westward advancing island arc is central to the model of Wagner and Srogi (1987) as well. In contrast, Hanan and Sinha (1989) argue that the ensialic character of the Baltimore mafic complex gabbros imply an island arc associated with continental crust. They propose that the Baltimore mafic complex formed in a backarc basin over a west-dipping subduction zone. However, they allow that the gabbros could be intrusive into an accretionary prism.

Regardless of direction of subduction, convergence during the Taconian telescoped the various lithotectonic components onto the Laurentian margin (Figures 22B-D or 23B-C). At least five thrust/nappe packages were formed. From north to south, these include: (1) the Octoraro/Peters Creek metasediments bounded by the Martic thrust below and Embreeville thrust above; (2) the West Chester/Woodville nappe(s) (Wagner and others, 1993) bounded by the Embreeville thrust below and Street Road and Doe Run faults above; (3) the Avondale/Mill Creek nappe(s) bounded by the Street Road fault below and Doe Run fault above; (4) the White Clay

Figure 23. **(A)** The Laurentian eastern margin in the Middle Cambrian, with Laurentia and the Brandywine microcontinent separated by the Octoraro Seaway. The Wilmington complex was part of a magmatic arc in Iapetus that developed over a subduction zone dipping away from Laurentia. Terrigenous/carbonate shelf deposits accumulated on the margins of Laurentia and the Brandywine microcontinent. An accretionary wedge grew over the subduction zone in front of the advancing Wilmington complex. Cover sequences: sf - shelf; sp - slope; s - shelf and slope, undifferentiated; aw - accretionary wedge. **(B)** The Laurentian eastern margin in the Middle(?) Ordovician. The Wilmington complex and associated accretionary wedge has been obducted over the Brandywine microcontinent, causing it to descend to great depth. a - Avondale massif, a fragment of the Brandywine microcontinent. Sedimentary sequences: s - shelf and slope deposits, undifferentiated; b - basinal deposits; aw - accretionary wedge and Iapetan sediments. **(C)** The Laurentian eastern margin in the Late Ordovician, at the end of the Taconian orogeny. Continued convergence has obducted the Brandywine microcontinent onto the Laurentian margin, collapsing the Octoraro seaway. m - Martic line, the contact (thrust fault) between the Octoraro basinal deposits and the Laurentian shelf; r - thrust faults of the Reading meganappe system. Fragments of the Grenville crust have been broken from the Laurentian margin and incorporated in the Reading nappes. Other symbols as in Figure 23A.


A
NW
SE
Octoraro seaway
sp
sh
sh-sp
Iapetus
Iapetus
shelf
sp
aw
Wilmington complex
Oceanic
Oceanic
Brandywine microcontinent
Oceanic
Laurentia
Mantle
B
Octoraro seaway
Iapetus
shelf
b
aw
s
s
a
Wilmington complex
Oceanic
Oceanic
Laurentia
Brandywine microcontinent
Oceanic
Mantle
C
m
aw
Wilmington complex
b
s
s
Iapetus
s
s
b
a
complex
r
Brandywine microcontinent
Oceanic
Oceanic
Laurentia
Oceanic
Mantle

nappe, riding on the Doe Run fault (Alcock, 1994); and (5) the Wilmington complex. The relative timing of the various faults is largely uncertain. The Martic thrust formed early, before the metamorphism of the enclosing sediments (Wise, 1970). The Street Road fault cuts the Doe Run thrust and therefore formed later. Other age relationships are not known, but must have occurred in two major phases. These phases are the assembly of the Iapetan components, and the telescoping of the Laurentian margin. We do not mean to imply that the phases were sequential; in fact, they may overlap in time.

The first phase involved thrusting of the magmatic arc (of which the Wilmington complex was the hot base) westward into and over the forearc basin containing the "Glenarm Wissahickon" accretionary prism sediments. These sediments, being deeply buried within the prism, were already at an elevated temperature. The emplacement of the lower part (near the magmatic core) of the arc raised the temperature further, into the second sillimanite zone of the amphibolite facies (Wagner and Srogi, 1987). This combined package cooled somewhat before it was thrust over the Brandywine gneisses on the Doe Run fault (Alcock, 1994). The added load from this obduction caused the gneisses to descend (Figure 22C or 23B), perhaps as deep as 30 km, judging from the Taconian metamorphic overprint in the West Chester massif (Wagner and Crawford, 1975). It was probably at this stage, while the gneisses were deeply buried, that the regional Taconian metamorphism (as opposed to that associated with the obduction of the Wilmington complex) reach its peak.

The second phase resulted in upward transport and emplacement of the basement nappes onto the Octoraro/Peters Creek sediments along the Embreeville thrust (Figures 22C-D and 23C). At some point during this phase, the Octoraro sediments were emplaced over the carbonate shelf edge by movement on the Martic thrust. This event may have preceded movement on the Embreeville thrust by a significant amount of time (Figure 22B). The absence of continental or oceanic crustal rocks in the hanging wall indicates that the Martic thrust originated within the Octoraro/Peters Creek basin sediments. Similar thrusts below and/or northwest of the Martic broke through the upper Laurentian continental crust and overlying carbonate shelf, producing several thrust nappes that formed the Reading Prong meganappe system (Figure 23C). Continued convergence resulted in the formation of the Street Road fault late in this phase.

If the Grenville-age gneisses of the Brandywine terrane were from a separate microcontinent, then movement on the Embreeville thrust resulted in the collapse of the intervening Octoraro seaway. This could explain the source of several ultramafic bodies emplaced along the fault. If, on the other hand, the gneisses were always a part of Laurentia, the Embreeville thrust must extend to the base of the continental crust. In this latter case, the ultramafic bodies represent slices of subcrustal mantle.

The trace of the Embreeville thrust to the southwest of the Woodville massif has not been identified; it may follow the present-day contact between the "Glenarm Wissahickon" schists and the Peters Creek Formation. The amount of displacement is uncertain but is probably significant given the juxtaposition of different metamorphic grades and the presence of ultramafics along the fault to the northeast. The loss of fault definition and lack of gneisses to the southwest may suggest that displacement progressively decreases in that direction. To the northeast, along the Huntingdon Valley fault, it is not known if the displacement on the Embreeville thrust was greater than or comparable to that adjacent to the West Chester nappe. If the Philadelphia terrane lies at a higher structural level (assuming that it had been thrust over the Brandywine terrane), then the amount of displacement on the Embreeville may have decreased to the northeast.

The juxtaposition of the higher grade rocks against the lower grade sediments steepened the metamorphic gradient, leading to the present-day compressed isograds along the north side of the Cream Valley fault (Figure 24; see also Valentino, this guidebook). The decrease in grade in the "Glenarm Wissahickon" rocks southwest of the Woodville and Avondale massifs possibly reflects either (1) the structural climb (because of the southwest structural plunge) through the accretionary prism toward hypabyssal levels, where the metamorphism was at a lower grade, or (2) greater distance from the heating effects of the obducted Wilmington complex. Indeed, both factors may have been in effect.

Strike-slip displacement–the second stage of the Cream Valley fault

The history of this area for the next 180 m.y. following the Taconian orogeny is somewhat

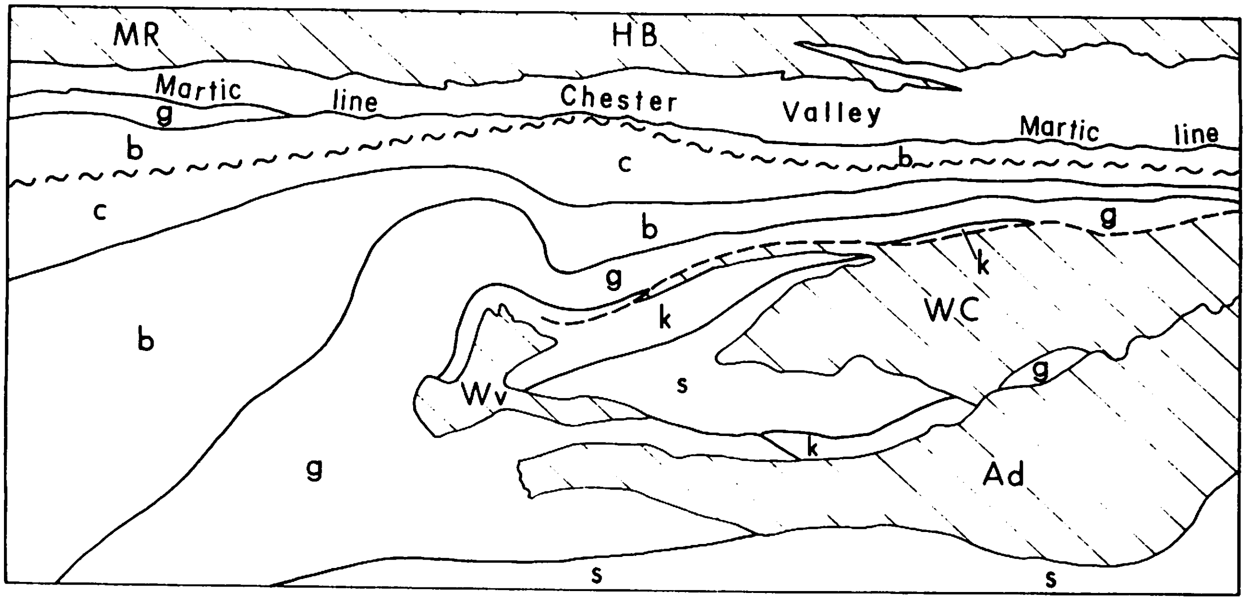


Figure 24. Metamorphic zone map for post-Grenville rocks in the Pennsylvania Piedmont (from Valentino and Faill, 1990). Greenschist facies: c - chlorite zone; b - biotite zone; g - garnet zone. Amphibolite facies: k - kyanite zone; s - sillimanite zone. Heavy dashed line - Cream Valley fault. Diagonal lines - Grenvillian rocks. Squiggle line - northern edge of the Pleasant Grove-Huntingdon Valley shear zone. Ad - Avondale massif; HB - Honey Brook Upland; MR - Mine Ridge; WC - West Chester massif; Wv - Woodville massif.

obscure. The rocks remained quite warm until well into the latter part of the Paleozoic, as evidenced by the Mississippian-age cooling dates (see Lapham and Root, 1971, and Faill, in press, for summaries of dates). The wide range in dates (420 to 305 Ma) from the Susquehanna to the Delaware Rivers suggests uplift was gradual, with no significant tectonic activity, not even any related to the Acadian orogeny of the northern Appalachians (Faill, 1985). It wasn't until the Alleghanian orogeny was underway that the transformation of the Embreeville thrust commenced.

The Alleghanian orogeny in Pennsylvania is best known for the arcuate large, long folds in the foreland that resulted from the northwestward shortening of the Paleozoic cover. The southeasternmost, presently-known structure reflecting this movement is the Oregon thrust, which passes through Lancaster Valley, and on which the Blue Ridge province (which includes the Pigeon and Hellam Hills, the Honey Brook Upland, and the intervening carbonates) overrode the carbonate shelf. No similar southeast-dipping Alleghanian thrusts have been described in the vicinity of the Cream Valley fault.

The principal Alleghanian tectonism that has been recently identified in this part of the Piedmont is a northeast-trending dextral transpression (Hill, 1987; Howard, 1988; Valentino, 1990; Valentino and Wiswall, 1991; and Wiswall, 1991). This tectonism is expressed by folding, cleavage development, and in subvertical regional shear zones that include the Cream Valley and Huntingdon Valley faults.

The Cream Valley fault is apparently but one segment of a regional shear zone that has been traced from where the Pleasant Grove shear zone emerges from under the Culpeper basin in Maryland to the end of the Huntingdon Valley fault at the Delaware River (Valentino and others, 1994). Based on foliation orientation, the Cream Valley fault appears to be a subvertical zone along most of its length, where the West Chester nappe and Poorhouse Prong gneisses lie on its southeast side. The similarity in sequence and style of structures within the fault zone with those outside it suggest that the fault zone foliation is the same as that seen elsewhere in the gneisses. If this is the case, then the foliation has been reoriented to a near vertical orientation by an early flattening deformation prior to dextral shearing. West of the Poorhouse Prong, this vertical fabric of the fault gradually dies out, leaving only the distinct metamorphic and structural break established by the Embreeville thrust within the enclosing siliciclastic rocks. In the vicinity of the Woodville nappe, the

Embreeville thrust swings to the northwest around the Woodville dome suggesting that it has been folded. This change in orientation corresponds with a set of localized, late (F3) folds with north plunging axes that occur only around the northern portion of the Woodville dome.

Several zones containing fabrics that record strike parallel, dextral shear have been identified between the Martic thrust and the Street Road fault. Extending northeastward from the Susquehanna River, both the Drumore and Peach Bottom shear zones consist of a steep phyllonite zone characterized by extreme grain size reduction and asymmetric porphyroblasts and pressure shadows indicating dextral displacement (Valentino, 1990; 1993; see also Valentino, this guidebook). The subvertical foliation along which slip occurred is superimposed on Taconian fabrics and is axial planar to tight, upright folds. To the northeast, these zones join and narrow; farther northeast, the combined zone appears to splay or bifurcate before dying out in the South Valley Hills southeast of Downingtown. Along the Cream Valley fault, the character and intensity of fabric development varies along strike. To the northeast where the gneissic foliation has been reoriented, the fabric associated with Cream Valley fault movement is primarily a subvertical mylonitic foliation (Howard, 1988). Meso- and microscopic kinematic indicators are rare, but macroscopic deflection of foliation within the fault zone indicates dextral displacement (Howard, 1988). To the southwest where the shallow orientation of the Embreeville thrust is preserved, a nonpenetrative crenulation cleavage showing Type II S-C relationships indicates oblique slip with both normal and dextral components. These features suggest that the magnitude of dextral displacement decreases to the southwest. Along the Street Road fault, Type I S-C fabrics indicate subhorizontal, dextral displacement. The extent of these fabrics is not presently known so the possible relationship with zones to the north is uncertain.

Thus, Alleghanian tectonism in the Piedmont is characterized by early folding and cleavage development followed by later dextral strike slip, the distribution of which was controlled by the Taconian architecture. The earlier structures formed in response to a strong subhorizontal flattening strain perpendicular to the tectonic trend. Within the Octoraro phyllite and shelf carbonates of the Chester Valley, mesoscopic folding resulted in tight, upright folds and a vertical axial planar schistosity. The Brandywine terrane cored by the West Chester and Avondale nappes were themselves folded into macroscopic, southwest plunging antiforms. Presumably, it was during this phase that the foliation in the West Chester gneisses was reoriented along the Cream Valley fault.

Subsequently, strike slip motion parallel to the tectonic trend replaced the earlier flattening. The fact that the combined Drumore/Peach Bottom zone apparently dies out to the northeast while transcurrent motion dies out along the Cream Valley fault southwestward suggests that the area between the two represents a transfer zone (Figure 25). In this model, dextral displacement on the Cream Valley/Huntingdon Valley fault farther east was transferred en echelon to the Drumore/Peach Bottom shear zone to the north. The late localized folds around the northern Woodville massif may represent the deformational response as displacement along the Cream Valley fault dropped to zero.

DISCUSSION – Age and significance of the Cream Valley fault

The age of movement on the Cream Valley fault is constrained within a rather generous time interval. The crosscutting fabric relationships and lower greenschist facies retrograde metamorphism associated with the Cream Valley fault indicates that the movement on it was late- to post-Taconian. The overlap of the Pleasant Grove shear zone by the Carnian Manassas sediments of the Culpeper basin in northern Virginia demonstrates a pre-Late Triassic movement. It is unlikely that a structure of this size was active during non-orogenic time, so its activity is probably restricted to one or more of the three major orogenies that occurred in the central Appalachian orogen with in this time span: the Taconian, Acadian, and Alleghanian orogenies.

Existing radiometric age data may indicate additional constraints on the timing of fault movement. The lower greenschist facies retrograde metamorphism associated with the fault could possibly have occurred in late in the Taconian orogeny (450 to 440 Ma) if the rocks involved cooled rapidly enough. However, radiometric age determinations in rocks close to the fault indicate that temperatures did not reach lower greenschist facies levels (300-350 degrees C) until late in the Devonian or early in the Carboniferous. A K-Ar date of 353 Ma

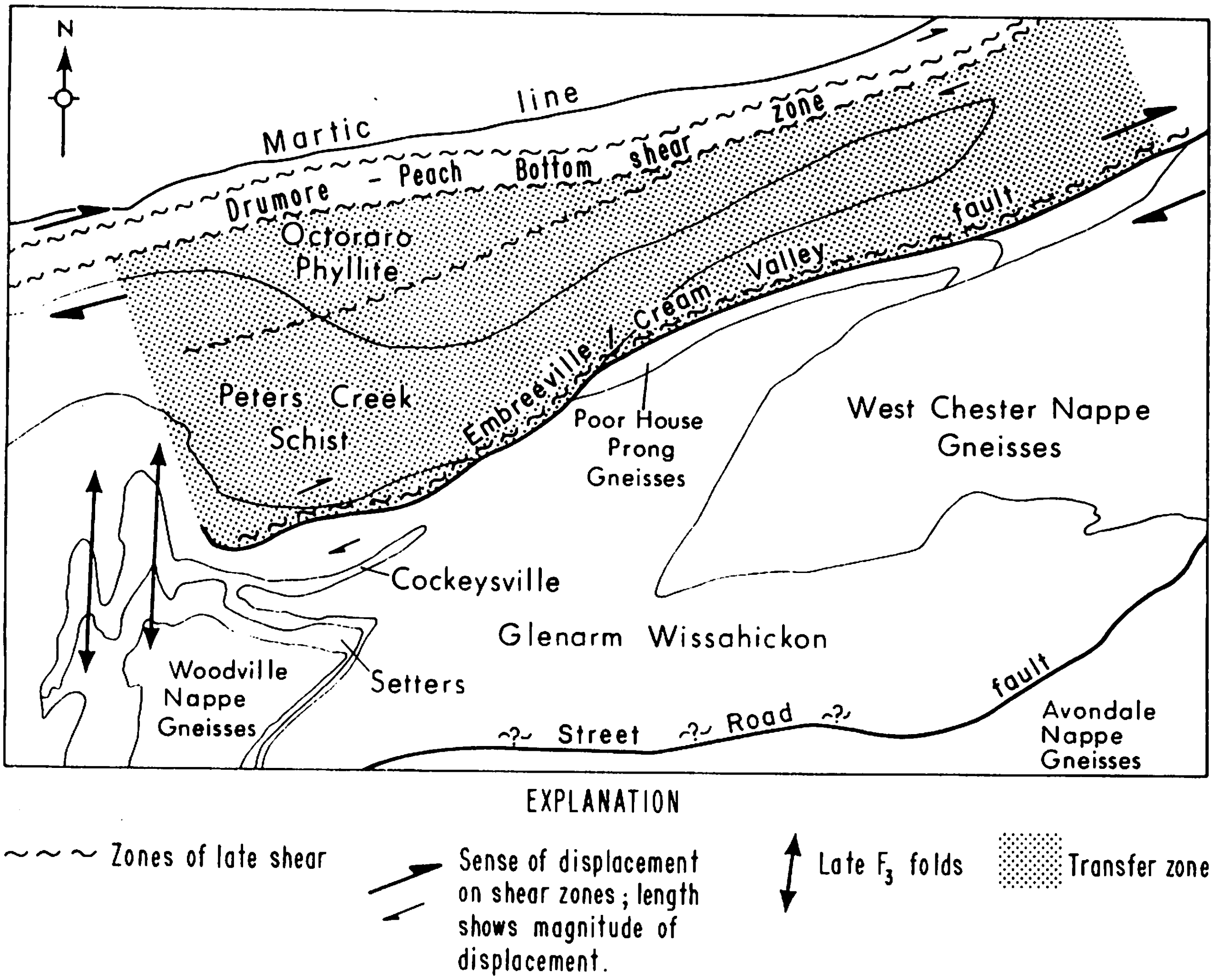


Figure 25. North-south trending folds in the transfer zone between the westward diminishing Cream Valley fault and the eastward diminishing Drumore-Peach Bottom shear zone.

was obtained on muscovite for the Wissahickon schist close to the Huntingdon Valley fault near Neshaminy Creek (Long and Kulp, 1962). 40Ar/39Ar dates of 336 Ma (on biotite) and 342 Ma (on muscovite) were determined for Mesoproterozoic gneiss near the Cream Valley fault in the eastern end of the West Chester massif (Sutter and others, 1990). A number of other dates farther from the Cream Valley fault and the shear zones similarly fall within the 360 to 305 Ma range (Lapham and Root, 1971), suggesting that the temperatures in this part of the Piedmont did not fall to lower greenschist facies levels until well after the Taconian orogeny.

The apparent absence of Acadian tectonism (380 to 360 Ma) in Pennsylvania (Faill, 1985), and the large proportion of Carboniferous radiometric dates (as described in the preceding paragraph), suggests that the Cream Valley fault was not active during the Acadian orogeny either. This leaves the Alleghanian orogeny as the time of activity.

We hasten to note that this analysis is not conclusive. Both K/Ar and 40Ar/39Ar determinations record only the latest cooling event below the blocking temperature for argon. This leaves open the possibility that fault movement did occur late in the Taconic, but the rocks remained above the argon blocking temperature until the Carboniferous. Further, the association of greenschist retrograde mineral assemblages associated with fault fabrics would suggest that the rocks were at least close to the argon blocking temperature during fault movement. Even if the rocks remained below the blocking temperature, deformation should result in argon loss, thereby yielding dates that are too young. Nevertheless, based on the recognition of major strike-slip motion during the Alleghanian in other parts of the orogen (e.g., Secor and others, 1986; Gates and others, 1986; Lefort and Van der Voo, *1981*; Goldstein, 1994), we consider Alleghanian movement most likely.

If the transpressional tectonics of the Cream Valley fault and associated structures is Alleghanian in age, then this fault should relate geometrically, kinematically, and dynamically to known Alleghanian structures in the central Appalachian foreland. Known Alleghanian deformation in the central Appalachians was entirely a décollement tectonism. Paleozoic and some Proterozoic rocks in a vast hanging-wall block were carried northwestward on a deep, subhorizontal décollement. One fact about the Alleghanian foreland is fundamental--the northern termination of the Valley and Ridge province of the Appalachian foreland, corresponding to a vast reduction in the amount of northwest shortening, lies in eastcentral Pennsylvania. The 200+ km horizontal shortening of the Paleozoic section from south-central Pennsylvania southward is an order of magnitude or more than the Alleghanian shortening in northeastern Pennsylvania and northern New Jersey. Clearly, a broad, major Alleghanian tectonic boundary lies between the central/southern Appalachians on the one hand, and eastern Pennsylvania/northern New Jersey on the other. This province boundary crosses central and southeastern Pennsylvania with an approximately northwest-southeast trend.

In the Valley and Ridge province, the boundary is represented by the northeastward diminution and termination of the first-order folds in east-central Pennsylvania, from the Nittany anticline across the anthracite basin to the Great Valley (Figure 26). No structures southeast of the Valley and Ridge have been recognized as corresponding to this boundary. The major Alleghanian structures southeast of the Valley and Ridge consist of low-angle thrusts, such as the Yellow Breeches, Grings Hill, and in the Piedmont, the Oregon thrust. All of these and similar thrusts post-date the foreland folding.

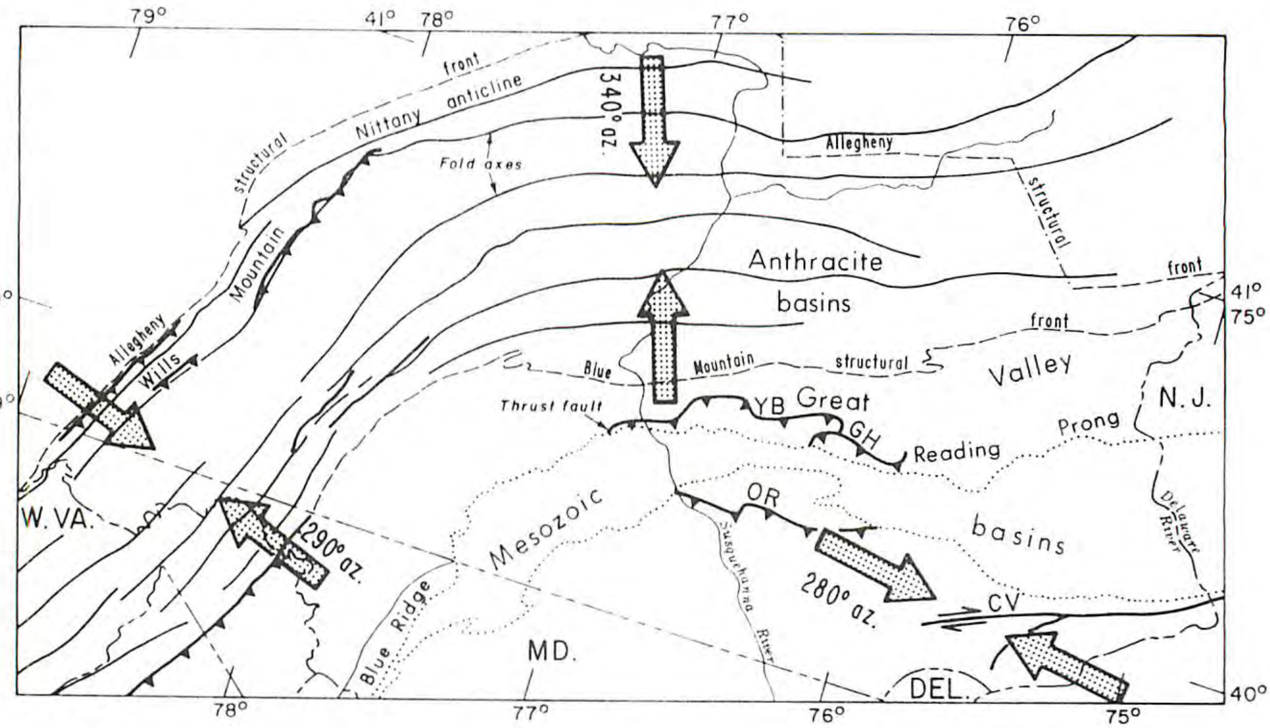


Figure 26. Large opposing arrows represent the inferred maximum principal stress directions for: the Cream Valley fault shear movement (280 degrees azimuth); Alleghanian foreland folds in south-central Pennsylvania (290 degrees azimuth); and Alleghanian foreland folds in central Pennsylvania (340 degrees azimuth). See text for discussion.

Geometrically and dynamically, the 070 degree azimuth trend of the Cream Valley fault is important if the fault was related to the foreland shortening. Dextral movement along such a trending fault/shear zone implies a driving stress oriented approximately 30 degrees clockwise from the fault, about 280 degrees azimuth (Figure 26). North of the Cream Valley fault, from Chester Valley and the Honey Brook Upland, across the Great Valley and Reading Prong, and over much of the Valley and Ridge province on either side of the Susquehanna River, the structural grain, represented mostly by upright folds, trends 070 degree azimuth. Although the folds parallel the Cream Valley fault, the fold-producing movements (at least north of the Mesozoic basins) were perpendicular to this trend, implying a maximum principal stress oriented 340 degrees azimuth, a 60 degree divergence from that inferred for the Cream Valley fault dextral movement. Therefore, it is unlikely that the strike-slip movement on the Cream Valley fault was in any related to the Alleghanian deformation to the north. However, the early flattening along the fault is geometrically, kinematically, and dynamically coincident with the Alleghanian folding. Thus, the dextral movement on the fault was later than the Alleghanian folding to the north.

It is only in southcentral Pennsylvania and west-central Maryland that the foreland and Blue Ridge trends swing around to 020 degrees azimuth, implying a maximum principal stress there oriented 290 degrees, close to that for the Cream Valley fault in Chester County. The correspondence of stress orientation between these two areas may indicate a genetic relationship, one in which the dextral movement on the Cream Valley fault is connected with additional late décollement movement in the southcentral Pennsylvania and west-central Maryland area, movement that has no counterpart in the shortening of the central and southeastern parts of Pennsylvania.

This is not to suggest that the first-order anticlines to the west all formed later than those to the north. The continuity of individual folds from the one area to the other suggests that each is a continuous structure in which the part in the west formed coevally with the corresponding part along trend to the north. In addition, the geometry and inferred kinematics of the anticlines indicate that they formed in reverse sequence, that is, from the Nittany/Wills Mountain anticline southeastward to the Blue Ridge anticlinorium (Faill, 1991). This reverse sequence in the foreland suggests that Alleghanian structure to the southeast in the Piedmont are even younger. If so, then the transpressional tectonics represented in part by the strike-slip movement on the Cream Valley fault may be very late Alleghanian in age.

The amount of strike-slip displacement on the Cream Valley fault is unknown, because the corresponding part of any feature on one side of the fault (or its extension, the Huntingdon Valley fault) has not been located on the other side. Hence, offsets cannot be measured. The suggestive "offset" between the Trenton Prong and the West Chester massif (see Figure 21) is spurious because the lithic differences between the Laurentian Trenton Prong gneisses and those of the non-Laurentian West Chester massif (e.g., Drake, 1984) preclude a pre-transpressive match-up. The possibility has been raised that the Brandywine gneisses (and the similar Baltimore gneisses to the southwest in Maryland) were attached to the Manhattan Prong throughout most of the Paleozoic, and were separated during the Alleghanian by dextral transpression (Valentino and others, 1994). However, the lithic dissimilarities between the Fordham gneiss (Hall, 1968) and the Brandywine massifs (e.g., Wagner and Crawford, 1975) argue against this hypothesis. In addition, the Fordham gneiss closely resembles the gneisses of the eastern Hudson Highlands (Ratcliffe and others, 1985). Measurement of the displacement is further complicated by the near coincidence of the Cream Valley strike-slip fault and the surface trace of the Embreeville thrust. Any features that were originally continuous across the incipient Embreeville thrust (and thus could be used to measure displacements) were offset by the thrusting by large amounts long before the transpression began, so there may be no features along the Cream Valley fault (and its extensions) that can be used to ascertain the strike-slip displacement.

Be that as it may, the amount of strike-slip displacement on the Cream Valley fault could not have been very large, certainly not tens of kilometers. The amount was probably no more than a few kilometers at most on its eastern end because the fault dies out some 50 km to the southwest, not far from the west end of the Poorhouse Prong. The displacement at its east end is a measure of the differential shortening of one side relative to the other. No structures have as yet been identified on either side that represents such shortening (or extension) parallel to the fault.

It is moot to what depth the Cream Valley fault extends. There is no direct evidence to indicate that it is a major crustal structure, extending through the continental crust. If the displacement on it is late Alleghanian, then it may possibly extend to the basal décollement, some 8 to 12(?) km below the surface. Considering its limited length and displacement, it probably extends only to one of the late splays from the décollement, probably the Oregon thrust, at a depth here of 3(?) km. If so, it represents only a significant tear in the hanging wall above the basal Alleghanian décollement or one of its splays.

THE PEACH BOTTOM PROBLEM IN LANCASTER COUNTY, PENNSYLVANIA

David W. Valentino
Concord College
Athens, WV 24712

ABSTRACT

The Peach Bottom syncline was defined by the apparent distribution of Cardiff conglomeratic quartzite around the northeastern and southwestern ends of the Peach Bottom slate belt in Lancaster County, Pennsylvania and Harford County, Maryland. It was suggested that the cleavage-bedding intersections and younging criteria, south of the slate belt indicate an anticlinal structure. The controversy over the Peach Bottom "fold" has existed for many decades because of the lack of hard information. This paper focuses on some new observations made during mapping along the length of the Peach Bottom structure in Lancaster County, Pennsylvania: (1) The new mapped distribution of the Cardiff Formation does not define a fold closure as portrayed by earlier workers; (2) At the Susquehanna River the Peach Bottom structure is a 1-1.5-km-wide zone of penetrative S3 schistosity that merges with the Pleasant Grove-Huntingdon Valley dextral shear zone toward the northeast. This suggests that the Peach Bottom structure is a large splay off the main shear zone; (3) Kinematic analyses in the Peach Bottom structure reveals both compressional and transcurrent deformation, and the transcurrent microstructures show a consistent dextral shear sense; and (4) The structural position and composition of the Peach Bottom and Cardiff Formations, and the regional stratigraphy of the Peters Creek Formation suggests that the Peach Bottom and Cardiff Formations are the structurally (stratigraphic?) lowest units in a monoclinal sequence containing rift-related clastics. This paper contains a summary of the work done on the Peach Bottom structure in Lancaster County exclusively. Future research will focus on the southwestern end of the structure in York County, PA and Harford County, MD.

INTRODUCTION

In southeastern Pennsylvania and northern Maryland the Peach Bottom slate occupies a narrow belt about 1.5 km wide and 25 km long (Figures 27 and 28) from Harford County, MD to southern Lancaster County, PA. For more than a century the Peach Bottom slate belt has been included in geologic maps and the focus of discussion and controversy in the geologic literature (Frazer, 1880; Lesley, 1885; Knopf and Jonas, 1929; Behre, 1933; Agron, 1950; Freedman and others, 1964; Wise, 1970; Smith, 1993). The slate belt was first reported in a volume written by Persifor Frazer in 1880. Nearly fifty years later the slate belt was included in a map and geologic report by Knopf and Jonas (1929), in which they portrayed the slate belt as residing in the core of a syncline, the *Peach Bottom syncline*.

The synclinal interpretation was based on the mapped distribution of Cardiff conglomeratic quartzite around the northeastern end of the Peach Bottom slate which seemed to define an apparent fold closure. The Peach Bottom slate and Cardiff quartzite were interpreted to be conformable with the surrounding Peters Creek Formation with the Peach Bottom slate interpreted as the youngest unit and the Peters Creek Formation interpreted as the older in the synclinal sequence.

Agron (1950) studied in detail the extensive exposures of the Peach Bottom slate along the strike of the belt and along the Susquehanna River where the belt is best exposed. Although Agron's structure and petrographic analysis was detailed, the only evidence supporting a syncline was the mapped distribution of the Cardiff conglomeratic quartzite around the ends of the slate belt. Agron's map was grossly compiled from the mapping of Knopf and Jonas (1929), Behre (1933), and Stose and Jonas (1939). Freedman and others (1964) reported on an investigation that dealt with the detailed geometry and relative timing of structures along the Susquehanna River and they interpreted the Peach Bottom structure to be a syncline (F2) that developed during the second phase of regional deformation (D2). Wise (1970) proposed a similar structural model to explain the position and geometry of the Peach Bottom syncline. Both Freedman and others (1964) and Wise (1970) did not deviate from the map patterns of earlier workers that portrayed a syncline.

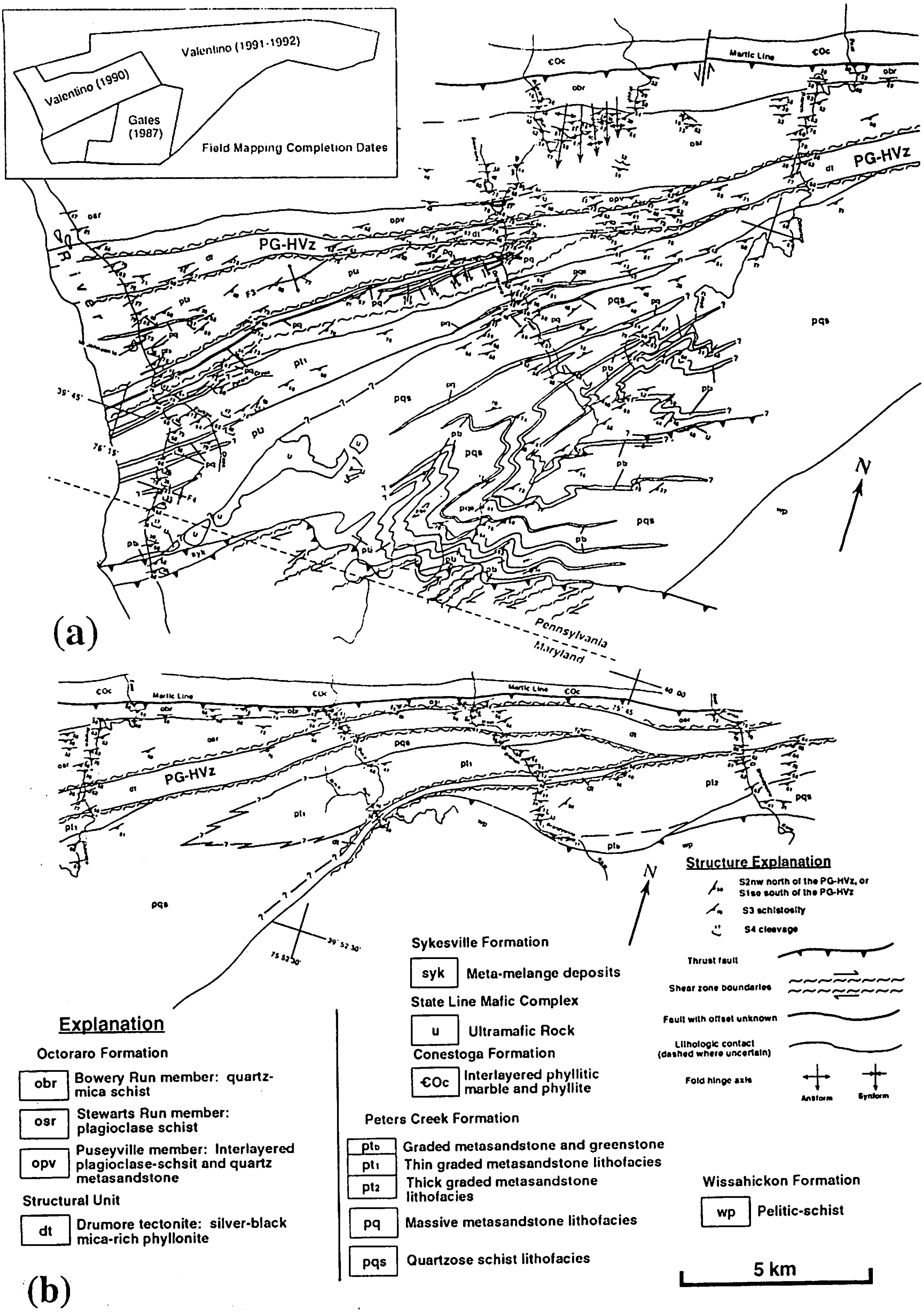


Figure 27. Regional map of the central Appalachian Piedmont (after Valentino and others, 1994). [A] Area west of Octoraro Creek. [B] Area east of Octoraro Creek.

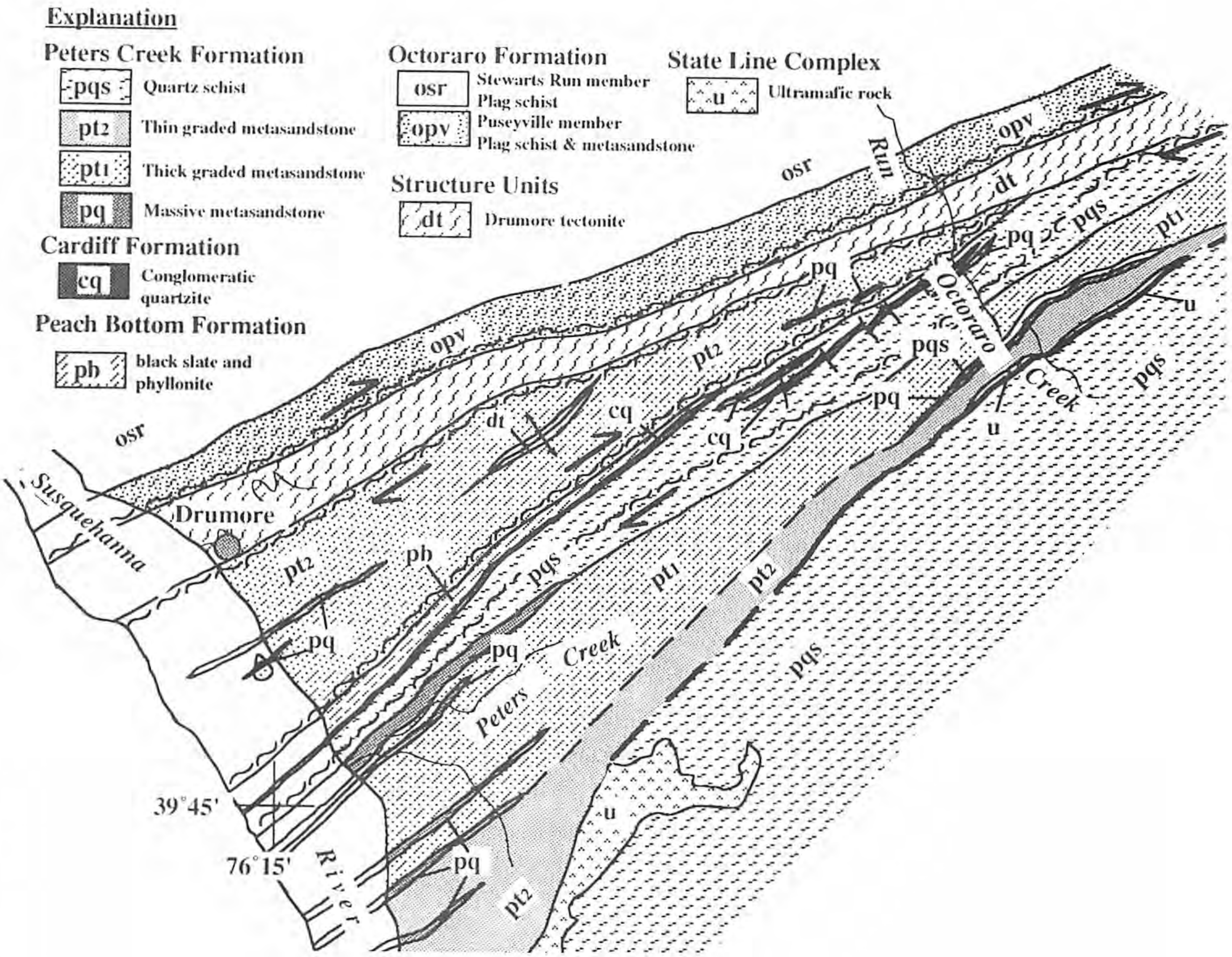


Figure 28. Geologic map of the Peach Bottom structure in Lancaster County, Pa.

Higgins (1972) was the first to formally question the original mapping and interpretation of Knopf and Jonas (1929) and of the workers that followed (Behre, 1933, Agron, 1950; Southwick and Fisher, 1967). In an elaborate review of the Piedmont nomenclature Higgins pointed out many inconsistencies in various reports and the lack of data to support a syncline in the Peach Bottom area. As an alternate interpretation Higgins (1972) proposed the Peach Bottom structure to be anticlinal based on local younging criteria in the Peters Creek metasandstones south of the belt and on cleavage-bedding intersection relationships. A narrow aeromagnetic low (Bromery and others, 1964) that corresponds directly to the Peach Bottom slate belt was interpreted by Higgins (1972) to support the anticline model. More recently, Faill and MacLachlan (1989) proposed that a "cryptic" suture coincides with the Peach Bottom structure based on the distribution of ultramafic bodies interpreted to be fragments of oceanic lithosphere. Smith (1993) suggested that the Cardiff quartzite and Peach Bottom slate were emplaced over the Peters Creek Formation along a talc-coated thrust fault and later folded into a syncline. Most recently a map-scale transcurrent shear zone was identified between the Peters Creek and Octoraro Formations just north of the Peach Bottom slate belt, and 150-200 km of dextral offset was interpreted across this zone (Valentino and others, 1994).

A tremendous amount of work has been done on the Peach Bottom problem since it was first proposed, and much of this work serves as the foundation for the current investigation. Except for the original mapping, very little work has focused on basic lithologic mapping in the region. In 1989 the Pennsylvania Geologic Survey initiated a mapping project to assess the detailed distribution of lithologies and structure in southern Lancaster County including the Peach Bottom slate, Cardiff quartzite, and the adjacent rock units. The area northeast of the Susquehanna River was mapped during the fall of 1989 to late spring of 1990 as the first stage of the project. This paper contains a summary of the analysis in Lancaster County. Future mapping will focus on the southwestern end of the structure in York County, PA and Harford County, MD.

MAPPING THE PEACH BOTTOM AREA OF LANCASTER COUNTY

A 7 x 13 km area was mapped for lithologic distribution and detailed structures that included the Peach Bottom slate belt, Cardiff conglomeratic quartzite, and numerous other members of the Peters Creek Formation (Figure 28). Since the original interpretation of the Peach Bottom syncline was based primarily on the mapped distribution of Cardiff quartzite around the northeastern end of the slate, a few critical areas were examined in detail: (1) the nose of the syncline as defined by the distribution of Cardiff quartzite (Figure 29), (2) the area where the Cardiff quartzite appears to split into two limbs in the fold (Figure 30), according to the earlier maps (Knopf and Jonas, 1929; Agron, 1950), and (3) the southwestward trace of the two conglomeratic quartzite layers (Figure 31) that define the syncline limbs.

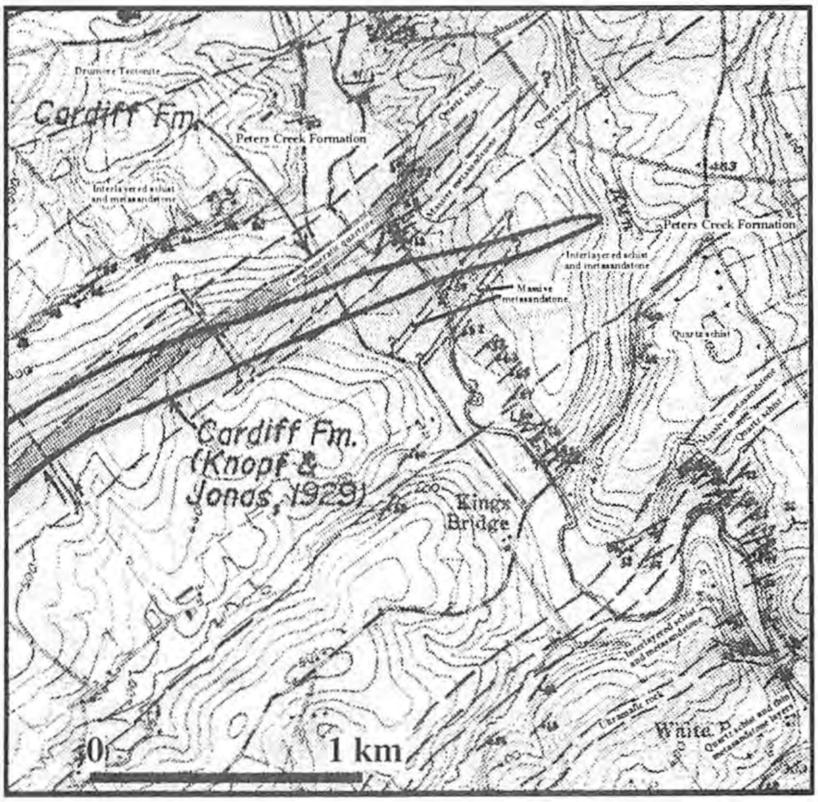


Figure 29. Geologic map of the northern end of the Cardiff Formation comparing the mapping of Knopf and Jonas (1929) and the current study.

Unit Descriptions

The Peach Bottom Formation. Black phyllitic-slate and slate of the Peach Bottom Formation outcrops in a narrow (30-500 m) belt trending northeast from the Susquehanna River for approximately 10 km where the slate is structurally cut off. The Peach Bottom formation contains up to 90 percent very fine muscovite and sericite, with accessory quartz, chlorite, chloritoid, and ilmenite. The slaty cleavage is the regional third generation fabric (Table 7) that is defined by the penetrative parallel alignment of the fine grained phyllosilicates. The slate occurs as a tabular shaped body that dips steeply to subvertically to the southeast and is parallel to the slaty cleavage.

Table 7. Deformation and metamorphic events in the western Piedmont of Pennsylvania (after Valentino, 1993; Valentino and others, 1994).

Deformation and Metamorphism for the Western Piedmont

D5 Weak north-south striking steeply dipping cleavage and associated crenulations		M3
D4: Post-transpressional extension Local conjugate cleavages $S4_s$ and $S4_a$. F4 open symmetric and conjugate box-folds		Chlorite-muscovite assemblage associated with discrete structures
D3: Dextral transpression Upright folds at all scales parallel (F3) to the shear zones and trending north-south ($F3_{NS}$), steeply dipping S3 schistosity in dextral shear zones, the Tucquan antiform, thrust fabrics ($S3_T$) in core of Tucquan antiform, juxtaposition of the Tucquan and Peters Creek structural blocks		
Tucquan block	$M2_{NW}$	Peters Creek block
$D2_{NW}$: Nappes $S2_{NW}$ schistosity and $F1_{NW}$ isoclinal folds	Regional metamorphism with higher grade at deeper structural levels in the Tucquan antiform.	$D1_{SE}$ Bedding parallel schistosity $S1_{SE}$ with rare intrafolial $F1_{SE}$ isoclinal folds
$D1_{NW}$: Early nappe $S1_{NW}$ schistosity and $F1_{NW}$ folds preserved in discrete domains and as inclusion trails within porphyroblasts of $M2_{NW}$ plagioclase	$M1_{NW}$ Mostly chlorite-muscovite-plagiocase assemblage, and rare garnet	$M1_{SE}$ Regional metamorphism with grade increase structurally higher

The Cardiff Formation. The Cardiff formation is a conglomeratic quartzite with deformed vein quartz pebbles with aspect ratios of 6:1 to 20:1. Quartz content ranges between 65% to 98 percent with less abundant minerals such as muscovite, chlorite, biotite, and chloritoid and accessory ilmenite, magnetite, tourmaline, zircon, and calcite from micro-veins. The quartz pebbles are clear and white with only minor inclusions of phyllosilicates. Some of the smallest clasts in the Cardiff Formation are blue rutilated quartz. The contact with the Peach Bottom black slate, structurally below, is very sharp and not gradational as described by earlier workers (Knopf and Jonas, 1929). The upper contact with chlorite-muscovite phyllitic-schist and muscovite-quartz phyllitic-schist of the Peters Creek Formation is gradational over a few meters. Smith (1993) showed that the Cardiff Formation is in contact with a very thin (<5 m) talc schist at the Susquehanna River. The Cardiff Formation is made up of three distinct tabular-shaped bodies separated by narrow belts of pelitic rock in Lancaster County, and only the northernmost Cardiff body is in contact with the Peach Bottom black slate (Figure 28). This field investigation resulted in a considerably different distribution of Cardiff Formation (Figures 29, 30, 31), as compared to the maps of previous workers.

The Peters Creek Formation. The Peters Creek Formation consists of three major lithologies: (1) numerous tabular shaped feldspathic metasandstone bodies (10's to 100's of meters thick and and as much as 6 km long), (2) a vast region of chlorite-muscovite-quartz schist, and (3) a thick sequence of interlayered lithologies (0.2-5.0 meters thick) containing metasandstones that grade into chlorite-muscovite schist. Minor lithologies are: (1) an ultramafic body composed of mostly serpentine with variable amounts of fibrous amphibole, (2) mafic meta-volcanics and volcaniclastics, and (3) muscovite-quartz phylitic schist. Compositional and stratigraphic analysis suggests that the Peters Creek Formation represents turbidite-fan deposits (Gates and Valentino, 1991; Valentino and Gates, in press; Valentino and Gates, this guidebook).

The Drumore Tectonite. The Drumore tectonite (Valentino and others, 1994) was named for the excellent exposures of this unit at the town of Drumore, Pennsylvania along the Susquehanna River, and this tectonite resides within a segment of the Pleasant Grove-Huntingdon Valley shear zone. This unit is characterized by silver-gray to silver-black, very fine grained, pelitic phyllonite. The Drumore unit is dominated everywhere by the regional

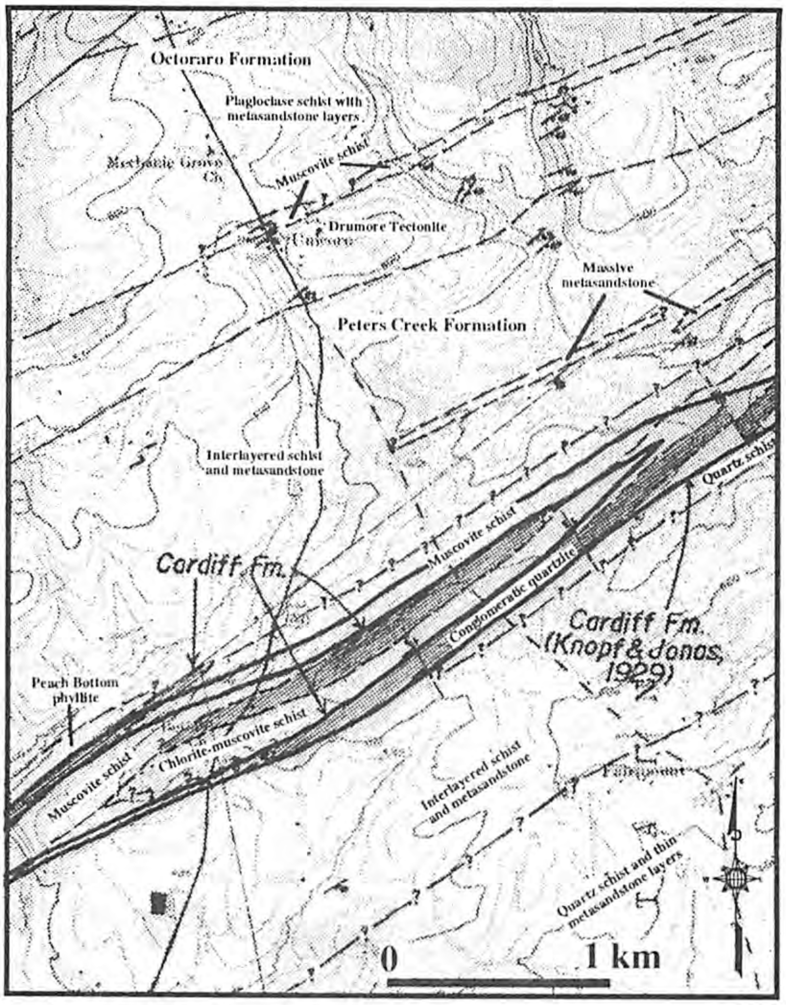


Figure 30. Geologic map of the Cardiff Formation comparing the mapping of Knopf and Jonas (1929) and the current study where the fold closure was interpreted.

S3 schistosity, but with close examination the preexisting metamorphic fabrics (S1nw & S1se of Table 7) can be observed. Contacts with the Peters Creek Formation on the south side and a metasandstone-rich section of the Octoraro Formation on the north are locally parallel to the regional foliation and often appear to be gradational. This unit has been mapped more than 40 km to the east of the Susquehanna River, and it outcrops in a relatively narrow belt ranging from 1 to 3 km wide. Retrograde metamorphism associated with D3 deformation (M3) is most complete within this phyllonite. Occasional vestiges of Peters Creek Formation metasandstone (to the south) and Octoraro Formation plagioclase schist (to the north) are preserved within the Drumore tectonite, suggesting this unit has a structural origin. Regionally, members of the adjacent Octoraro and Peters Creek Formations are truncated at a low angle against the northern and southern boundary of the tectonite (Figure 28).

The interpreted syncline nose at West Branch of Octoraro Creek

The easternmost exposures of Cardiff conglomeratic quartzite are on the hillside on the northeast bank of the West Branch of Octoraro Creek (Figure 29). Knopf and Jonas (1929) mapped the Cardiff formation approximately 300 m south of this location. The contacts of the Cardiff Formation are well constrained by the exposure along Octoraro Creek. Adjacent to the Cardiff Formation to the north and south is a chlorite-muscovite schist bearing medium- to finegrained muscovite, chlorite, quartz, and oxides. The Cardiff Formation is approximately 75 meters thick measured perpendicular to the internal foliation that strikes 046^{o} and dips 68^{o} southeast. The internal foliation (S1se) is parallel to the compositional layering defined by localized concentration of micas. Parallelism of the foliation and compositional

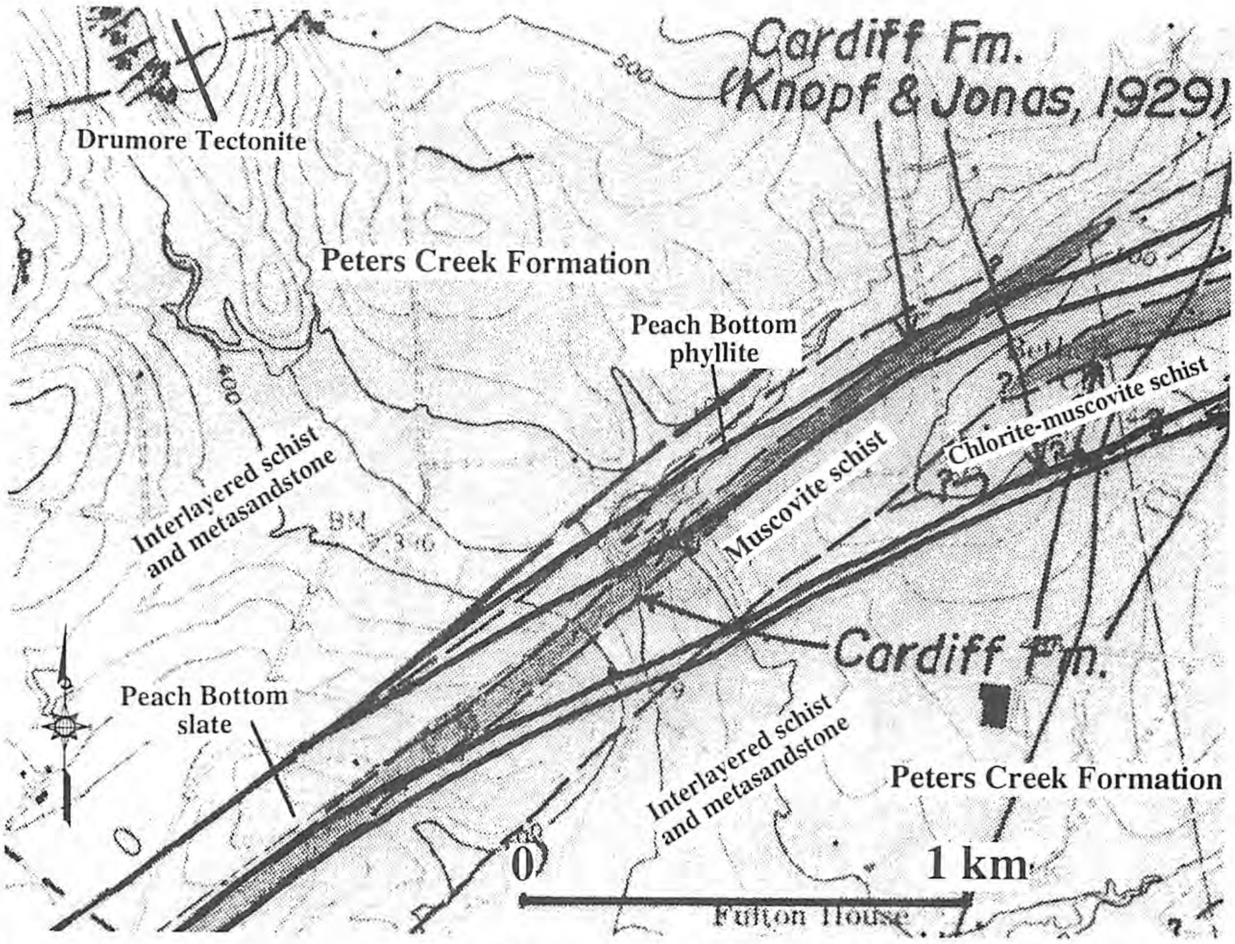


Figure 31. Geologic map of the Cardiff Formation and adjacent units comparing the mapping of Knopf and Jonas (1929) and the current study where the limbs of the fold were interpreted.

layering exists across the entire width of the outcrop at this location.

The Cardiff Formation is not present along strike to the northeast. However, a single narrow ridge is underlain by quartzite along strike to the southwest (Figure 32). Knopf and Jonas (1929) mapped the Cardiff Formation as nearly 200 m wide, including the narrow ridge and part of the northwest slope shown in Figure 29. Float and blocks of conglomeratic quartzite were found at the base of the hill, but this relationship is not uncommon for a narrow quartzite ridge. Aerial photography and oblique aerial photos show a narrow continuous ridge on top of the hill (80 m wide) that projects along strike to the exposures observed on the West Branch of Octoraro Creek (Figures 32 and 33). Along the narrow ridge there are no outcrops in place. Therefore, the structural information collected at the creek and the air photography are the primary data from which the narrow linear distribution of Cardiff Formation is interpreted. More exposure and a wider ridge would be expected from the distribution of quartzite that the earlier workers portrayed on their maps.

Cardiff Formation east of Pennsylvania Route 222

Figure 30 shows the detailed distribution of Cardiff Formation of Knopf and Jonas (1929) as compared with the distribution of this investigation. Crossing a small valley to the southwest, the narrow ridge previously discussed projects into the only outcrops of Cardiff Formation in the area (Figure 33A). Most of the exposure is slumped, but where it is in place the internal foliation is consistent with that measured along the West Branch of Octoraro Creek. A belt of quartzite more than 300 m wide was portrayed by Knopf and Jonas (1929) to divide into the limbs that define the Peach Bottom syncline just west of this small valley. It is interesting that their 300-m-wide quartzite only forms a narrow ridge less than 60 m

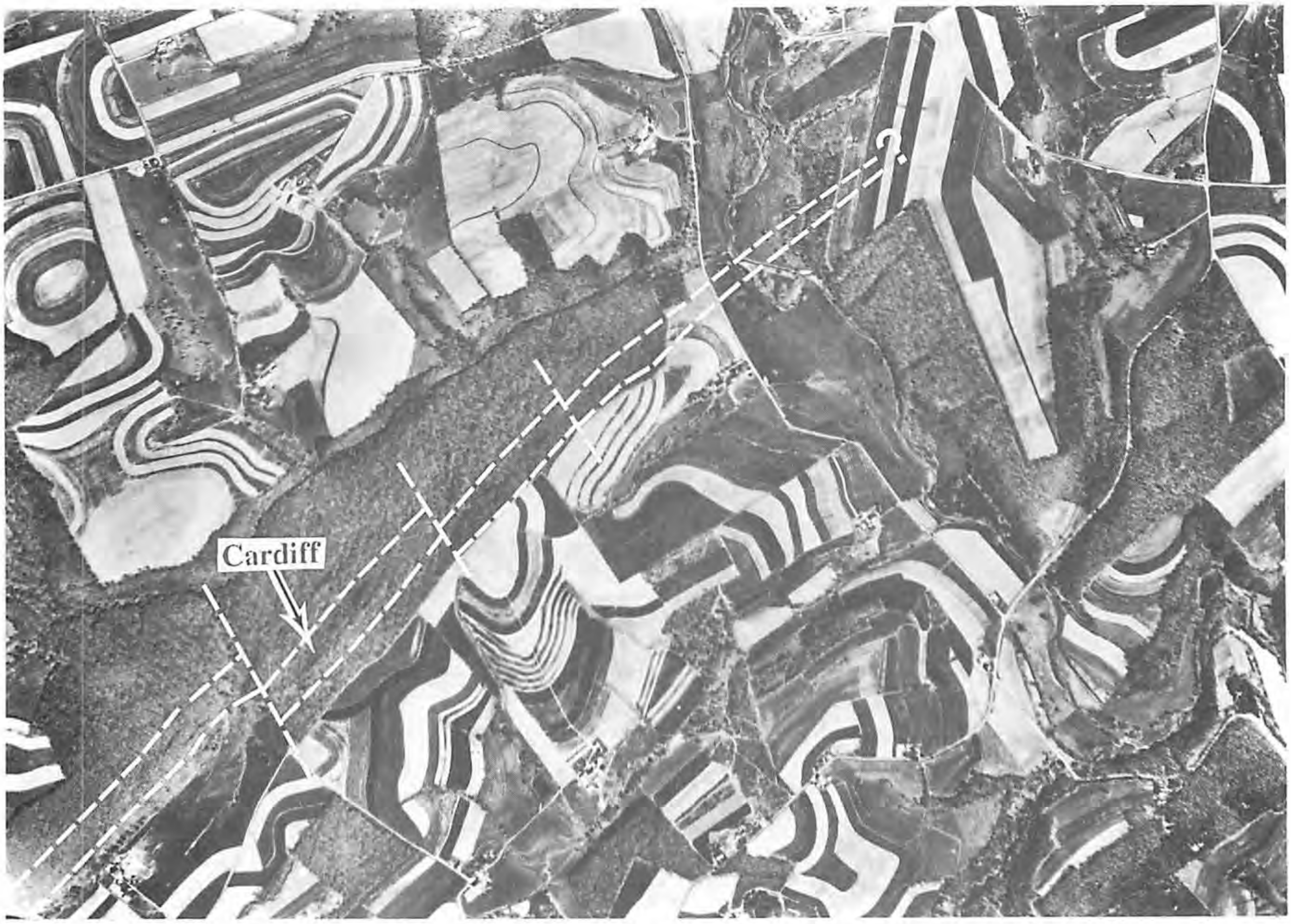


Figure 32. Aerial photograph of the ridge underlain by Cardiff Formation just southwest of the West Branch of Octoraro Creek.

wide. The southern Cardiff Formation of Knopf and Jonas (1929) coincides with the steep southern slope of the hill while the northern quartzite resides on the gentle northern slope. The terrain where the northern and southern quartzites apparently merge (Knopf and Jonas, 1929) is gently sloped to the northeast with no prominent ridge or exposures. The lack of a topographic high and exposures is anomalous for a sequence of conglomeratic quartzite that is more than 300 m thick and is surrounded by deeply weathered pelitic lithologies. Float of part of the Peach Bottom slate was found in the valley just northeast of the proposed fold nose. The northeastern limit of Peach Bottom slate as mapped by Knopf and Jonas (1929) is more than 400 m southwest of this locality (Figure 30). The lack of a prominent ridge for the proposed northern quartzite and the presence of black slate float where the northern and southern quartzites are suppose to merge suggests that the distribution of the Cardiff quartzite of Knopf and Jonas (1929) is erroneous.

New excavation along PA Route 222 in the winter of 1990 produced exposures of silver-tan muscovite-quartz phyllitic-schist in the location where the northern conglomeratic quartzite was shown by Knopf and Jonas (1929) and Agron (1950). Abundant float of fine grained chlorite-muscovite schist was observed on the south side of the hill. This area was mapped by earlier workers as Peach Bottom slate, however, float of schist was found on top of the broad hill in a freshly plowed corn field. Abundant float of Cardiff Formation, including some very large blocks (1-2 m across), is present in the wooded area on the northern side of the hill (Figure 30). The edge of the corn field (in 1990) approximately marks the northern limit of chlorite-muscovite schist float and the beginning of quartzite float. The abundant Cardiff Formation float in the wooded area is interpreted to be derived from a second quartzite body for the following reasons: (1) the colluvium and float of quartzite on the northern slope of the hill resides topographically higher than quartzite on the southern slope, and (2)

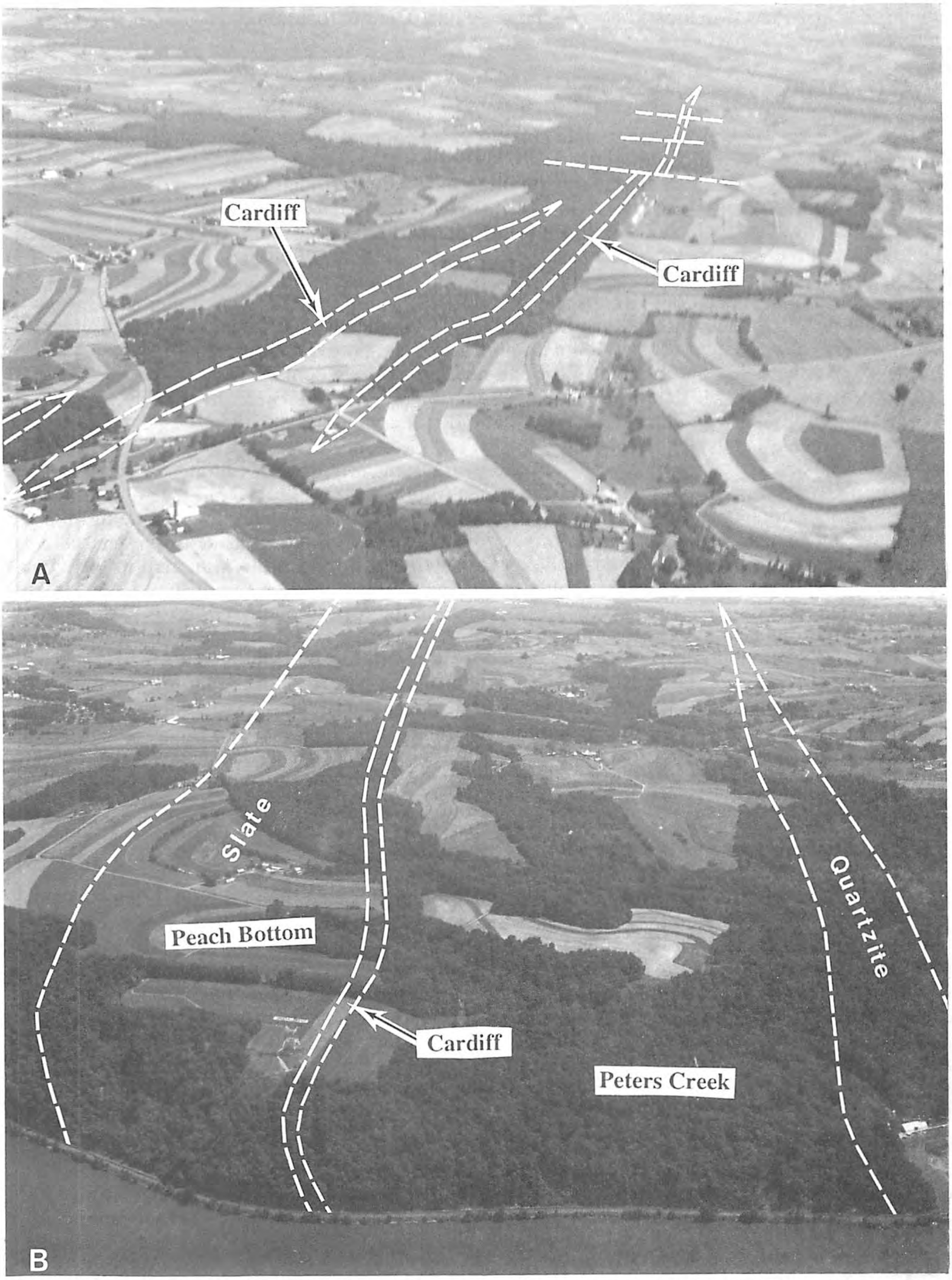


Figure 33. [A] Oblique aerial photograph of the ridge underlain by Cardiff Formation just southwest of the West Branch of Octoraro Creek. The view is looking northnortheast. [B] Oblique aerial photograph of the Peach Bottom section at the Susquehanna River. The view is looking northeast.

phyllitic schist float was found between the conglomeratic quartzite on the southern slope and the conglomeratic quartzite float on the northern slope.

Cardiff Formation at Shoemaker Road

There is a small exposure of granule conglomerate containing rounded grains of blue and gray quartz in the field along Shoemaker Road (Figure 31). The foliation strikes 033^{o} and dips 61^{o} southeast. Float of silver-gray phyllitic material was found northwest of this quartzite while float of silver-tan phyllitic-schist and chlorite muscovite-quartz schist were found to the southeast. Knopf and Jonas (1929) mapped the Peach Bottom slate south of this quartzite, but there is no evidence for any of the Peach Bottom lithologies. However, the silver-gray phyllitic material on the north side is very similar to phyllite associated with the Peach Bottom Formation. Knopf and Jonas (1929) also mapped a second quartzite farther south that projects through the Aument Dairy Farm. Abundant float of chlorite-muscovite-quartzite schist was found in that area with no float of conglomeratic quartzite. The sequence of lithologies suggested by float and outcrop along Shoemaker Road is as follows from southeast to northwest: (1) chlorite-muscovite-quartz schist (Peters Creek Formation), (2) silver-tan phyllitic-schist (Peters Creek Formation), (3) conglomeratic quartzite (Cardiff Formation), and (4) silver-gray phyllite-schist (Peach Bottom Formation).

Southwest across the Conowingo Creek valley the same sequence of lithologies was observed in a freshly plowed corn field on top of the hill in March, 1990 (Figure 31). The quartzite present along Shoemaker Road crosses a zone of conglomeratic quartzite float on the hill top when the Shoemaker Road quartzite is projected southwestwards along strike of the internal foliation. This indicates that the quartzite on the southwest hilltop is the same unit as the quartzite at Shoemaker Road. During winter months one can stand on top of the western hill where conglomeratic quartzite appears as float and see the exposure at Shoemaker Road directly along strike to the northeast. It is apparent that these quartzites are the same unit. Knopf and Jonas (1929) mapped two quartzites on the western hill top, but, there is no field evidence to support their conclusion. The single quartzite observed on the hill during this investigation was considered by earlier works to be the southern quartzite that defines the southern limb of the Peach Bottom syncline. The quartzite at Shoemaker Road was considered the northern quartzite that defined the northern limb. The quartzite at Shoemaker Road lies along a discontinuous linear outcrop belt of quartzite that was traced from the Susquehanna River.

Lithologic sequence across the Peach Bottom structure

For the Peach Bottom structure to be synclinal or anticlinal the lithologic sequence northwest of the interpreted hinge axis should be a mirror image of the lithologic sequence southeast of the hinge axis, assuming one of the limbs is not faulted. This investigation demonstrated that the Cardiff conglomeratic quartzite most likely is not continuous around the northeastern end of the slate belt as mapped by Knopf and Jonas (1929) and Agron (1950), nor does the lithologic sequence of the Peters Creek Formation occur immediately adjacent to the Cardiff Formation. Earlier workers partially explained the lack of a repeating lithologic sequence by the proposed faulted northern limb of the Peach Bottom syncline. The following is a presentation of the detailed lithologies across the Peach Bottom structure and the sequences northwest and southeast of the Peach Bottom Formation at various localities along strike in Lancaster County:

Lithologic Sequence at the Susquehanna River

Southern Sequence	**Formation**
Black slate	Peach Bottom
Slate grades into silver-black phyllitic-slate (50 m)	Peach Bottom
Conglomeratic quartzite (40 m)	Cardiff
Talc schist (2-3m)	
Chlorite-muscovite phyllitic-schist (30 m)	Peters Creek
Silver-tan muscovite-quartz phyllitic-schist (125 m)	Peters Creek
Chlorite-muscovite-quartz schist with metasandstone (>300 m)	Peters Creek

Northern Sequence

Black slate	Peach Bottom
Dolomite-quartz-talc rock (~10 m)	
Talc schist (~2 m)	
Muscovite-chlorite-quartz schist and metasandstone (60 m)	Peters Creek
Interlayered greenstone and schist (~10 m)	Peters Creek
Muscovite-chlorite-quartz schist and metasandstone (>700 m)	Peters Creek

Lithologic Sequence at Cherry Hill Road

Southern Sequence

Black Slate	Peach Bottom
Slate grades into silver-black phyllitic-slate (10 m exposed)	Peach Bottom
Fine to coarse grained conglomeratic quartzite (40 m)	Cardiff
Chlorite-muscovite phyllitic-schist (30 m)	Peters Creek
Silver-tan muscovite-quartz phyllitic-schist (125 m float in woods)	Peters Creek
Chlorite-muscovite-quartz schist with metasandstone (>400 m)	Peters Creek

Lithologic Seuqence at Tanyard Hollow Road

Southern Sequence

Black slate (exposed along the road)	Peach Bottom
Conglomeratic quartzite (exposed in old quarry)	Cardiff
Silver-tan muscovite-quartz phyllitic-schist (contact exposed)	Peters Creek

Northern Sequence

Black slate	Peach Bottom
Black slate with abundant vein quartz (10-15 m)	Peach Bottom
Chlorite-muscovite-quartz schist	Peters Creek

STRUCTURE OF THE PEACH BOTTOM AREA

The mapped distribution of the Cardiff, Peters Creek, and Peach Bottom Formations do not define an antiform or a synform in Lancaster County, regardless of the oriention of the interpreted fold (Figures 28, 29, 30, and 31). This may not be the case for the southwestern end in Harford County, MD. The mapped distribution of mesoscopic structures defines a steeply southeast-dipping, northeast-striking, tabular shaped zone containing third generation upright folds (F3), axial planar schistosity (S3) and moderate to shallow east plunging extension lineations (L3). Microstructural analysis revealed abundant evidence for transcurrent slip and horizontal compression on S3 parallel to L3 throughout the tabular zone and along the length. New growth of chlorite and muscovite associated with D3 structures is restricted to the limit of S3 schistosity that is confined to a relatively narrow zone containing the Peach Bottom Formation. The Peach Bottom and Cardiff Formations reside within a zone of S3 schistosity that merges with the Pleasant Grove-Huntingdon Valley shear zone toward the northeast (Figures 27 and 28). This investigation portrays the Peach Bottom structure including the Peach Bottom, Cardiff, and portions of the Peters Creek Formations as residing in a segment of a transcurrent shear zone splay off of the main Pleasant Grove-Huntingdon Valley zone. The southwestward extension of this transcurrent splay is not well delineated because the mapping has yet to be completed, however, if the splay continues parallel to the Peach Bottom and Cardiff Formations it appears that it probably merges with the main shear zone south of Broad Run in Maryland.

Regional Structure

South of the Pleasant Grove-Huntington Valley zone the regional metamorphism spans the chlorite- to biotite-zones (Faill and Valentino, 1989; Valentino and Faill, 1990; Valentino and Faill, this guidebook) and is associated with the schistosity that Freedman and others (1964) termed S1, being the first generation for the area. This schistosity is termed S1se because it is the first schistosity south of the Pleasant Grove zone (Table 7). This early schistosity was related to Taconian (Lapham and Bassett, 1964) nappe emplacement to the northwest (Freedman and others, 1964; Wise, 1970). There are very few folds associated with the S1se schistosity, and the folds that are present are intrafolial folds. South of the Peach Bottom Formation the regional schistosity strikes approximately northeast-southwest and dips moderately to steeply to the southeast. Late D4 deformation is characterized by development of symmetric open folds, conjugate box-folds (F4), and conjugate cleavages (S4).

S3 schistosity at Peach Bottom

In the area of the Peach Bottom structure, the regional third generation steeply southeast dipping schistosity cross cuts and deforms the S1se schistosity (Freedman and others, 1964; Valentino, 1990; Valentino, 1991; Valentino and others, 1994). Freedman and others (1964) and Wise (1970) portrayed an even distribution of the S3 schistosity along the length of the lower Susquehanna River valley, however, although evidence for S3 exists across the entire region as weakly developed crenulation cleavage, the S3 is mostly restricted to a relatively narrow zone approximately 3 km wide (Valentino, 1990; 1991; Valentino and others, 1994) between the town of Drumore and just north of Peach Bottom. Within the Peach Bottom area the S3 schistosity is generally defined by recrystallized muscovite and chlorite, as well as planar layers of recrystallized quartz. The intensity of schistosity developed is heterogeneously distributed throughout the zone. In competent metasandstone layers of the Peters Creek Formation and the Cardiff Formation, the S3 foliation is generally defined by flattened quartz pebbles and recrystallized quartz matrix. In pelitic portions of the Peters Creek Formation, the Drumore tectonite, and Peach Bottom Formations, the S3 schistosity is generally very penetrative and defined by recrystallized muscovite and chlorite, and rare recrystallized thin quartz ribbons.

F3 folds at Peach Bottom

The S3 schistosity is axial planar to folds that preserve earlier structural fabrics in the hinge regions. In most cases these F3 folds are defined by folded S1se schistosity, but in some of the metasandstone units of the Peters Creek Formation the folds are defined by folded compositional layering interpreted to represent primary sedimentary bedding (Valentino and Gates, this guidebook). Although the new map pattern of lithologies does not support the Peach Bottom fold models of earlier workers, a map-scale F3 fold was recognized. A lens-shaped outcrop pattern of Drumore tectonite is located within the Peters Creek Formation northwest of the Peach Bottom slate belt (Figure 28). The trend of this lens is oblique to the trend of the S3 structure zone approximately 10-15^{o} counter-clock-wise. The position of this lens of Drumore tectonite with respect to the belt of tectonite, which dips steeply beneath the Peters Creek Formation about 0.5-1.0 km to the northwest, suggests that this lens represents the reemergence of Drumore tectonite in an F3 antiform at the present level of erosion. The size and geometry of this proposed antiform is consistent with F3 folds that occur in the Lancaster Valley tectonite zone located 25 km to the north (Valentino and MacLachlan, 1989; Valentino, 1990; Faill and MacLachlan, 1990).

Linear fabrics at Peach Bottom

Lineations in the Peach Bottom area are defined by three elements: (1) intersection lineations between S3 and S1se/S0 which are termed $L_{S3XS1se}$, (2) F3 fold hinge axes including the hinge axes of crenulations that are micro-F3 folds (L_{F3}), and (3) mineral elongation lineations (L3) defined by mica streaks, symmetric and asymmetric quartz and mica pressure fringes on porphyroclasts, and elongate quartz pebbles in sheared metasandstone units.

The local orientation of intersection lineations and fold hinge axis lineations is dependent on the local original orientation of the fabric that was folded or cross cut. These fabrics are generally moderately to shallowly northeast plunging. Mineral elongation lineations generally trend to the northeast along the length of the structure, however, the plunge of extension lineations may vary from 0^{o} to 40^{o} depending on the location in the deformation zone. At the Susquehanna River, there is a southern and northern domain. The southern domain contains subhorizontally-oriented, mineral-elongation lineations defined by elongate quartz pebbles in the Cardiff Formation and mica streaks in the Peters Creek

Formation. In addition, symmetric and asymmetric quartz and muscovite pressure fringes on chloritoid and pyrite grains define a subhorizontal extension lineation in the Peach Bottom Formation. The northern domain contains mineral-elongation lineations that are defined by chlorite pressure fringes on magnetite grains, mica streaks, and linear aggregates of quartz from metasandstones and which plunge 30^{o} to 40^{o} NE. Along strike toward the northeast the shallowly plunging mineral elongation lineations are dominant in the Peach Bottom structure.

Structure Boundaries

The Peach Bottom structure is distinguished by a broad zone of penetrative S3 schistosity, F3 folds, and L3 lineation. The zone varies in structural width along strike. The boundaries of this zone range from broad transition zones up to 1 km wide, to transition zones less than 100 m wide. The northern boundary of the Peach Bottom structure at the Susquehanna River is defined by the abrupt appearance of semi-penetrative S3 schistosity and F3 folds in the pelitic Drumore tectonite. This structural boundary has been traced parallel to the contact between the Drumore tectonite and the Octoraro Formation 50 km to the northeast. A more complex situation exists along strike for the southern boundary of the Peach Bottom structure. At the Susquehanna River, where rocks are best exposed, the S3 schistosity transition zone occurs over about 100 m of outcrop. Rocks dominated by the S3 schistosity gradually give way to weaker and weaker S3 and more preserved evidence for earlier metamorphic and deformation fabrics. Freedman and others (1964) claimed that S3 merged with S1se south of the Peach Bottom syncline, however, microscopic analysis of these rocks revealed no evidence for this conclusion, therefore confining the effects of S3 schistosity development to this relatively narrow zone.

KINEMATIC ANALYSIS

Kinematic analysis was completed along the mapped length of the Peach Bottom structure at the map scale, outcrop scale and microscopic scale. Freedman and others (1964) and Wise (1970) concluded that, on the basis of the orientation and geometry of F3 folds, these folds developed as the result of northwest-southeast directed compression. The presence of abundant mineral elongation lineations that trend approximately parallel to the strike of the Peach Bottom structure suggest a component of lateral slip or non-coaxial deformation.

Micro- and Meso-scopic kinematic analysis

The distribution of pressure fringes about porphyroclasts, S-C fabrics, quartz preferred grain shape, and quartz crystallographic preferred orientation data were used during microscopic structural analysis. Small laths of chloritoid in the Peach Bottom Formation commonly occur with pressure fringes of quartz and muscovite. The long axis of the pressure fringes is generally subhorizontally oriented consistent with mesoscale mineral elongation lineations. In X-Z sections pressure fringes are distributed both symmetrically and asymmetrically about the chloritoid grains. The asymmetrically distributed pressure fringes have a geometry indicative of dextral shear. The Drumore tectonite contains abundant pyrite porphyroclasts with pressure fringes of quartz consistent with dextral shear. Magnetite porphyroclasts with asymmetrically distributed chlorite pressure fringes in deformed Peters Creek Formation reveal a consistent dextral shear sense. Some samples collected in the region of moderately plunging mineral elongation lineations gave results of local oblique dextral slip. The Cardiff conglomeratic quartzite often contains well developed type I S-C mylonitic fabrics (Lister and Snoke, 1984) that are consistent with dextral shear, as well as type II S-C fabrics in the Peters Creek Formation and the Drumore tectonite. Thin mylonitized quartz veins often have preferred grain shape orientation consistent with dextral offset. Numerous mesoscopic kinematic indicators were observed in the field. Mylonitized quartz veins in chlorite-muscovite schist from the Peters Creek Formation often show asymmetry of the vein with respect to the mylonitic foliation consistent with dextral shear. The rotation sense of the S1se schistosity into discrete shear bands also demonstrates the dextral shear sense.

Map-scale kinematic analysis

The bedrock map pattern within and near the boundaries of the zone of penetrative S3 schistosity support right lateral offset. A chlorite-muscovite phyllitic member of the Peters Creek Formation, located adjacent to the Cardiff Formation at the Susquehanna River, pinches out against the sheared Cardiff Formation to the northeast where a muscovite-quartz schist member of the Peters Creek Formation is in contact with the Cardiff Formation (Figure 28). The truncation of the chlorite-muscovite phyllite is most likely the result of extensive dextral strike-slip deformation. Similarly the northern boundary of the zone of penetrative S3 schistosity crosses the area where the Peach Bottom Formation and northern lens of the Cardiff Formation end abruptly (Figure 28). This pattern of discontinuous lithologies near the structural boundary of a regional zone of high strain is also suggestive of dextral strike-slip truncation and offset. With this model, structural blocks of Peach Bottom slate or Cardiff quartzite could potentially exist along the southern margin of the Drumore tectonite farther east.

STRUCTURAL MODEL FOR THE PEACH BOTTOM STRUCTURE

Deformation in the Peach Bottom structure is characterized by a 2-km-wide zone of upright F3 folds combined with variably developed axial planar S3 schistosity. There are two zones of well developed S3 schistosity in this region: (1) the northern zone is about 1.5 km broad and correlates directly with a silver-black phyllonite of the Drumore tectonite which is part of the Pleasant Grove-Huntingdon Valley shear zone (Valentino and others, 1994); (2) the southern zone is about 1 km wide and correlates directly with the black slate (ultraphyllonite?) of the Peach Bottom Formation, mylonitized quartzite of the Cardiff Formation, and mylonitized metasandstone and phyllonite of the Peters Creek Formation. These two deformation zones merge approximately 12 km northeast of the Susquehanna River (Figure 28), where the zone including the Drumore tectonite continues eastward. There is a wedge shaped body of Peters Creek Formation that is intensely deformed by F3 folding and only moderately developed S3 schistosity located between the Drumore tectonite and the Peach Bottom structure at the Susquehanna River (Figure 28). This wedge-shaped body of Peters Creek Formation is possibly a large strike-slip duplex.

The maximum elongation direction within the Drumore tectonite and Peach Bottom structure was subhorizontal as revealed by abundant extension lineations and the quartz petrofabrics. Kinematic analysis at all scales consistently shows a component of dextral shear associated with the S3 schistosity parallel to the maximum elongation direction. The orientation of the abundant F3 folds suggests a component of subhorizontal compression with the bulk shortening direction oriented at a high angle to the Peach Bottom structure. Microscopic and map-scale kinematic analysis suggests that a major component of the deformation was non-coaxial shear parallel to the Peach Bottom structure.

The combination of contractional and strike-slip structures that developed during the same deformation event is consistent with a model of dextral transpression. There is a clear relationship between the relative timing of lateral slip and schistosity development. Since all the evidence for strike-slip deformation is parallel to the steeply dipping S3 schistosity, this schistosity must have developed prior to or as the result of shearing. In many places the S3 schistosity is only axial planar to the F3 folds and has not experienced any component of strike-slip deformation. It can be argued that the lateral slip component of the transpressive deformation only occurred after development of a favorably oriented structure that was steeply dipping.

LITHOFACIES INTERPRETATION

The Peach Bottom Formation is not a simple prograde slate because it experiences multiple penetrative deformation and metamorphic episodes. The fine grained slate-like appearance may be due exclusively to grain size reduction of micas in the shear zone to produce an ultramylonite composed of primarily micas similar to the Drumore tectonite. Although it was reported in the geologic literature that the Peach Bottom slate contains carbon (e.g. Smith, 1993) no geochemical analyses have been reported to support that conclusion. The dark color of the Peach Bottom Formation is most likely due to the presence of extremely fine grained micas. The Peach Bottom slate and Cardiff quartzite was correlated with black slates and conglomerate in the lower Chilhowee Group of the Chickies Formation of Early Cambrian age by Knopf and Jonas (1929). Later, Fisher and others (1979) and Smith (1993) correlated the Peach Bottom slate with the Arvonia slate belt of northern Virginia based on lithologic similarities and because the Arvonia slate resides in a synclinal structure. Both of these regional correlations are problematic.

The composition of the Cardiff Formation is not unusual for large massive metasandstone bodies in the Peters Creek Formation (Figure 34). Grain size appears to be the only difference between Peters Creek Formation massive metasandstone and the conglomeratic quartzite of the Cardiff Formation. Because the Cardiff Formation is in very close spacial association with the Peters Creek Formation and has an identical composition to other local metasandstone bodies, it is reasonable to assume that the Cardiff Formation is part of the same lithotectonic sequence. This association is much more reasonable than correlating the Cardiff Formation with the Chickies Formation tens of kilometers across structural strike crossing a major strike-slip shear zone (the Pleasant Grove-Huntingdon Valley zone), and a multitude of other thrust-related structure that occurs in the Piedmont.

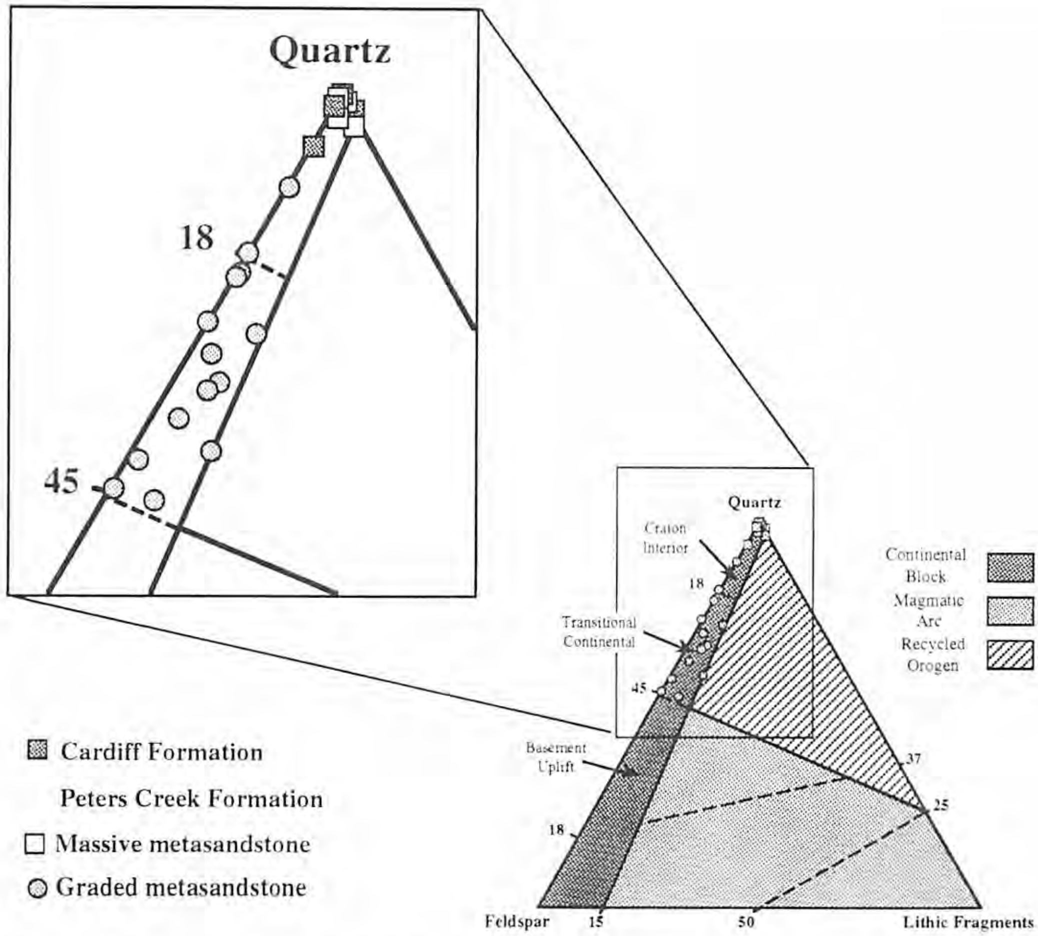


Figure 34. QFL plot for Peters Creek Formation massive metasandstone bodies and for the Cardiff Formation conglomeratic quartzite.

Correlation of the Peach Bottom Formation with the Arvonia slate belt of northern Virginia requires that the Peach Bottom Formation be Ordovician in age (Fisher and others, 1979; Smith, 1993), and requires correlation over hundreds of kilometers. The Peach Bottom Formation contains the regional S1se schistosity that is associated with the Taconian orogeny (Freedman and others, 1964; Wise, 1970), therefore, by the principle of cross cutting relationships the Peach Bottom Formation must be at least pre-Taconic in age. Although it is stated elsewhere that the Peters Creek Formation is part of a tectonic melange (Horton and others, 1989; Smith, 1993), the stratigraphy and composition of coarse grained detritus suggest that the Peters Creek Formation was deposited as a turbidite-fan sequence with a granitic basement source (Gates and Valentino, 1991; Valentino, 1993; Valentino and Gates, in press; Valentino and Gates, this guidebook). The presence of minor metabasalt and metavolcaniclastic interlayered with feldspathic metasandstone suggests the tectonic depositional setting was rifting, most likely Late Proterozoic-Early Cambrian rifting (Gates and Valentino, 1991). A protolith for the Peach Bottom Formation would be pelitic shales with minor siltstone, and such a lithology could have originated as a restricted basin deposit, either deep water environment or lacustrine. It is not unreasonable for the Peach Bottom Formation to represent the earliest lithofacies in a rift-related depositional tectonic environment. This model simplifies the regional stratigraphy and does not require tenuous correlation of the Peach Bottom Formation over hundreds of kilometers, but only requires that Peach Bottom Formation to be in association with the adjacent Cardiff and Peters Creek clastic sequence.

CONCLUSIONS

The main conclusions of this investigation on the Peach Bottom area of Lancaster County, Pennsylvania are the following:

1. The distribution of the Cardiff Formation as determined by new mapping shows that it does not wrap around the northeastern end of the Peach Bottom Formation as indicated by earlier workers (Knopf and Jonas, 1929; Agron, 1950). Therefore, a fold structure, syncline or anticline, does not adequately explain the lithologic distribution.
2. At the Susquehanna River the Peach Bottom structure is a 1-1.5-km-wide zone of penetrative S3 schistosity separated from Pleasant Grove-Huntingdon Valley zone by a 2 km thick section of Peters Creek Formation. The Peach Bottom structure merges with the Pleasant Grove-Huntingdon Valley zone toward the northeast.
3. Kinematic analyses conducted in the Drumore tectonite and Peach Bottom structure reveal a consistent dextral shear sense.
4. The Peach Bottom and Cardiff Formations are the structurally lowest units in a monoclinal sequence containing the Peters Creek Formation rift-related clastics. By including the Peach Bottom and Cardiff Formations in the same lithotectonic and lithostratigraphic section as the Peters Creek Formation this eliminates the need for correlations over tens to hundreds kilometers as proposed by earlier workers.

PATTERNS OF REGIONAL METAMORPHISM IN THE CENTRAL APPALACHIAN PIEDMONT OF PENNSYLVANIA, THE APPLICATION OF GARNET-CHLORITE THERMOMETRY, AND DIFFERENCES IN PALEOZOIC TECTONOTHERMAL HISTORIES

David W. Valentino[1] and Rodger T. Faill[2]
(1) Concord College
Athens, WV 24712
(2) Pennsylvania Geological Survey
P. O. Box 8453, Harrisburg, PA 17105-8453

ABSTRACT

A map of metamorphic zones was assembled for the central Appalachian Piedmont of Pennsylvania to delineate regional patterns of metamorphism with respect to known geologic structures. Four main phases of metamorphism affected the rocks of the Piedmont: (1) Grenvillian age amphibolite to granulite facies confined to basement massifs; (2) a 500 Ma granulite facies local to the Wilmington Complex and adjacent units; (3) a Taconian regional metamorphism that ranges from greenschist to granulite facies; and (4) a Late Paleozoic greenschist facies overprint related to transpressional shearing and folding. The Taconian metamorphism in the Pennsylvania Piedmont can be divided into two northeast-trending belts which have contrasting metamorphic patterns.

In the southeastern belt, Taconian metamorphism increases in grade from northwest to southeast, corresponding to increasingly higher structural levels toward the southeast. Earlier workers interpreted this pattern to be a consequence of Taconian obduction of a magmatic arc that provided a heat source from above. In the northwestern belt, Taconian metamorphic zones are symmetrically distributed on each side of the large (27 km broad) Tucquan antiform. This structure, defined by folded Taconian schistosity, is a basement-cored, upright, open arch that developed during Late Paleozoic dextral transpression. From limb to crest, the characteristic assemblages in the pelites are: (a) chlorite-muscovite- $\pm$ plagioclase; (b) chlorite-biotite-muscovite $\pm$ plagioclase; (c) biotite-chloritoid; and (d) chlorite-garnet $\pm$ chloritoid. Rare kyanite also occurs in the Tucquan core. A high-Al bulk composition precluded a biotite-garnet assemblage in these rocks. These Taconian metamorphic assemblages were heterogeneously affected by a lower greenschist facies overprint during development of the antiform in the Late Paleozoic.

The exposed sequence containing garnets [assemblage (d) above] in the hinge of the Tucquan antiform is structurally about 3 km thick. Despite late chlorite growth on the edges of some garnets, electron-microprobe imagery demonstrated that the garnet internal chemical zonation is pristine. With a 20:1 chlorite/garnet ratio, minor production of late chlorite at the expense of garnet had little effect on the primary chlorite compositions. Chlorite-garnet thermometry yielded temperatures between 400-550^{o}C, with the lowest temperatures occurring generally near the two edges of the garnet zone (the structurally highest part), with progressively higher temperatures toward the center (antiform hinge). This pattern suggests the thermal source was from below. The contrast in metamorphic pattern between the southeast belt and the northwest belt in the Pennsylvania Piedmont suggests these two belts experienced different metamorphic and tectonic histories.

INTRODUCTION

The metamorphic history of the Pennsylvania Piedmont is complex, as has been recognized for some time (e.g., Crawford and Crawford, 1980). Of the four metamorphic episodes mentioned in the abstract, this paper focuses on the most regional of the metamorphisms, the one related to the Late Ordovician Taconian orogeny. Widespread as it was, the development and distribution of the various mineral assemblages do not form a simple pattern. The complexities of the metamorphic zones reflect both the variation in the Taconian tectonism across the Piedmont (Wise, 1970), and the subsequent Alleghanian transpression (Valentino and others, 1994). The essence of this paper centers on the contrast between two parts of the Piedmont that have undergone very different metamorphic histories, and that were later juxtaposed. It is a tale of two adjacent, northeast-southwest trending belts that were formed in different settings, and of the boundary formed between them when they were brought together. The two belts comprise most of the Piedmont in southeast Pennsylvania.

The northwestern belt encompasses the Tucquan antiform along its entire length (Figure 35). On its northeast end, the Grenvillian gneisses of Mine Ridge constitute the core, which plunges gently to the southwest. This gneiss core is overlain by the Chilhowee sequence. Farther to the southwest, the Tucquan antiform is expressed as a single large fold of the Taconian foliation in the Octoraro schists. This Tucquan antiform persists farther to the southwest into north-central Maryland (Figure 35). The northwestern zone is bounded on the north by the Marburg phyllite and the carbonates of the Lancaster Valley; on its south side, the Pleasant Grove-Huntingdon Valley shear zone constitutes the boundary along its entire length (Valentino and others, 1994).

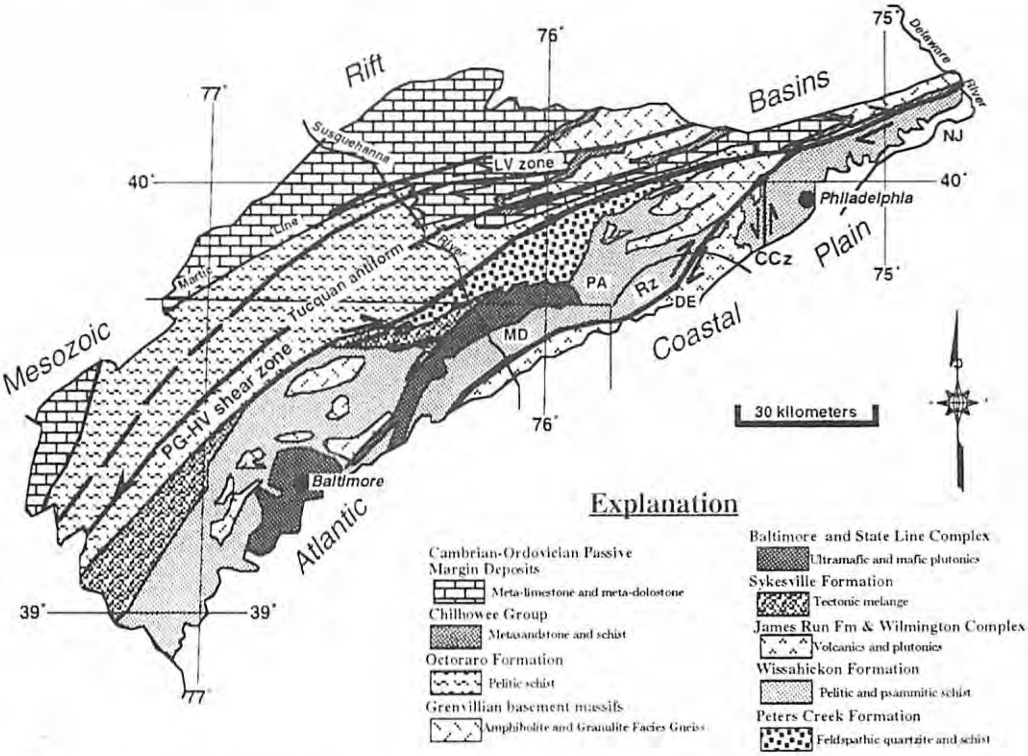


Figure 35. Generalized tectonic map for the north-central Appalachain Piedmont (modified from Williams, 1980).

The geology of the southeastern belt is more complex. On its east end, the Philadelphia, Brandywine, and White Clay structural blocks were overridden by the Wilmington complex, which probably represents the infrastructure of an island arc from Iapetus (Wagner and Srogi, 1987; Wagner and others, 1991). The metamorphism of this obduction was relatively local, although intense (Crawford and Crawford, 1980; Wagner and Srogi, 1987). To the southwest, the relations are different, and not as clear. The Grenvillian age massifs in the central Piedmont are overlain by the Setter and Cockeysville Formations, which in turn underlay, in thrust contact, the Wissahickon siliciclastic schists (Alcock, 1994). These schists, in turn, pass underneath (in thrust contact) the eastern end of the State Line mafic complex (Muller and others, 1989). On the north side of this southeast belt, a narrow outcrop band of the Peters Creek Formation is present (Figure 35: also see Valentino and Gates, this guidebook). The narrow band of this rift facies sequence (Gates and Valentino, 1991; Valentino and Gates, 1994) broadens to the southwest, where is contains the Peach Bottom structure, which has been variously interpreted as a syncline (Knopf and Jonas, 1929), an anticline (Higgins, 1972), and a segment of a regional transcurrent shear system (Valentino, 1990; Valentino and others, 1994).

The boundary between these two belts is the Pleasant Grove-Huntingdon Valley shear zone (Figure 35), a 1.5-2.0-km-wide zone of subvertical schistosity and retrograde metamorphism.

It contains fine-grained phyllonite and mylonite, which are pervaded by subhorizontal lineations parallel to the zone. The Pleasant Grove-Huntingdon Valley shear zone is a product of Alleghanian dextral transpressional tectonism (Valentino, 1993; Valentino and others, 1994).

METAMORPHISM OF THE NORTHWESTERN BELT

The northwestern metamorphic belt of southeastern Pennsylvania experienced a complex deformation history accompanied by metamorphism. This belt is structurally dominated by the Tucquan antiform (27 km width, 65 km length, and 7 km amplitude at widest point), that is defined by an arch of penetrative schistosity (S1 of Freedman and others, 1964; Wise, 1970). Freedman and others (1964) suggested that the regional schistosity and isoclinal folding was related to northwesterly directed nappe emplacement associated with the Taconian orogeny, and later folding of the regional schistosity into the Tucquan antiform was related to vertical movement of Grenvillian basement blocks. More recently, detailed investigations in this region suggest that the post-Taconian deformation history was one dominated by dextral strike-slip tectonics (Valentino and others, 1994).

Deep erosion of the Tucquan antiform exposes a sequence of metamorphic rocks ranging from schists of the chlorite zone on the antiform limbs up to the garnet zone in the core (Valentino and Faill, 1990). These Taconian metamorphic zones are symmetrically distributed about the axial trace (Figure 36). From limb to crest, the characteristic assemblages in pelitic rocks are: (1) chlorite-muscovite $\pm$ plagioclase; (2) chlorite-biotite-muscovite $\pm$ plagioclase; (3) biotite-chloritoid; and (4) chlorite-garnet $\pm$ chloritoid. Locally near the core of the antiform a small kyanite zone is exposed. High-Al bulk composition is responsible for the lack of a biotite-garnet mineral assemblage in the sequence from limb to core.

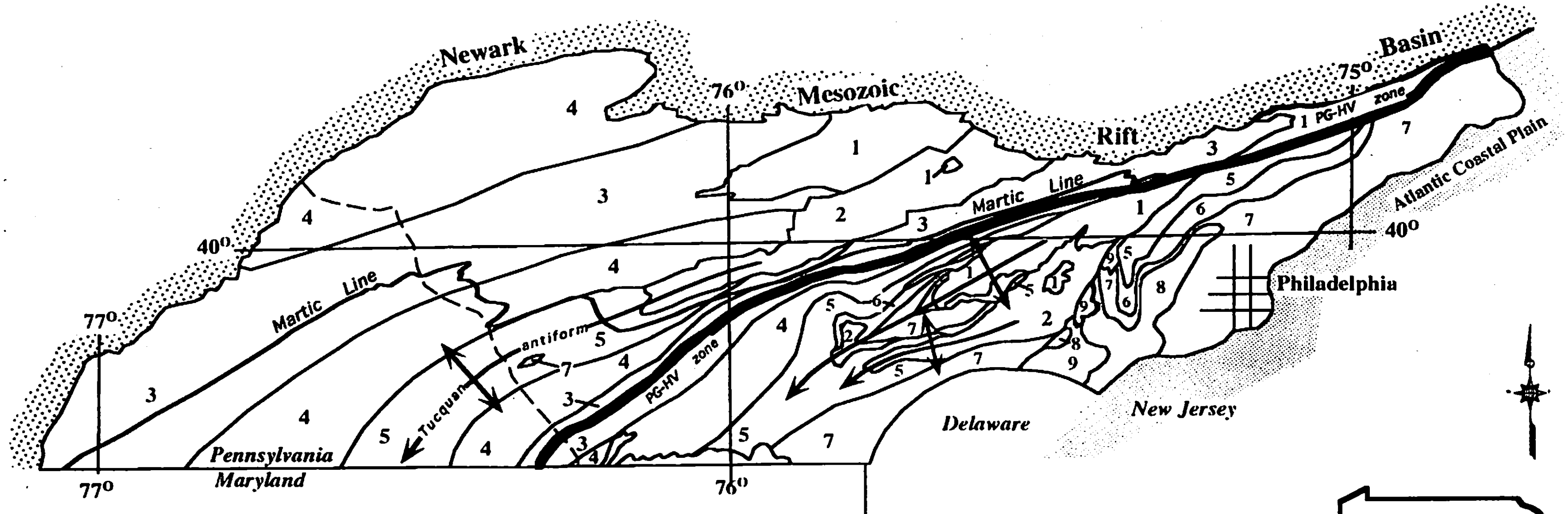
The rocks in the core of the Tucquan antiform are generally muscovite-chlorite-quartz schist with diagnostic minor phases such as chloritoid, biotite, garnet and kyanite. The approximate modal mineral percentages for these rocks were determined by point counting thin sections and estimation with color density charts (Figure 37). Five metamorphic mineral assemblages occur in the rocks of the Tucquan antiform core: (1) garnet-chlorite, (2) garnet-chlorite-biotite, (3) biotite-chlorite-chloritoid, (4) garnet-chlorite-chloritoid and (5) kyanite-chloritoid-chlorite. Although assemblage (2) was observed in some rocks, the occurrence of biotite and garnet in the same thin-section is rare (Figure 37c). Garnet is most often in association with chlorite $\pm$ chloritoid. The rare occurrence of garnet-biotite bearing schist, and common occurrence of biotite-chlorite without garnet and garnet-chloritoid-chlorite without biotite, is the result of the bulk composition of the rocks. Calculated bulk compositions for two samples plot above the garnet-chlorite join on the AFM diagram, suggesting the high aluminum content of the rocks may be a controlling factor for the absence of biotite in the presence of garnet.

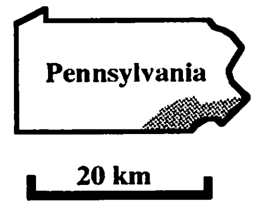
Garnet-chlorite exchange thermometry in the garnet zone

To obtain quantitative information for the conditions of metamorphism in the Tucquan antiform, various thermobarometric techniques were considered, however, the metamorphic mineral assemblages were most applicable to the garnet-chlorite exhange thermometer (Dickenson and Hewitt, 1986; Laird, 1988; Berman, 1990). Late lower greenschist facies retrograde metamorphism associated with the development of the Tucquan antiform heterogeneously affected the pervasive Taconian metamorphism that is portrayed in the metamorphic zone map (Figure 36). In general, this late retrogression is confined to discrete shear zones, but, some minor overgrowths of chlorite on primary garnet were documented (Figure 38).

Two potential problems in obtaining quantitative data from the overprinted samples are: (1) disturbance of the internal composition of the primary garnets and matrix chlorite during retrograde reequilibration, and (2) a change in the matrix chlorite composition in response to growth of secondary chlorite at the expense of the primary garnet. The digital data to generate the relative cation composition images (Figure 39) was obtained using the Cameca Microprobe in the Department of Geological Sciences at Virginia Technological Institute and State University. Garnets were imaged for CaO, FeO, MgO, and MnO to test for alteration in the internal chemical zonation. In conjunction with these qualitative images, samples V1 and V2 were calibrated with selective spot analyses to produce a quantitative data set from which

Generalized Metamorphic Zone Map of the Pennsylvania Piedmont





Explanation

Grenvillain Metamorphic Zones
1- Granulite facies gneisses
2- Amphibolite facies gneisses

Paleozoic Metamorphic Zones
3- Chlorite
4- Biotite
5- Garnet
6- Kyanite
7- First Sillimanite
8- Second Sillimanite
9- Wilmington complex granulite and amphibolite facies gneisses
10- State Line complex greenschist and amphibolite facies mafic and ultramafic rocks

modified from Valentino and Faill (1990)

Figure 36. Metamorphic zone map of the Pennsylvania Piedmont (after Valentino and Faill, 1990).

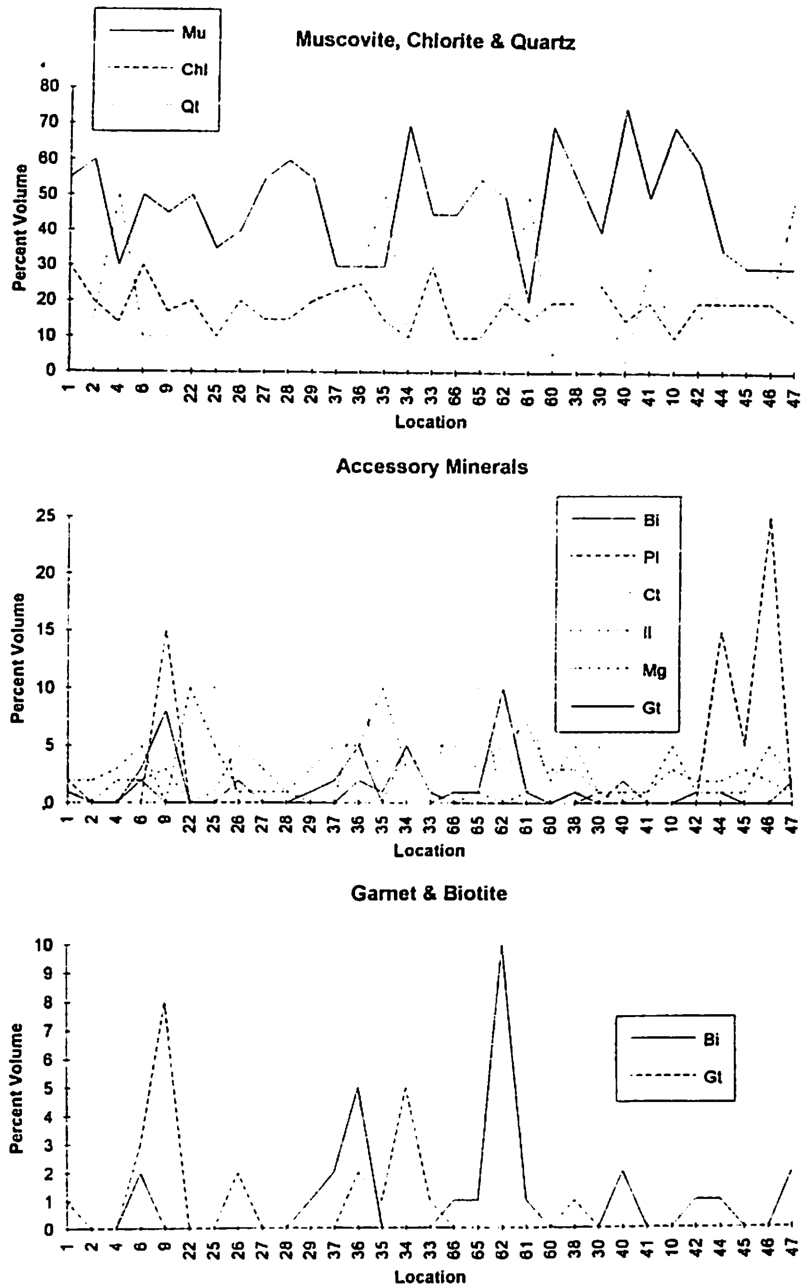


Figure 37. Graphs of the modal mineralogy for Octoraro Formation in the core region of the Tucquan antiform.

calibrated profiles were generated (Figure 39). Normal greenschist facies compositional zonation was observed using the images (Figure 38). The MnO images show typical high concentration in the core with progressively lower relative concentrations toward the garnet edges. Conversely, the FeO images show a pattern with the highest concentrations near the edge of the garnets and lowest in the cores. In general, the CaO and MgO components of these garnets are relatively low, however, the amplified calibrated profiles (Figure 39) demonstrate normal internal compositional patterns. The resorbed edges of some garnets where secondary chlorite has grown, particularly sample V1, shows truncation of the internal compositional zonation (Figure 38). The secondary growth of chlorite at the expense of garnet appears to have been a surficial process relative to the garnet, therefore the internal compositional zonation does not appear to be affected by the late lower greenschist facies overprint. The implications of this are that, although these garnets appear to be altered by patchy chlorite overgrowth, the internal chemistry is preserved, therefore, possibly permitting the use of unaltered garnet rim compositions for determination of paleo-thermal conditions.

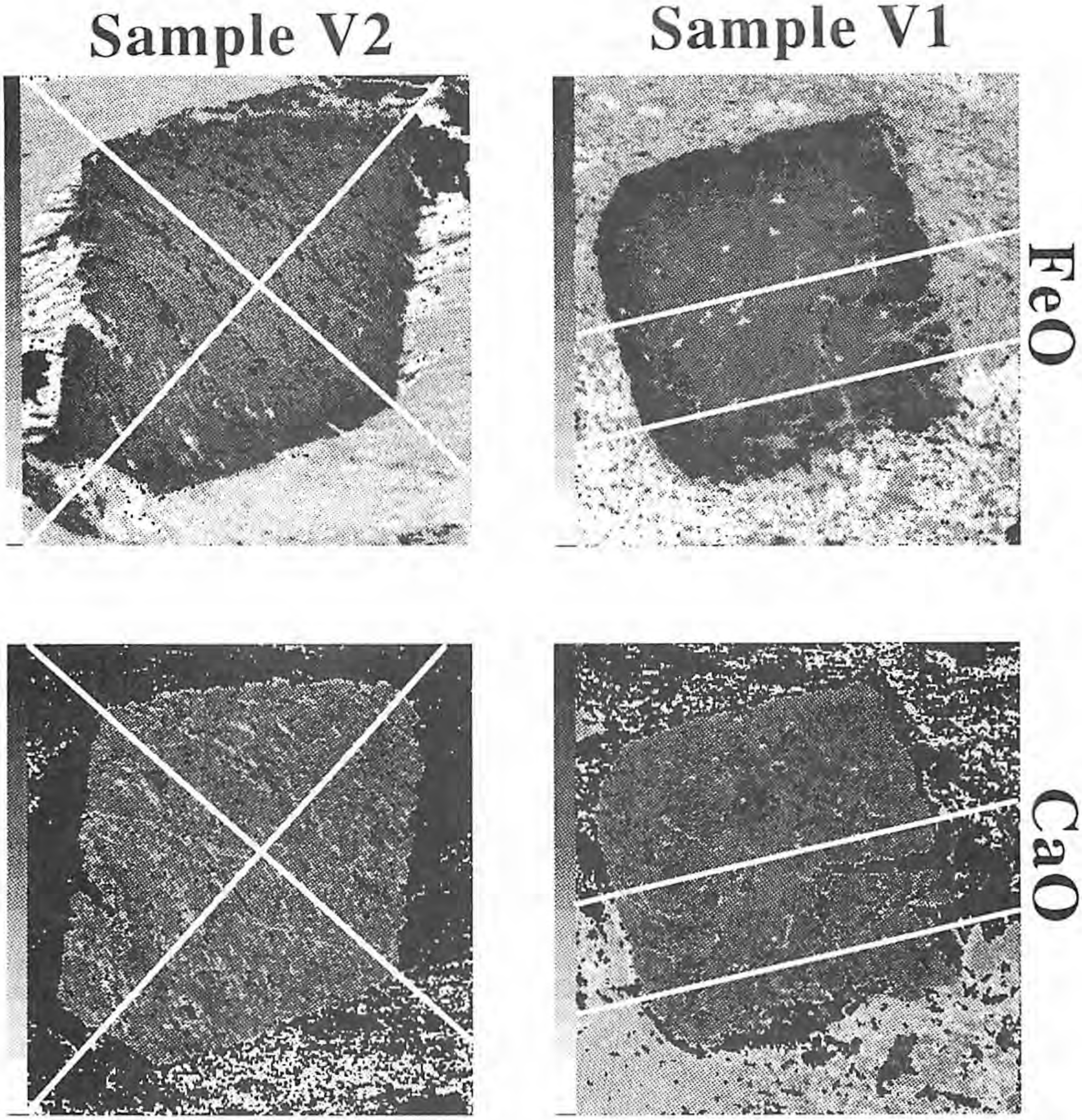


Figure 38. False gray-scale images of garnets from samples V1 and V2. Locations for these samples are provided on Figure 40.

The second part of this preliminary test was to assess the degree of primary chlorite compositional change in response to growth of secondary chlorite at the expense of garnet. Point counting thin sections from the Octoraro Formation in the core of the Tucquan antiform determined that the average ratio of matrix chlorite to garnet is approximately 20:1 in garnet bearing samples. The secondary chlorite comprises less than 1 percent of the total chlorite in the samples and microprobe spot analyses determined that the primary matrix chlorite and the secondary overgrowth chlorite have the same range of compositions for any given sample. This suggests that the addition of small amounts of secondary chlorite to the matrix had minor effect on the primary matrix chlorite composition. Therefore, the matrix chlorite compositions and garnet rim compositions were used in the garnet-chlorite exchange thermometer (Dickenson and Hewitt, 1986; Laird, 1988; Berman, 1990).

Analytical data was obtained from the garnet zone with the least amount of secondary chlorite overgrowth determined petrographically. The garnet-chlorite exchange thermometer was applied to *estimate* the variation in thermal conditions in the core of the Tucquan antiform, and summarized results are presented in Table 8. All the samples, except sample V1, resulted in average temperatures between 400°C and 525°C, and these results are reasonable for the upper greenschist facies metamorphic mineral assemblages documented. However, sample V1 produced an average temperature of 600°C, approximately 75°C higher than the other samples.

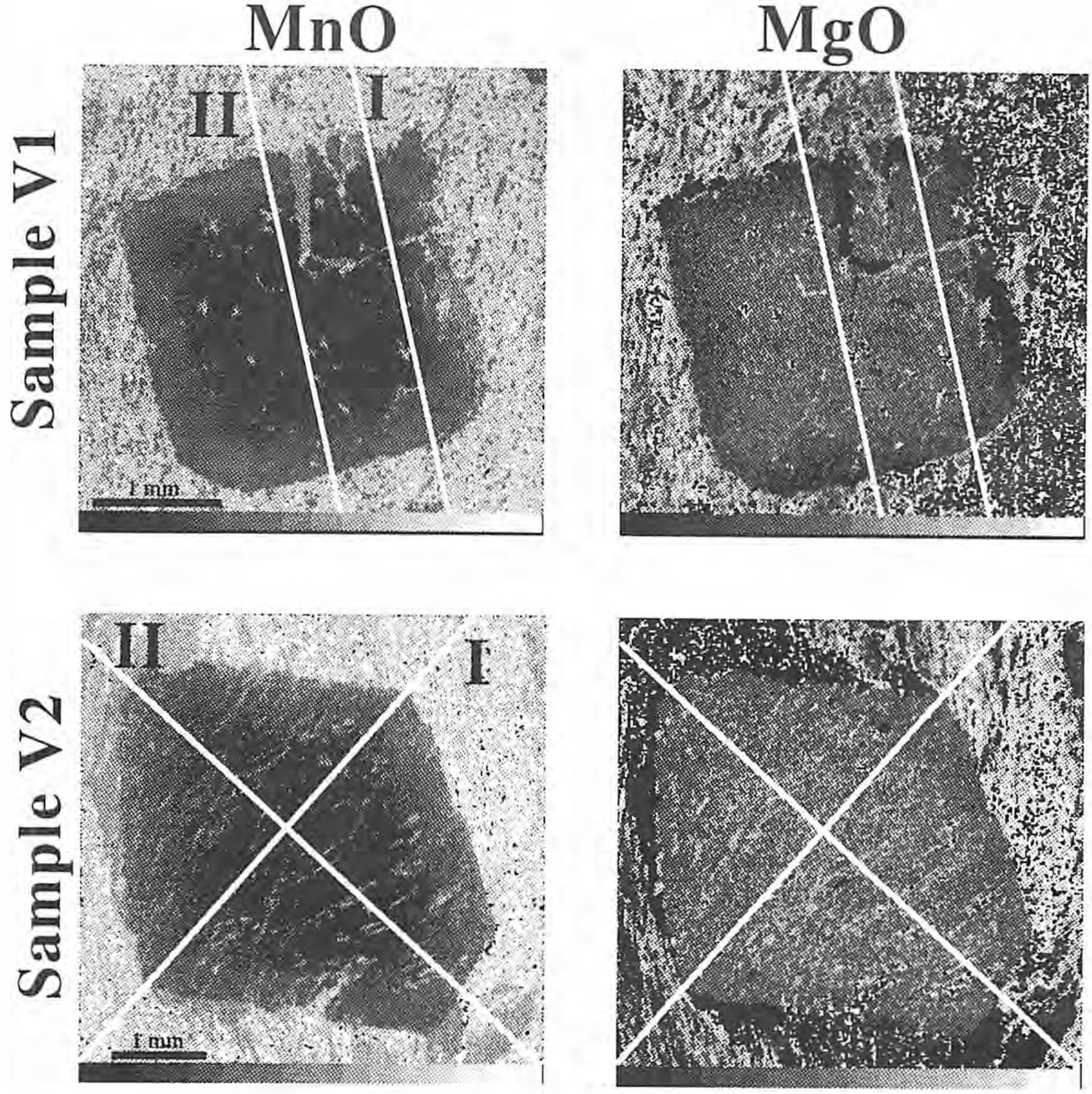


Figure 38. Continued.

At first this temperature presented a potential problem, but, a careful examination of the V1 garnet image in conjunction with the coordinates for the garnet spot analyses revealed that most of the garnet analyses for this sample were collected from portions of the garnet rim that was severely resorbed by secondary chlorite overgrowth. The ratio of FeO:MgO was noticeably lower for sample V1 as compared to samples with less overgrowth. Although the almandine component is an order of magnitude greater that the pyrope component in these garnets, near the edge of the garnet the rate of change for the almandine component versus distance along the garnet radius is much higher than for pyrope (Figure 39). Therefore, small amounts of garnet alteration would drastically effect the apparent edge composition with respect to the FeO:MgO ratio. A higher MgO:FeO ratio paired with normal chlorite compositions would produce a higher temperature, therefore possibly explaining the high temperature for Sample V1.

Thermal structure of the Tucquan antiform

The average temperature estimates were plotted on a structure map (Figure 40) of the Tucquan antiform core region and a cross section was constructed (Figure 41) perpendicular to the antiform hinge axis. Approximate isotherms based on the five average temperatures were constructed on the structure map and the cross section. Systematically the highest

Table 8. Summary of Garnet-Chlorite exchange thermometry results for samples from the Tucquan antiform core region in Lancaster County, Pennsylvania. Temperature values were calculated using an algorithm based on the garnet-chlorite exchange thermometer of Dickenson and Hewitt (1986) and associated corrections (Laird, 1988; Berman, 1990). The calculated average temperatures do not necessarily occur in the middle of the reported temperature range.

Location	Sample	Number of Calculated Temperatures	Temperature Range (^{o}C)	Average Temperature (^{o}C)
Holtwood 35	XA8	34	385-450	440
Holtwood 36	XA9	42	415-505	455
Conestoga 4	XB0	32	490-525	510
Holtwood 40	XB1	25	420-440	425
Holtwood 9	V1	55	575-625	600
Holtwood 38	V2	68	420-445	422

temperatures occur at the structurally deepest level and the lowest temperatures occur at the structurally highest level. Truncation of the Taconian structure and isotherms occurs on the northwestern limb of the antiform where a late thrust fault (Valentino, 1990; Valentino, 1993) breached the crest of the antiform to displace rocks from the garnet zone toward the northeast over rocks of the biotite zone (Figure 41). It is interesting to note that this thrust fault occurs along a segment of the Martic Line mapped by Ernst Cloos (see Cloos and Hietanen, 1946). Using the southern boundary of the garnet zone as a structural marker, the calculated temperatures were plotted against relative structural depth to approximate the Taconian thermal gradient in the core of the Tucquan antiform (Figure 42). A linear regression through the data points produced a gradient of $5.7^{o}C/100$ m of structure thickness. Although this thermal gradient is reasonable for a tectonically active region it is only a first pass estimate when all uncertainties in this analysis are considered. For example, volume loss related to metamorphism is not taken into account in this construction.

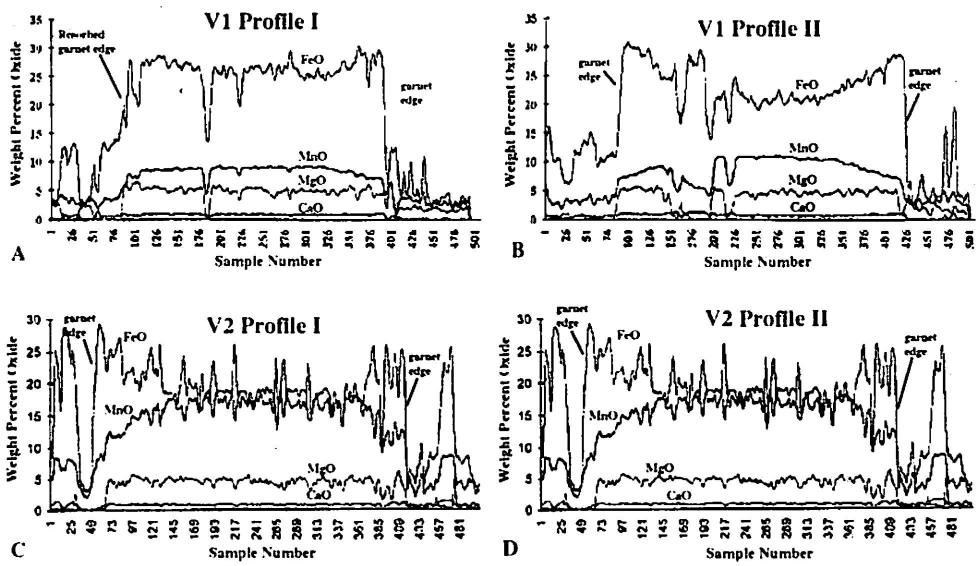


Figure 39. Calibrated chemical profiles of MnO, CaO, FeO, and MgO for samples V1 and V2.

THE BOUNDARY BETWEEN THE METAMORPHIC BELTS

The northwest and southeast metamorphic belts are separated by the Pleasant Grove-

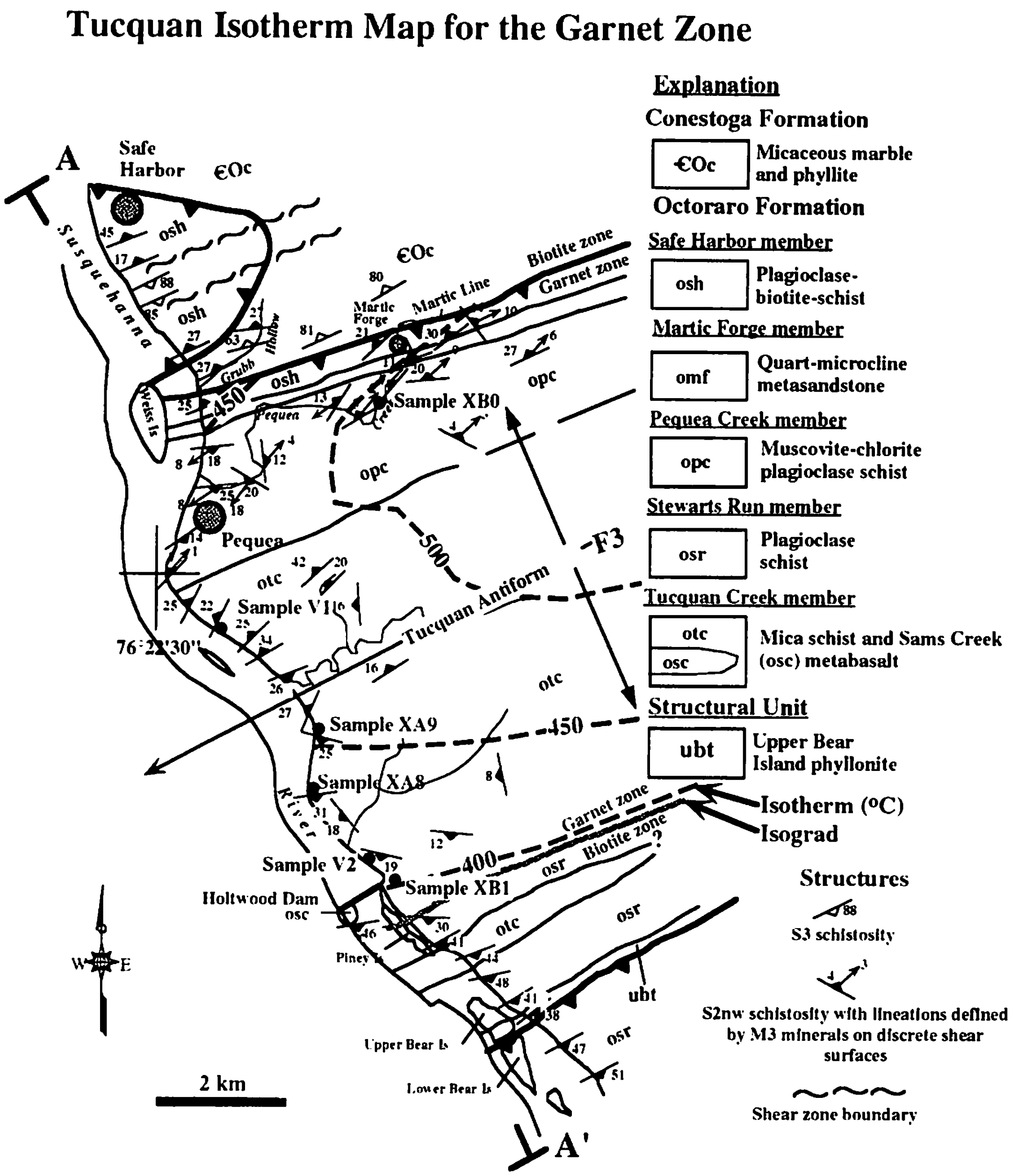


Figure 40. Structure map for the Tucquan antiform core region with approximate isotherms.

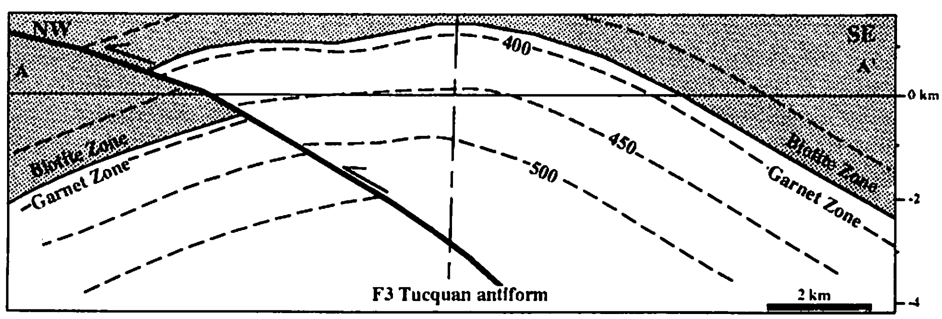


Figure 41. Cross section for the Tucquan antiform core region with approximate isotherms.

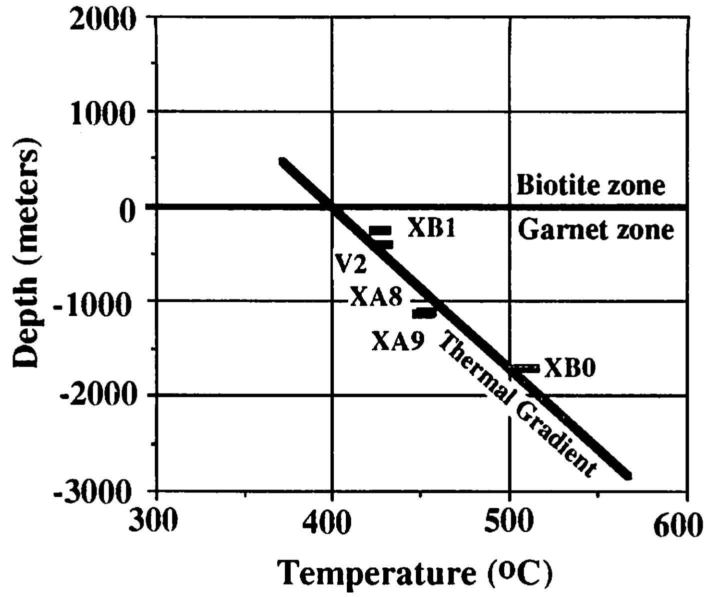


Figure 42. Plot of depth versus calculated temperature for samples from the Tucquan antiform.

Huntingdon Valley shear zone. The Pleasant Grove-Huntingdon Valley shear zone extends from the Delaware River near Trenton, New Jersey to the Culpepper Basin in northern Maryland west of Baltimore (Figure 35). The segment of the shear zone studied in detail is defined by a 2 to 3 km broad zone of steeply dipping and northeast striking S3 schistosity defined by recrystallized muscovite and chlorite, subhorizontally plunging L3 extension lineations, and upright F3 folds defined by folded Taconian schistosities. The segment of the shear zone located on the southeastern flank of the Tucquan antiform separates the pelitic schist of the Octoraro Formation (northwest of the shear zone) from the interlayered metasandstone and quartz-schist of the Peters Creek Formation (southeast of the shear zone). Farther east the shear zone forms the northern boundary of the Philadelphia block, and separates the Wissahickon schist from the Grenvillian Trenton massif (Figure 35).

The Pleasant Grove-Huntingdon Valley shear zone is defined by steeply dipping (65-90^{o}SE) S3 schistosity that cross cuts regional Taconian structures. Regional systematic variation in the strike of the S3 schistosity from 040^{o} near the Susquehanna River to 070^{o} farther east shows the gradual change in shear zone orientation that parallels the Pennsylvania reentrant. In the west, the shear zone occurs as a 1-1.5 km thick belt of silver-gray fine-grained phyllonite (locally called the Drumore tectonite), developed between the Peters Creek and Octoraro Formations (Valentino and others, 1994). The easternmost exposures of the shear zone occur as a 1-1.5 km broad mylonite zone developed in the Grenvillian-age gneisses of the Trenton massif and adjacent Wissahickon schist (Armstrong, 1941). Mineral elongation lineations in the shear zone occur as pressure fringes on porphyroclasts in most rock types (Figures 49A and 49B), elongate quartz grains and pebbles in metasandstone (Figure 43B), and chlorite and muscovite streaks in metapelite. Various types of pressure fringes are developed on porphyroclasts: (1) chlorite fringes on magnetite (Figure 43B), (2) quartz fringes on euhedral pyrite (Figure 44A), and (3) quartz-muscovite fringes on chloritoid (Figure 43C). These lineations generally plunge shallowly throughout the zone of S3 schistosity, but rare plunges of up to 35^{o} northeastward were observed in the Susquehanna River region. Over a distance more than 65 km the L3 mineral lineations show a consistently shallow plunge, and only the region near the Susquehanna River has lineations that plunge moderately northeastward.

Mineral lineations associated with the S3 schistosity in the Pleasant Grove-Huntingdon Valley zone are indicative of horizontal transport. Dextral discrete shear zones were observed in quartzite of the Peters Creek Formation in X-Z parallel outcrop surfaces (Figures 43D and 43E). Similarly X-Z sections reveal that the porphyroclasts with pressure fringes display an asymmetry consistent with dextral shear (Figures 43A, 43B, and 43C). Mylonitized quartz veins commonly have dextral preferred grain shape orientation, and conglomeratic

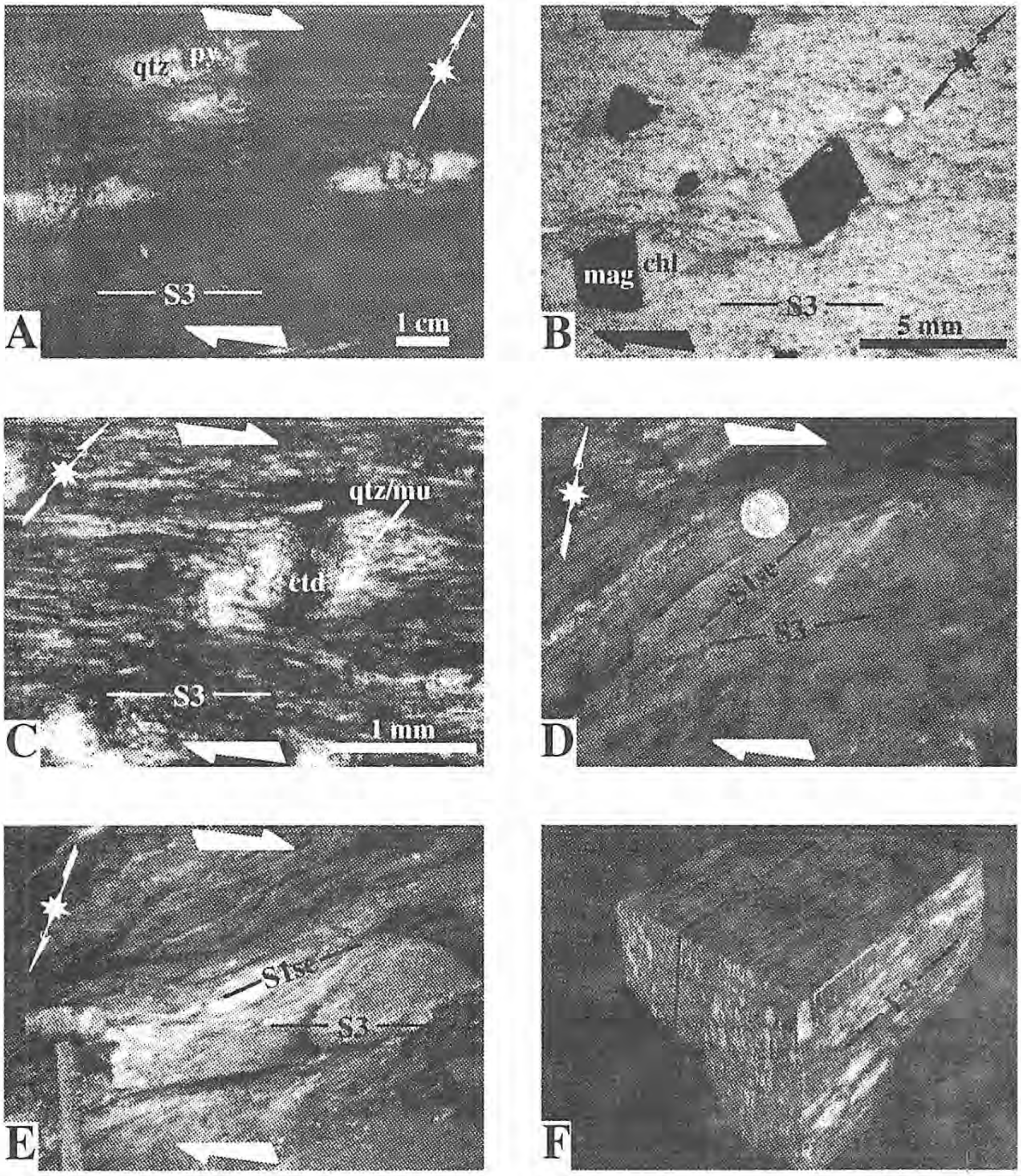


Figure 43. Deformation fabrics in the Pleasant Grove-Huntingdon Valley shear zone in Lancaster County, Pennsylvania. [A] Quartz pressure fringes developed on a pyrite porphyroclast. [B] Chlorite pressure fringes developed on magnetite porphyroclasts. [C] Quartz-muscovite pressure fringes on chloritoid porphyroclast. [D] and [E] Discrete shear zones developed in Peters Creek Formation quartzite. [F] Block of Cardiff quartzite with flattened and elongate pebbles.

quartzite contains flattened pebbles of quartz (Figure 9f). All the shear sense indicators consistently provided a dominantly dextral sense of shear along the width and length of the Pleasant Grove-Huntingdon Valley shear zone.

METAMORPHISM OF THE SOUTHEASTERN BELT

The southeast belt south of the Cream Valley-Huntingdon Valley shear zone is structurally complex, consisting of a number of separate structural blocks of diverse origin that were juxtaposed, and all of these structural blocks were thrust onto the Laurentian margin during

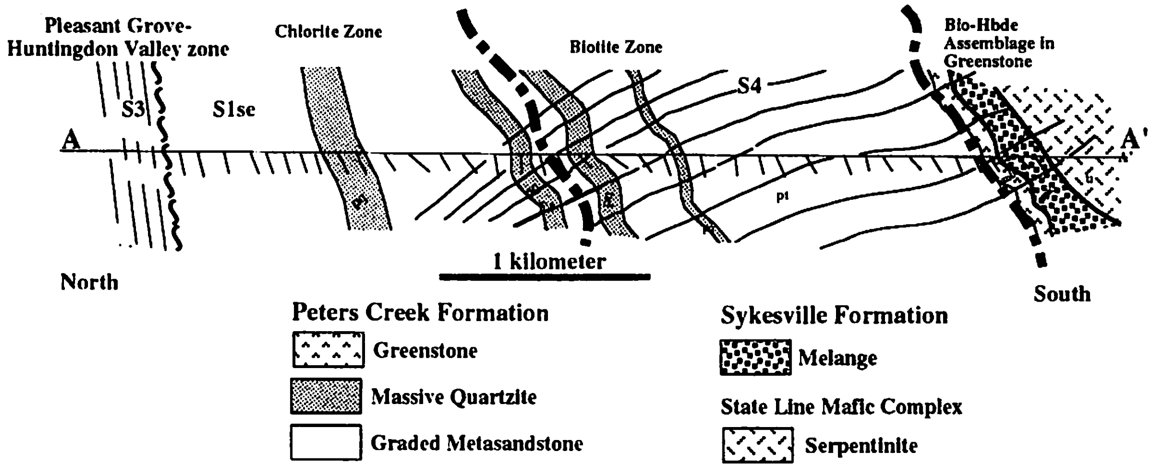


Figure 44. Schematic cross section of the Peters Creek formation at the Susquehanna River with metamorphic zones shown.

the Taconian orogeny. The Taconian metamorphic pattern is largely a consequence of this tectonic stacking.

The Southeast Metamorphic Belt at the Susquehanna River

Immediately south of the Tucquan antiform in the region of the Susquehanna River the Peters Creek Formation is well exposed. The Peters Creek Formation contains a layer-parallel schistosity (S1se) that occurs as a monoclinal structure that strikes about 045^{o} and dips moderately to steeply (50-80^{o}) southeast. From northwest to southeast, structurally lowest to highest, the metamorphic mineral assemblages that occur in the Peters Creek Formation are: (1) chlorite-muscovite, and (2) chlorite-biotite-muscovite. In the structurally highest portions of the Peters Creek Formation metabasalts contain the assemblage chlorite-biotite-hornblende. These relationships are illustrated in the cross section of Figure 44. This distribution of metamorphic mineral assemblages in conjunction with the mapped structures indicates that the metamorphic grade increases at higher structural levels.

The Southeast Metamorphic Belt in Eastern Pennsylvania

The central and eastern end of the southeast metamorphic belt comprises the Philadelphia block, the Wilmington complex, and the Brandywine and White Clay structural blocks (Figures 45A and 45B). The Peters Creek sequence and the State Line complex form the western end of this belt. Under current interpretation, all four of the eastern blocks were outboard of the Laurentian hinge zone during the late Neoproterozoic and much of the Lower Paleozoic, occupying or being formed in separate parts of the Iapetan Ocean. The convergence and collapse of the western Iapetus during the Middle and Late Orodvician caused each of these blocks to be obducted onto the continental margin, creating a structural stacking with the lower elements in the northern part of the southeast belt, and the highest parts on the south side.

The northernmost, and structurally deepest, element is the Brandywine structural block, an aggregate of three massifs of Mesoproterozoic age along with (where present) the unconfomably overlying Setters quartzite and schist, and Cockeysville marble. Whether the Brandywine Grenvillian massifs originally were detached microcontinents within Iapetus as portrayed for the Baltimore Grenvillian gneiss massifs by Fisher and others (1979) and Muller and Chapin (1984), or were connected to some other part of Laurentia, as suggested by Valentino and others (1994), is presently moot. The Brandywine is overlain, in thrust contact on three sides (Figure 45B) by the Glenarm Wissahickon schists and gneisses of the White Clay structural block (Alcock, in press). This block consists of a siliciclastic sequence with intercalated basaltic flows, deposited either in a deep basin (Iapetus) offshore of the Laurentian rifted margin, or in a back-arc or fore-arc basin associated with the Taconian

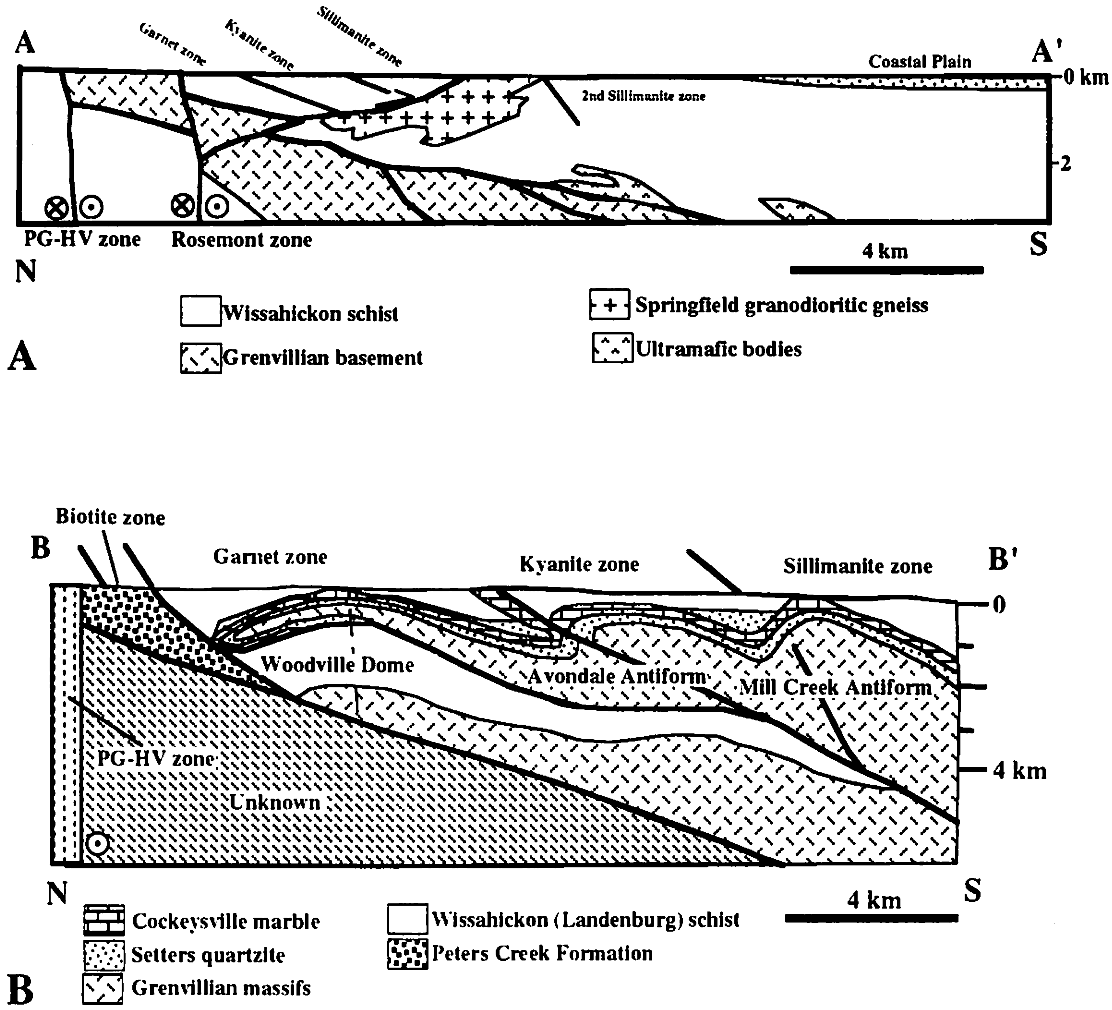


Figure 45. Generalized cross sections of the southeastern metamorphic belt (after Faill, in press) with metamorphic zones represented. **[A]** The Philadelphia block, and **[B]** The White Clay and Brandywine blocks.

magmatic arc.

The Brandywine and White Clay blocks are presently joined to the Philadelphia block across the Rosemont shear zone, a post-Taconian subregional shear zone (Valentino, 1988; Valentino and others, in press). The nature of the pre-shear contact is uncertain, but the present contact geometry, defined by the Rosemont shear zone, was strongly influenced by transcurrent faulting (Valentino and others, in press). Presumably the Philadelphia block was thrust over the Brandywine at an earlier time (Wagner and Srogi, 1987), and the stacking relation between the White Clay and Philadelphia blocks is unknown. The Philadelphia block, consisting mostly of Wissahickon Formation pelitic and psammitic schist and gneisses, was apparently a deep basin siliciclastic sequence that was subsequently intruded with intermediate to felsic bodies, such as the Springfield granodioritic gneiss.

The Wilmington complex was thrust (Wagner and Srogi, 1987) over the Philadelphia and White Clay blocks, and thus represents the structurally highest component of the southeast metamorphic zone in Pennsylvania (Figure 45A). The complex, containing mafic and felsic gneisses intruded by mostly gabbroic and intermediate plutons, is considered part of the infrastructure of a Cambrian (?) magmatic arc (Crawford and Crawford, 1980; Wagner and Srogi, 1987; Wagner and others, 1991).

The relation between the White Clay block and the Peters Creek Formation rift-related sequence (Valentino and Gates, in press; Valentino and Gates, this guidebook) is not well delineated, but mapping in the contact region immediately north of the State Line mafic complex (A. E. Gates, personal communication, 1987) suggests a composition gradational contact between the two units.

The Metamorphism

In general, the metamorphic grade in the southeast belt increases to the southeast, from the structurally lowest- to the highest-level structural block (Figure 36). This pattern is in distinct contrast to the symmetric pattern of the northwest belt. The metamorphic zones in the northernmost part of the southeast belt are exceptionally narrow, indicating a strong, northwest-southeast metamorphic gradient across a short distance. These zones possibly represent retrograde assemblages formed as a consequence of shearing and recrystallization within the Pleasant Grove-Huntingdon Valley shear zone during the Alleghanian orogeny.

The Grenvillian gneisses of the Brandywine terrane immediately to the south were affected by the Taconian metamorphism that replaced granulite facies minerals with an upper amphibolite assemblage, and this metamorphism occurred at the depth range of 33 to 38 km (Wagner and Crawford, 1975). A metamorphic discontinuity exists between the "Glenarm Wissahickon" schist surrounding the western end of the Brandywine block (Avondale massifs of the Grenvillian gneisses and the lower Glenarm rocks including the Setters and Cockeysville Formations (Alcock, 1994)). This discontinuity is a metamorphic inversion, reflected by upper amphibolite facies "Glenarm Wissahickon" rocks overlying Cockeysville marbles that exhibit a lower temperature assemblage (Alcock, 1989; 1994). This inversion indicates that the "Glenarm Wissahickon" schists were thrust over the Brandywine block after peak metamorphism. The "Glenarm Wissahickon" schists south of the Brandywine block exhibit kyanite- and staurolite-bearing assemblages overprinting andalusite- and cordierite-bearing assemblages. This suggests that an earlier, moderately shallow-depth metamorphism of this southern part of the White Clay block was followed by a second metamorphism much deeper in the crust, at a depth of as much as 25 to 30 km (Crawford and Mark, 1983).

The Wilmington complex consists of foliated gneisses exhibiting granulite facies assemblages; these gneisses were intruded by non-foliated gabbroic and granodioritic plutons which cooled under granulite-facies conditions, at interpreted depths of 21 to 28 km (Wagner and others 1991). The schists and gneisses of the underlying "Glenarm Wissahickon" just to the north and west of the complex, and of the Wissahickon to the east, were metamorphosed at conditions above the second sillimanite isograd. Metamorphic grade decreases away from the Wilmington complex, with the implication that the complex was thrust over the adjacent structural blocks while it was itself very hot (Crawford and Mark, 1983; Wagner and Srogi, 1987; Wagner and others, 1991).

The southeast belt (central and eastern part) displays a metamorphic and tectonic pattern quite different than that of the northwest belt. Rather than the symmetric pattern as found in the northwest belt, the metamorphic grade increases, and the structural level becomes higher, to the southeast. This metamorphic gradient was produced largely by the tectonics of the belt. The successively higher facies to the southeast were a result of the obduction and stacking of structural blocks, particularly the hot Wilmington complex.

TECTONIC MODELS AND CONCLUSIONS

The pattern of metamorphism for the northwest belt suggests a heat source from below. Potential tectonic models to explain this are: (1) burial metamorphism associated with the development of the Laurentian Passive Margin, (2) delamination of the eastern edge of the lower Laurentian lithosphere as it enters the subduction zone below the Taconian magmatic arc resulting in an aesthenospheric rise and increased geothermal gradient (a major heat source from below) (Glover and others, 1992), and (3) oblique pre-Taconian subduction and oblique collision would cause localized transtensional deformation and possibly develop pull-apart basins providing a rift-related heat source to the Laurentian passive margin (Glover and others, 1992), prior to obduction of the magmatic arc.

Burial metamorphism as a mechanism to explain a heat source from below is problematic because the metamorphism in the core of the Tucquan antiform occurred synchronously with deformation (D1 of Freedman and others, 1964; Wise, 1970; Valentino and others, in press), and deformation of this scale and intensity would not be expected in a passive margin burial metamorphic environment. The delamination model was proposed by Glover and others (1992) to explain metamorphic relationships in the western Piedmont of Virginia and North Carolina Blue Ridge cover sequences, as well as the occurrence of 500 Ma mafic and felsic plutons that apparently intruded the Laurentian crust. Although a model of localized transtensional extension as a mechanism for a heat source from below is considered, it is not likely since no Taconian transtensional structures have yet to be documented in the central Appalachian Piedmont.

Published tectonic models of Wagner and Srogi (1987) and Wagner and others, (1991) explain the metamorphism of the southern belt as the result of subduction of the Laurentian passive margin beneath the Taconian magmatic arc containing the Wilmington complex. Regional Taconian metamorphic patterns in the southeastern belt are consistent with this model. The grade of metamorphism generally increases from northwest to southeast, with the exception where later thrusting and folding occurred near the Grenvillian basement massifs. This pattern of increasing grade toward the southeast corresponds to increasingly higher structural levels toward the interpreted magmatic arc.

TERRACES, FLUVIAL EVOLUTION, AND UPLIFT OF THE LOWER SUSQUEHANNA RIVER BASIN

Frank J. Pazzaglia[1] and Thomas W. Gardner[2]
(1) Department of Earth and Planetary Sciences
University of New Mexico
Albuquerque, NM 87131
(2) Department of Geosciences
The Pennsylvania State University
University Park, PA 16802

ABSTRACT

Lower Susquehanna River fluvial terraces offer a unique opportunity to investigate the late stage geologic and geomorphic evolution of the U.S. Atlantic passive margin. Petrography and elevation distinguish and provide a basis of correlation for two groups of terraces, the upland terraces and lower terraces, through the Piedmont, Newark-Gettysburg Basin, and Great Valley. Downstream correlation to dated upper Coastal Plain and Fall Zone fluvial deposits, relative weathering, and soil profile development characteristics establish terrace age. Upland terraces (Tg1, Tg2, and Tg3), middle to late Miocene strath terraces 80 to 140 m above the present channel, occur only along the Piedmont reach. They are underlain by unstratified, texturally-mature, quartz-dominated, roundstone diamictons. In contrast the lower terraces (QTg, Qt1 - Qt6), Pliocene and Pleistocene strath and thin aggradational terraces within 45 m of the present channel, are underlain by stratified and unstratified, texturally and compositionally immature sand, gravel, and pebbly silt.

Terrace age and longitudinal profiles suggest complex interactions between relative base level, long-term flexural isostatic processes, climate, and river grade. A model for terrace genesis requires the Susquehanna River to attain and maintain a characteristic graded longitudinal profile over graded time. For the U.S. Atlantic margin, the model proposes that straths are continually cut along this graded profile during periods of relative base level stability, achieved by slow, steady, isostatic continental uplift acting in concert with eustatic rise. Change in an external modulating factor, such as eustatic fall or climate change, results in fluvial incision and subsequent genesis of strath terraces. Convex-up longitudinal profiles of lower Susquehanna River terraces, which converge at the river mouth, diverge through the Piedmont, and reconverge north of the Piedmont, stand in contrast to their hypothesized, original concave-up profiles. Progressive and cumulative flexural upwarping of the Atlantic margin accounts for terrace profile deformation suggesting flexural isostasy as a first-order, regional deformation mechanism. These results offer new interpretations of terrace age, correlation, and geologic significance that require modification of previous studies suggesting uplifted or anticlinally-warped peneplains on the U.S. Atlantic margin.

INTRODUCTION

Fluvial terraces preserved along the lower reaches of the Susquehanna River, record the late-stage geologic and geomorphic evolution of the U.S. Atlantic passive margin (Pazzaglia and Gardner, 1993). Early studies, driven by the need to understand peneplain genesis and uplift (Davis, 1889; Barrell, 1920; Bascom, 1921; Knopf, 1924; Stose, 1928; Knopf and Jonas, 1929; Ashley, 1930, 1933; Campbell, 1933; Hickok, 1933), recognized the geomorphic importance in identifying, correlating, and dating these terraces (Wright, 1892; Bashore, 1894, 1896; Stose, 1928, 1930; Campbell, 1929, 1933; Jonas and Stose, 1930; Ashley, 1933; Hickok, 1933; Leverett, 1934; Mackin, 1934; Stose and Jonas, 1939; Peltier, 1949). Previous terrace correlations suggested uplifted, seaward-sloping peneplains (Stose, 1928, 1930), a broad northeast-southwest trending elongate dome called the Westminster Anticline (Campbell, 1929, 1933), or erosion surfaces graded to tectonically and eustatically-generated knickpoints (Hickok, 1933). These incongruous interpretations arose because: (1) systematic petrographic and textural criteria were not used to identify and correlate fluvial terraces; (2) no distinction or genetic relation was made between depositional and erosional fluvial features;

(3) a relationship linking terrace genesis to the complex interaction between passive margin isostatic, eustatic, and climatic processes was not established; and perhaps most importantly, (4) terrace age was generally poorly constrained.

Recently, fluvial deposits at the mouth of the Susquehanna River have been dated by stratigraphy and petrography-based downdip correlations to marine deposits in the Salisbury Embayment (Pazzaglia, 1993). Within the context of this new age control, fluvial terraces along the lower Susquehanna River could be mapped, correlated, and dated. The terraces could then be used to develop a model for terrace genesis on a U.S. Atlantic-type passive margin suggesting the nature, magnitude, and rates of passive margin neotectonic deformation (Pazzaglia and Gardner, 1994).

Terraces were mapped at a scale of 1:24,000, primarily from field identification of fluvial deposits, in Cecil and Harford Counties Maryland, and York and Lancaster Counties, Pennsylvania. Some terraces less than 30 m above the modern channel retain a characteristic terrace morphology and were also mapped, in part, by air photo interpretation. Fluvial deposits underlying terraces vary from stratified sand, gravel, and pebbly-silt several meters thick, to a roundstone diamicton 1-3 m thick, to a few surficial lag clasts per square kilometer. A clast density of at least 1 clast per 100 m^2 is required for the definition of a mappable terrace. Petrography, relative weathering characteristics, and to a lesser degree, elevation define the terrace correlation. Downstream petrographic correlation to dated fluvial deposits on the upper Coastal Plain and Fall Zone estimates age for terraces greater than 30 m above the modern channel. Field-based relative soil profile development and upstream correlation to glacial deposits estimates age for terraces less than 30 m above the modern channel.

RIVER PROFILE

The Susquehanna River flows through a broad, moderately-incised valley underlain by steeply-folded Paleozoic and Mesozoic sandstone, siltstone, shale, and limestone of the Ridge and Valley, Great Valley, Gettysburg Basin, and Low Piedmont Physiographic Provinces, falling 65 m over 130 km from Northumberland to Columbia, Pennsylvania, for an overall gradient of 0.0005 (Figure 46). Along this stretch, the longitudinal profile is relatively straight exhibiting few knickpoints except for those lithologically controlled by resistant sandstone or quartzite. As the river enters the amphibolite-grade gneiss, schist, phyllite, and quartzite underlying the High Piedmont, its valley abruptly narrows and deepens to a 180 m-deep steep-walled gorge flanked by a gently-rolling upland with less than 30 m of local relief (Figures 46 and 47). Here the longitudinal profile is strongly convex, falling 70 m in 70 km for an average gradient of 0.001. Local lithologic and/or structural features may accentuate major knickpoints through the High Piedmont. Longitudinal profiles of Piedmont tributary streams exhibit similar convexities and knickpoints in their lower reaches and a concave or straight profile for their middle and upper reaches. The relative proportion of convexity in tributary-stream longitudinal profiles is greatest for streams in the Holtwood gorge region and tends to decrease upstream (Thompson, 1990).

An interpretation of a straight-line projection of the longitudinal profile across the deeply-incised Piedmont reach (Figure 46, curve 1) suggests that the Susquehanna River was previously graded to a base level at least 50 m above present (Reed, 1981). An interpretation of an exponential curve fit to the entire profile (Snow and Slingerland, 1987; Ohmori, 1991), with the origin near the confluence with the Juniata River, suggests that the river was graded to a base level about 62 m above present (Figure 46, curve 2). Both projections across the Piedmont convexity intersect the middle Miocene, paleo-Susquehanna River Bryn Mawr Formation on the upper Coastal Plain (Pazzaglia, 1993). Absence of significant late Cenozoic tectonic uplift for the upper Coastal Plain at the head of Chesapeake Bay (Pazzaglia and Gardner, 1994) is interpreted to mean that the exponential profile (Figure 46, curve 2) represents a reasonable middle Miocene paleo-longitudinal profile of the Susquehanna River.

LOWER SUSQUEHANNA RIVER FLUVIAL DEPOSITS

Petrography and landscape position define two groups of terraces, the upland terraces (Tg1, Tg2, and Tg3) occurring between 80 and 140 m above the present channel, and the lower

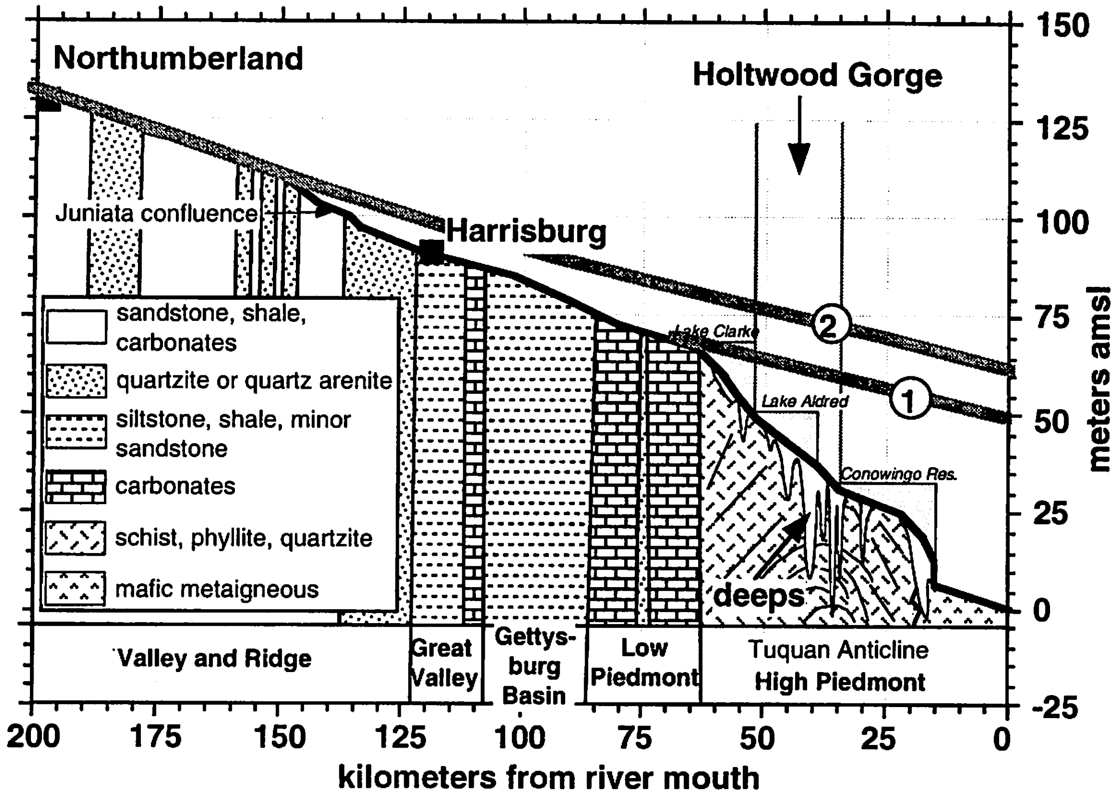
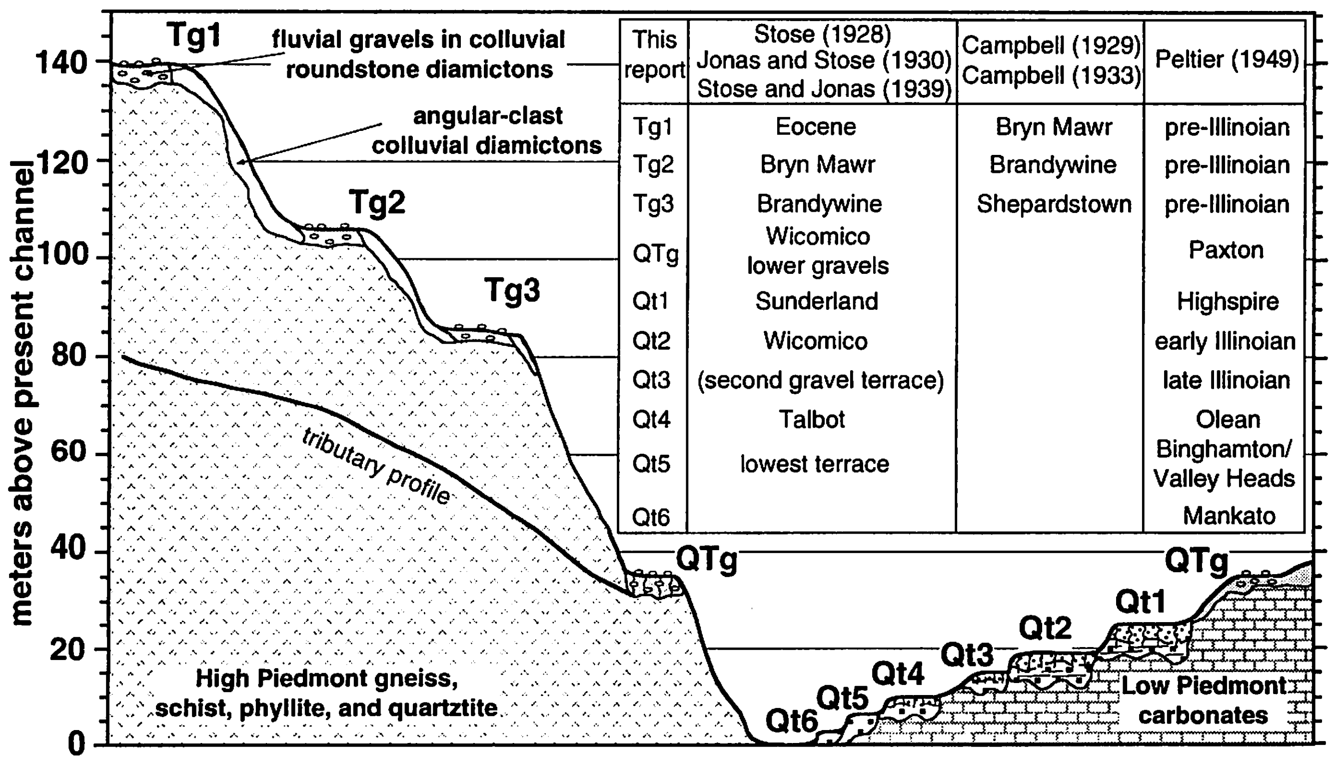


Figure 46. Longitudinal profile of the Susquehanna River from Northumberland to the head of Chesapeake Bay. Spoon-shaped, discontinuous thalweg "deeps" (Mathews, 1917) are illustrated in the Piedmont reach. Curves 1 and 2 represent a straight line and exponential projection, respectively, of the upper profile from the confluence with the Juniata River across the lower profile convexity.



This report	Stose (1928) Jonas and Stose (1930) Stose and Jonas (1939)	Campbell (1929) Campbell (1933)	Peltier (1949)
Tg1	Eocene	Bryn Mawr	pre-Illinoian
Tg2	Bryn Mawr	Brandywine	pre-Illinoian
Tg3	Brandywine	Shepardstown	pre-Illinoian
QTg	Wicomico lower gravels		Paxton
Qt1	Sunderland		Highspire
Qt2	Wicomico		early Illinoian
Qt3	(second gravel terrace)		late Illinoian
Qt4	Talbot		Olean
Qt5	lowest terrace		Binghamton/ Valley Heads
Qt6			Mankato

Figure 47. Composite cross section of the lower Susquehanna River valley showing the relative location and elevation of terraces.

terraces (QTg and Qt), lying within 45 m of the present channel (Figures 47 and 48). Upland terraces occur only along the Piedmont reach and are underlain by texturally and compositionally mature colluvial roundstone diamictons on relatively flat interfluve surfaces. Lower terraces occur both within and north of the Piedmont reach and are underlain by stratified and unstratified, texturally and compositionally immature sand and gravel. Only terraces Qt1 - Qt6 north of the High Piedmont exhibit well-defined terrace surfaces underlain by stratified sand, gravel, and pebbly-silt deposits.

Upland Terraces

Upland terraces (Tg1, Tg2, Tg3) are degraded strath terraces misidentified in early studies as peneplains mantled with upland gravels (Stose, 1928, 1930; Campbell, 1929, 1933; Figure 47). Upland terrace deposits almost always occur within 1 km of the river, rarely separated by more than 2 km along river length on flat or very gently sloping (< 5% slope) interfluves. Various origins for the colluvial roundstone diamictons underlying upland terraces have been proposed including: (1) residuum currently forming in the saprolite weathering profile; (2) surficial lag gravels derived from an older, now completely stripped sedimentary deposit such as the Cretaceous Potomac Group; (3) clasts rounded by transport during hillslope colluvial processes; (4) till, or other glacially-derived sediment; or (5) anthropogenic artifacts. The spatial geometry of upland terraces (Figures 2 and 3), their distribution at three discrete elevations above the present channel and characteristic petrology (Tables 9 and 10), affirms a paleo-Susquehanna River fluvial origin for these deposits.

Clasts in upland terrace roundstone diamictons are composed of subangular to very well-rounded vein quartz, metamorphic quartz, and quartzite, with lesser amounts of quartz arenite sandstone, sandstone, siltstone, and chert pebbles ranging in diameter (b-axis) from 2 to 20 cm with boulders up to 50 cm in diameter (Figure 49; Tables 9 and 10). Angular, 2-30 cm, hematite-cemented, imbricate, quartz pebbles and quartz sand ironstone clasts also occur. Specific terrace petrography varies with elevation. Tg1 exhibits only vein and metamorphic quartz and less than 15% quartzite and chert. Clasts at the Kirk Farm locality (Figure 48), comprise approximately 50% polycrystalline metamorphic quartz (relatively equal amounts of smokey/dark and clear/light varieties), 35% monocrystalline vein quartz, and 15% quartzite derived from local Piedmont sources such as the lower Cambrian Chickies Quartzite or from more distant sources such as the Tuscarora Formation (Table 9). Deposits underlying upland terraces Tg2 and Tg3 exhibit more sandstone and quartzite clasts derived from the Ridge and Valley (Tables 9, 10). For example, terrace Tg3 at the Brinton Farm locality (Figure 48) exhibits approximately 35% Piedmont vein and metamorphic quartz and 65% sandstone, quartzite, red siltstone, and chert (Table 9).

The soil developed through upland terrace roundstone diamicton deposits suggests multiple episodes of colluvial deposition and subsequent pedogenesis. Field mapping reveals the source for the Kirk Farm roundstone diamicton as a ridge capped by broken ironstone clasts approximately 6 m above and 100 m west of the deposit. The relatively small total distance of transport, both horizontally and vertically at the Kirk Farm locality is typical of most upland terrace deposits and consistent with the amount of colluvial transport observed for weathered bedrock and saprolite elsewhere in the Piedmont (Pazzaglia and Cleaves, in press).

Lower Terraces

Fluvial deposits of moderate textural maturity and heterolithic composition below the upland terraces define the lower terraces, QTg and Qt1 through Qt6 (Figures 47 and 48; Tables 9 and 10). Within the gorge, lower terraces are best preserved at the confluence of large tributary streams and the Susquehanna River within 50 m of the present channel. North of the Piedmont, the lower terraces define a stair-step like sequence of strath and fill-top deposits capping limestones, siltstones, and shale interfluves. These deposit are particularly well preserved at Washington, Marietta, and Middletown, Pennsylvania (Figure 48).

QTg deposits consist of subangular to rounded clasts ranging from 2 to 8 cm in diameter (b-axis) (Table 10) with occasional boulders up to 1 m across. Clast composition ranges from nearly all quartz and quartzite to more heterolithic assemblages dominated by sandstones,

Table 9. Clast identifications[1] for lower Susquehanna River terraces and selected upper Coastal Plain and Fall Zone fluvial deposits.

TERRACE OR COASTAL PLAIN DEPOSIT	NAME AND LOCATION [2]	vein quartz	meta quartz[3]	quartzite	sandstone siltstone chert limestone	ironstone granite gneiss schist
	Qt4, at 3 km south of Washington	3.5	-	19	74	3.5
lower terraces	Qt4, at Marietta (Peltier, 1949)	2	-	12	81	5
(Qt1 - Qt6)	Qt2, at Highspire	1	2	31	64	2
	Qt2, at Marietta (Peltier, 1949)	5	-	17	75.5	2.5
and	Qt1?, at Coudon Farm	15	40.5	30.5	12	2
	Pensauken, at Turkey Point	18.5	39.5	16	22.5	3.5
Pensauken Fm.	Pensauken, at Turkey Point	7	20	10	54	9
	Pensauken on Delmarva (Jordan, 1964)	46	-	36	16	2
lower terraces	QTg, at Broad Creek	16	16	41	24	3
(QTg)	QTg, combined count for entire terrace	25	12	37.5	25	<1
and	Perryville, at Mountain Hill	12	45	41	1	1
Perryville fm.	Perryville, north of Havre de Grace	10	15	31.5	31.5	12
	Tg3, at Brinton Farm, Lancaster Co., PA	16	21	52.5	10.5	<1
Upland terraces	Tg2, combined count for entire terrace	26	15	55	4	<1
(Tg1 - Tg3)	Tg1, at Kirk Farm, Lancster Co., PA	33	51.5	14.5	<1	1
	Tg1, combined count for entire terrace	5	55	30	9	1
and	Bryn Mawr, at York quarry, (2 - 4 cm)	18.5	42	37.5	2	<1
	Bryn Mawr, at York quarry, (4-10 cm)	25	30	40	4	1
Bryn Mawr Fm.	Bryn Mawr, Fall Zone, (Owens, 1969)	95	-	<1	3	2
	Bryn Mawr, on Elk Neck Peninsula	29	61	2	5	3

1 = clast size: 2-8 cm unless otherwise specified
2 = see Figure 3 for locations.
3 = dash means that vein and metamorphic quartz were not separately identified. Total non-quartzite quartz is listed in the vein quartz column.

siltstones, and local Piedmont lithologies (Table 9). A typical composition is 50% quartzite and sandstone, 30% vein and metamorphic quartz, and 20% red siltstone, chert, and locally-derived lithologies such as Piedmont schist, phyllite, metagreywacke, and slate (Tables 9 and 10).

The only known stratified QTg deposit, the Paxton terrace, occurs at Harrisburg, Pennsylvania (Peltier, 1949). The Paxton terrace contains exotic lithologies such as granite and gneiss (Peltier, 1949) and exhibits a 6-m thick deeply-weathered, well-developed, red (2.5 YR) argillic horizon through a matrix-supported diamicton and stratified gravel and sand (Peltier, 1949). Red, matrix-supported QTg roundstone diamictons also occur near the terminus of Johnson Road along the northern slope of the Muddy Creek-Susquehanna River confluence, and at the Otter Creek campground along the south side of the Otter Creek-Susquehanna River confluence (Figure 48). Both diamictons exhibit a ~2 m-thick, deeply-weathered, truncated, red (2.5 YR), well-developed, argillic horizon that grades down into weathered bedrock or saprolite. In both cases, a brown, matrix-supported, angular-clast, colluvial diamicton exhibiting a late Pleistocene soil profile (Pazzaglia and Cleaves, in press) unconformably overlies the red roundstone diamictons. Clasts within the red roundstone diamictons exhibit weathering rinds > 1 cm thick and are often completely saprolitized.

Terraces Qt1-Qt6 generally contain from 1 to 6 m of stratified, but poorly-sorted yellow, tan and buff, coarse sand, gravel and boulders up to 2 m across interbedded with medium sand or silty beds. Clast identifications (Tables 9 and 10) demonstrate the heterolithic nature of terraces Qt1 to Qt6 which include significant amounts of extrabasinal granite and gneiss. Buff-colored, occasionally-laminated, massive, pebbly silt 1-3 m thick unconformably overlies the sand and gravel (Peltier, 1949).

Terraces Qt1 and Qt2 lie approximately 32 and 22 m, respectively, above the present channel. Qt1 is equivalent to the Highspire terrace of Peltier (1949) where it was formerly

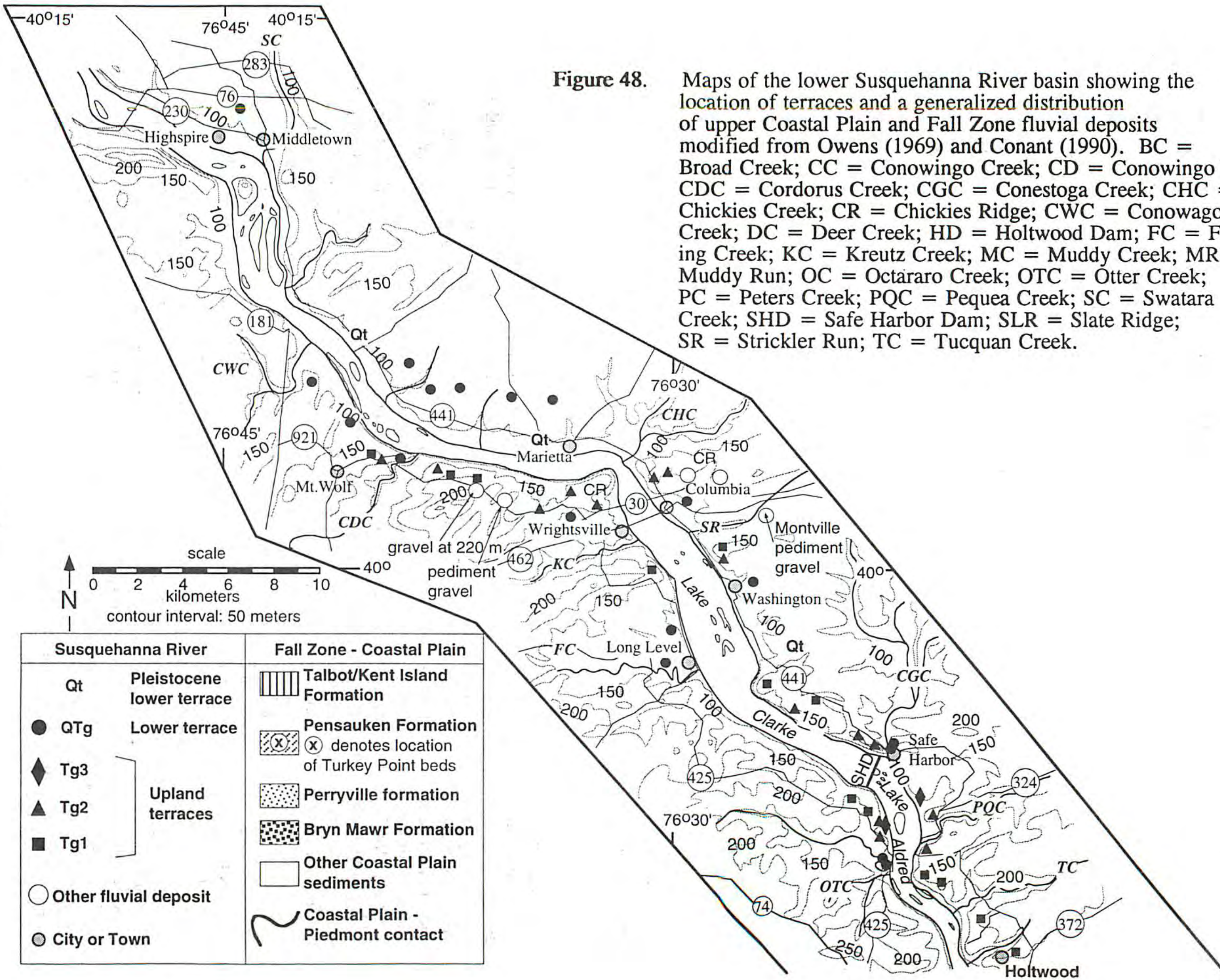


Figure 48. Maps of the lower Susquehanna River basin showing the location of terraces and a generalized distribution of upper Coastal Plain and Fall Zone fluvial deposits modified from Owens (1969) and Conant (1990). BC = Broad Creek; CC = Conowingo Creek; CD = Conowingo Dam: CDC = Cordorus Creek; CGC = Conestoga Creek; CHC = Chickies Creek; CR = Chickies Ridge; CWC = Conowago Creek; DC = Deer Creek; HD = Holtwood Dam; FC = Fishing Creek; KC = Kreutz Creek; MC = Muddy Creek; MR = Muddy Run; OC = Octararo Creek; OTC = Otter Creek; PC = Peters Creek; PQC = Pequea Creek; SC = Swatara Creek; SHD = Safe Harbor Dam; SLR = Slate Ridge; SR = Strickler Run; TC = Tucquan Creek.

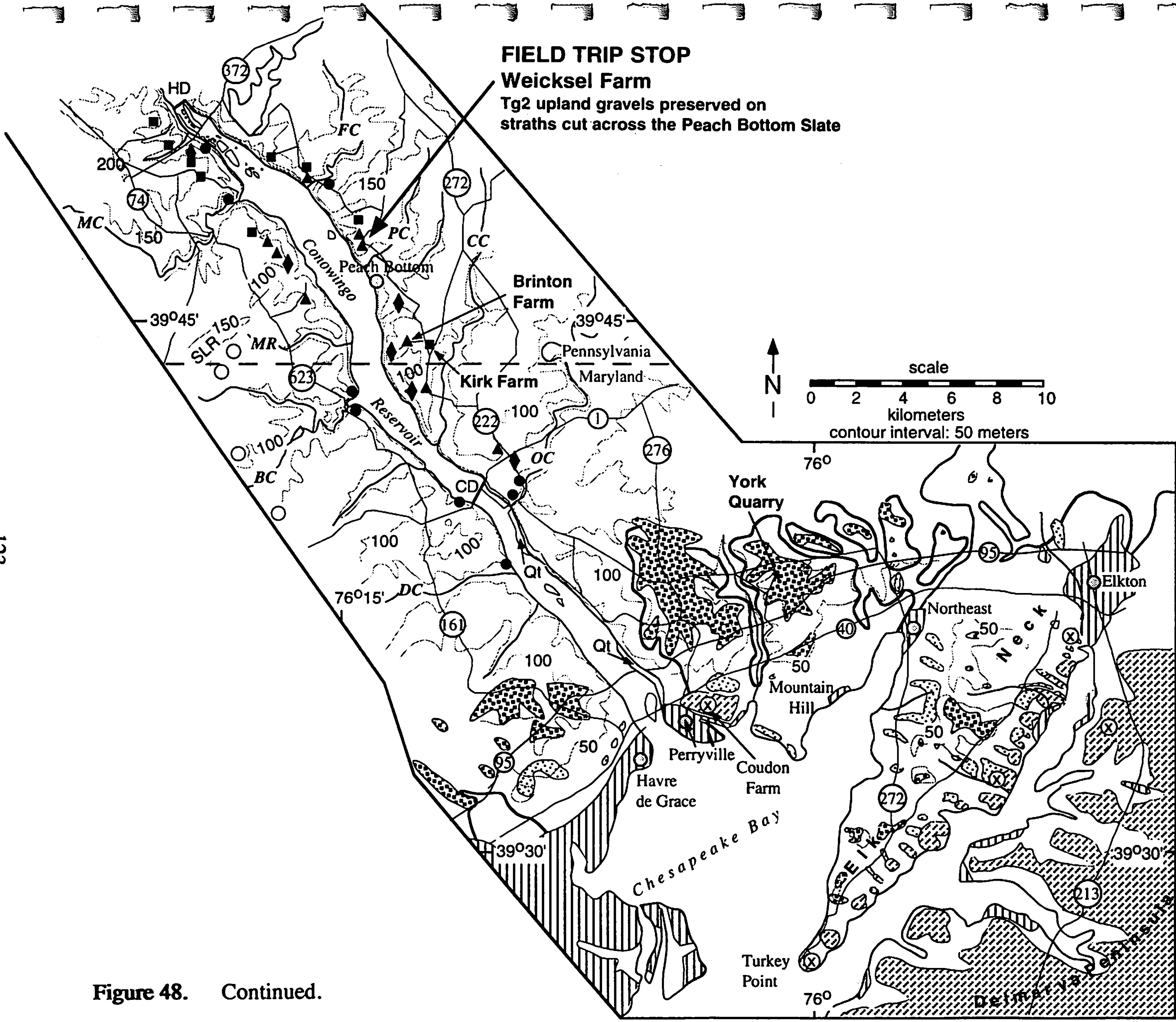


Figure 48. Continued.

Table 10. Compositional, textural, and weathering characteristics of lower Susquehanna River terraces. Data compiled from this study and from Peltier (1949).

Terrace	Composition	Shape	Size (cm)	Soil characteristics [1]	Oxidation depth	Clast rind thickness
Qt6				10 YR Bw: 0.1 m	1	0
Qt5				10 YR Bw: 0.3 m	< 5	0
Qt4				7.5 YR Bt: 0.5 m	4 - 6	0 - 0.25
Qt3	heterolithic; at least 50% quartzite, sandstone, and siltstone; up to 3% granite and gneiss	subangular to well-rounded	2 - 200	-	-	-
Qt2				7.5 YR Bt1, 5 YR and 2.5 YR 2Btb: 1 - 1.5 m*	6 m	0.5 - 1 infrequent saprolitized clasts
Qt1				-	-	0.5 - > 1 frequent saprolitzed clasts
QTg	30% vein and meta quartz 50% quartzite and sandstone 20% red siltstone and chert	subangular to well-rounded	2 - 8 with boulders 1 meter across	2.5 YR Btb: 6 m*	> 6 m	> 1 *frequent* saprolitized clasts
Tg3	35% vein and meta quartz 65% quartzite, sandstone, red siltstone, and chert	subrounded to well-rounded	2 - 20 with cobbles up to 50 cm	-	-	white, leached rinds on quartzites 0.25 to 1 cm
Tg2	50% vein and meta quartz 50% quartzite and sandstone	well-rounded	2 - 20 with cobbles up to 50 cm	-	-	white, leached rinds on quartzites 0.25 to 1 cm
Tg1	75% vein and meta quartz 25% quartzite	well-rounded	2 - 20 with cobbles up to 50 cm	10 YR Bt1, 7.5 YR 2Btb, 5 YR 3Btb*	at least 3 m	white, leached rinds on quartzites 0.25 to 1 cm

1 = soil symbols: B = illuvial zone; w = cambic horizon; t = illuvial clay; 2, 3 = change in parent material; b = buried (Soil Survey Staff, 1975).
* = soil developed in colluvial roundstone diamictons derived from originally-stratified terrace gravels

exposed as a 6 m-thick stratified deposit. At Marietta, Qt1 occurs mostly as a roundstone diamicton less than 3 m thick overlying carbonate bedrock. No soil profile descriptions exist for terrace Qt1 but labile clasts such as granite or gneiss are typically saprolitized and friable. Sandstone clasts typically exhibit weathering rinds 0.5 - greater than 1 cm (Table 10). Terrace Qt2, the best preserved and areally most extensive lower terrace, is equivalent to the early Illinoian terrace of Peltier (1949), the 46-foot terrace of Wright (1892), second terrace of Bashore (1894), and the second gravel of Stose (1928). Deeply-weathered, yellowish-red sand and gravel locally cemented by manganese-oxide (Peltier, 1949) underlies Qt2. The upper 1 to 2 m have been reworked into several distinctive roundstone diamicton

Figure 49. Photograph of Tg1 upland gravels at Kirk Farm. Pencil is 15 cm long.

units. Soil profiles developed in the diamictons are clearly polygenetic, characterized by a truncated, reddish-brown (7.5YR and 5YR), pebbly-silt, argillic horizon 1-1.5 m thick overlain by a brown argillic horizon less than 1 meter thick (Peltier, 1949; Table 10).

Terraces Qt3 and Qt4 lie approximately 10 and 15 m, respectively, above the channel and are roughly equivalent to the late Illinoian and Olean terraces of Peltier (1949). Qt4 retains original stratification and exhibits a 0.5 m-thick brown and strong brown (7.5YR and 10YR) argillic horizon developed primarily in the pebbly-silt cap (Table 10). Qt5 and Qt6 occur about 3-6 m above the current channel and have been inundated by historic floods. Terrace Qt5 is equivalent to the Binghamton and Valley Heads terraces which grade to a single terrace south of Conewago, Pennsylvania whereas Qt6 is equivalent to the Mankato terrace of Peltier (1949). Qt5 and Qt6 terrace deposits have poorly-developed brown soils exhibiting 0.1 to 0.3 meter-thick brown cambic and argillic horizons (Peltier, 1949; Table 2).

Within the Piedmont gorge, two distinct terraces underlain by stratified, heterolithic sand, gravel, and boulders, and thought to be roughly equivalent to Qt4 and Qt5, occur 6 and 12 m above the modern channel at Otter Creek, Long Level, Peach Bottom, Muddy Creek, Octararo Creek, and the reach between the river mouth and Conowingo Dam (Figure 48). At least two much more poorly-preserved, higher terraces, thought to be roughly correlative to Qt1 and Qt2, are recognized by the presence of heterolithic pebbles and large boulders along the west bank of the river from Holtwood Dam to the mouth of Muddy Creek. A narrow, gravel-mantled bedrock bench 20 m above the modern channel defines the higher terrace, while the boulder mantled, flat, concordant surface of bedrock islands defines the lower terrace. Relative weathering characteristics for lower terraces Qt1 - Qt6 in the Piedmont reach could not be obtained because of the lack of exposure and extensive anthropogenic disturbance. In most places, terraces Qt3 to Qt6 are drowned beneath reservoir pool levels.

UPPER COASTAL PLAIN AND FALL ZONE STRATIGRAPHY

Upper Coastal Plain and Fall Zone fluvial deposits at the mouth of the Susquehanna River (Figure 50) represent the proximal, updip equivalents to well-dated deposits of the marine Chesapeake Group. Petrography-based lithostratigraphic correlations along the Fall Zone and downdip into the Salisbury Embayment establish an age range and allow for a regional chronostratigraphic framework for these fluvial deposits (Pazzaglia, 1993; Figure 51). Fluvial deposits at the mouth of the Susquehanna River include the late Oligocene - late Miocene Bryn Mawr Formation quartz arenite, the Pliocene Perryville formation arkosic sublitharenite, and the Plio-Pleistocene Pensauken Formation arkose and subarkose.

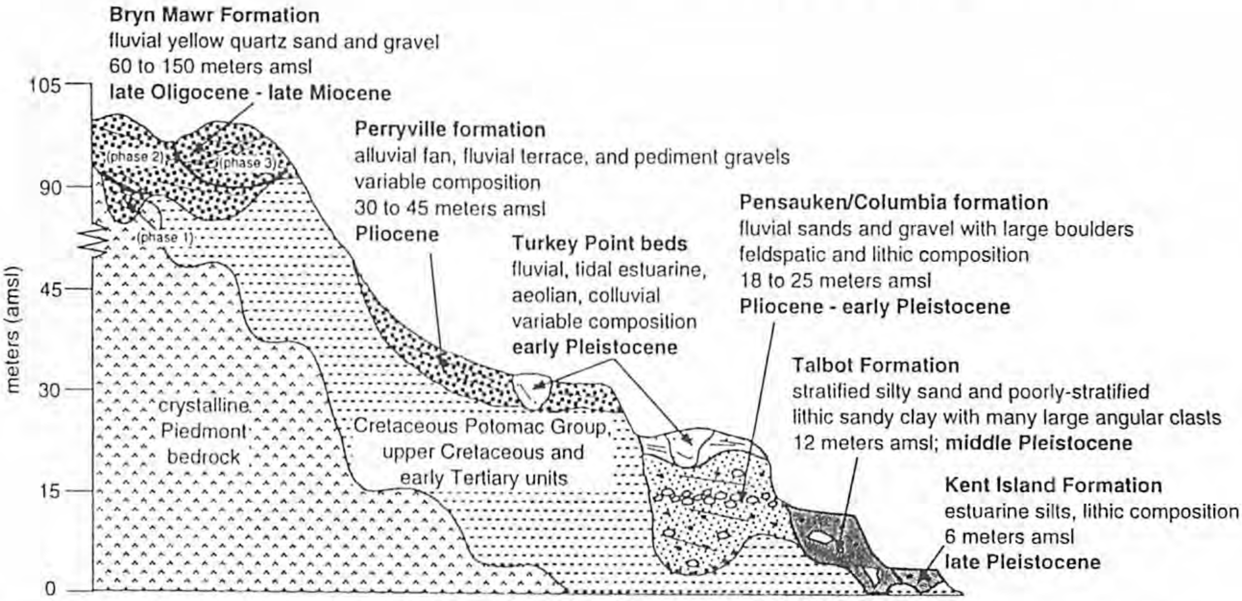


Figure 50. Schematic diagram of upper Coastal Plain and Fall Zone fluvial stratigraphy at the head of Chesapeake Bay.

The Bryn Mawr Formation (Lewis, 1880; Bascom, 1924) is a 30 m-thick, mature, well-sorted, paleo-Susquehanna River, braided-alluvial-plain, sandy gravel (Pazzaglia, 1993). Three lithofacies-defined members or phases separated by three distinct unconformities occur within this deposit. If the unconformities represent major hiatuses correlative downdip to sequence boundaries, the phases would signify fluvial aggradation in response to three third-order marine transgressive events (Figure 51). Conceptually, Bryn Mawr Formation phase 1 represents late Oligocene, upper-braid-plain deposition correlative downdip to the Oligocene, marine, Old Church Formation; Bryn Mawr Formation phase 2 represents a major period of extensive, distal-braid-plain deposition chronostratigraphically equivalent downdip to the middle to late Miocene, marine, Calvert, Choptank, and St. Marys Formations; and Bryn Mawr Formation phase 3 represents upper-braid-plain deposition inset into phases 1 and 2, correlative downdip to the late Miocene, fluvial-deltaic Bethany and Manokin formations.

Heterolithic, but quartz-dominated, fan, terrace, and pediment deposits graded to a late Tertiary, probably Pliocene, upper Chesapeake base level are inset into the Bryn Mawr Formation and informally designated the Perryville formation (Pazzaglia, 1993; Figures 50 and 51). This deposit, largely derived from the Bryn Mawr Formation and older Coastal Plain formations, also contains labile clasts including saprolitized mafic and felsic igneous rocks contributed by Piedmont streams and the Susquehanna River (Table 9).

The Pensauken Formation (Salisbury and Knapp, 1917) (Columbia Formation of the Delaware Geological Survey nomenclature; McGee, 1888; Jordan, 1964, 1974) is a bouldery, feldspathic sand representing paleo-Delaware-Hudson River fluvial channel deposition across the Delmarva Peninsula (Owens and Minard, 1979; Owens and Denny, 1979; Figures 50 and 51). This formation contains a full heavy mineral suite characterized by up to 50% labile components petrographically most similar downdip to the Pliocene Beaverdam Formation (Jordan, 1964, 1974; Pazzaglia,

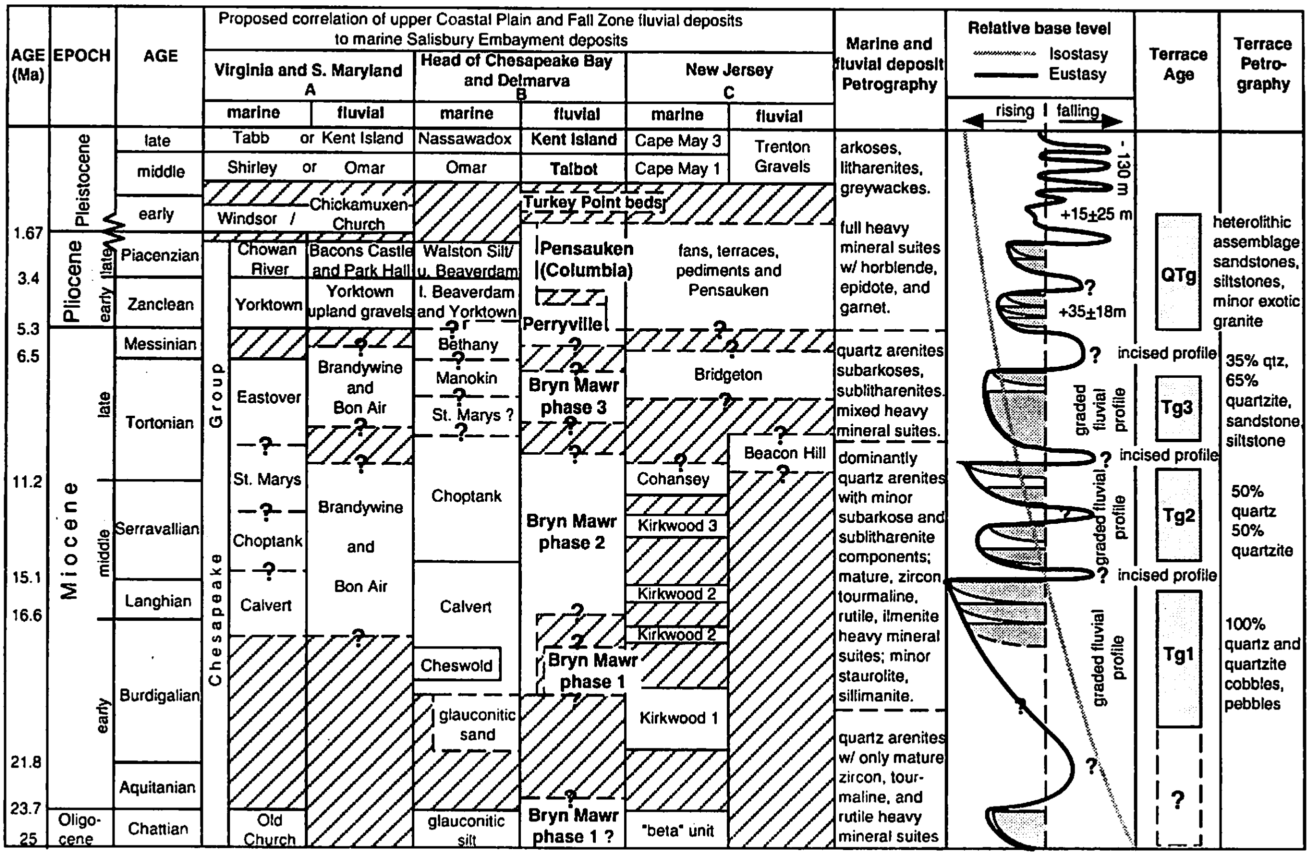


Figure 51. Regional correlation of upper Coastal Plain and Fall Zone fluvial stratigraphy between the James and Hudson Rivers and relation to lower Susqehanna River fluvial terraces. Stratigraphy compiled form various sources provided in Pazzaglia (1993). Eustasy from Haq and others, 1987; Ward and Powars, 1991; Dowsett and Cronin, 1990; Cronin and others, 1981; Greenlee and Moore, 1988).

1993). In contrast, a 2-5 m-thick bouldery lithofacies locally present in Cecil County, Maryland commonly exhibits petrographic and paleoflow characteristics more consistent with Susquehanna River fluvial deposition. Large, relatively fresh and weathered angular boulders 1 to 2 m in diameter representing virtually every Appalachian lithology, as well as those exotic to the Susquehanna River basin such as granite and gneiss, occur in this lithofacies (Table 9).

Paleo-Chesapeake Bay fluvial-estuarine and eolian deposits similar in texture, depositional environment, and relative stratigraphic position to the Windsor Formation and Walston Silt (Figures 50 and 51) called the Turkey Point beds (Pazzaglia, 1993), unconformably overlie the Perryville and Pensauken Formations. The base of the Turkey Point beds retain a reversed magnetic polarity and therefore are not younger than 720 ka. Given the late Pleistocene surficial soil developed in the Turkey Point beds (Pazzaglia, 1993), the paleomagnetic data, and the inset stratigraphic relation between the Pensauken and Perryville Formations, these units are inferred to be early Pleistocene, late Pliocene-early Pleistocene, and Pliocene age respectively.

TERRACE AGE AND CORRELATION

Interpretation of petrography (Tables 9 and 10) and downstream longitudinal projection indicate that the three upland terraces correlate to the Bryn Mawr Formation on the Fall Zone and upper Coastal Plain (Figures 50, 51, and 52a). Petrography alone cannot determine unequivocally which Bryn Mawr Formation phases correlate to the upland terraces. Most simply Tg1, Tg2, and Tg3 could correlate to Bryn Mawr Formation phase 1, phase 2, and phase 3 respectively. Based on the paucity of phase 1 deposits, their inferred advanced age, and Piedmont

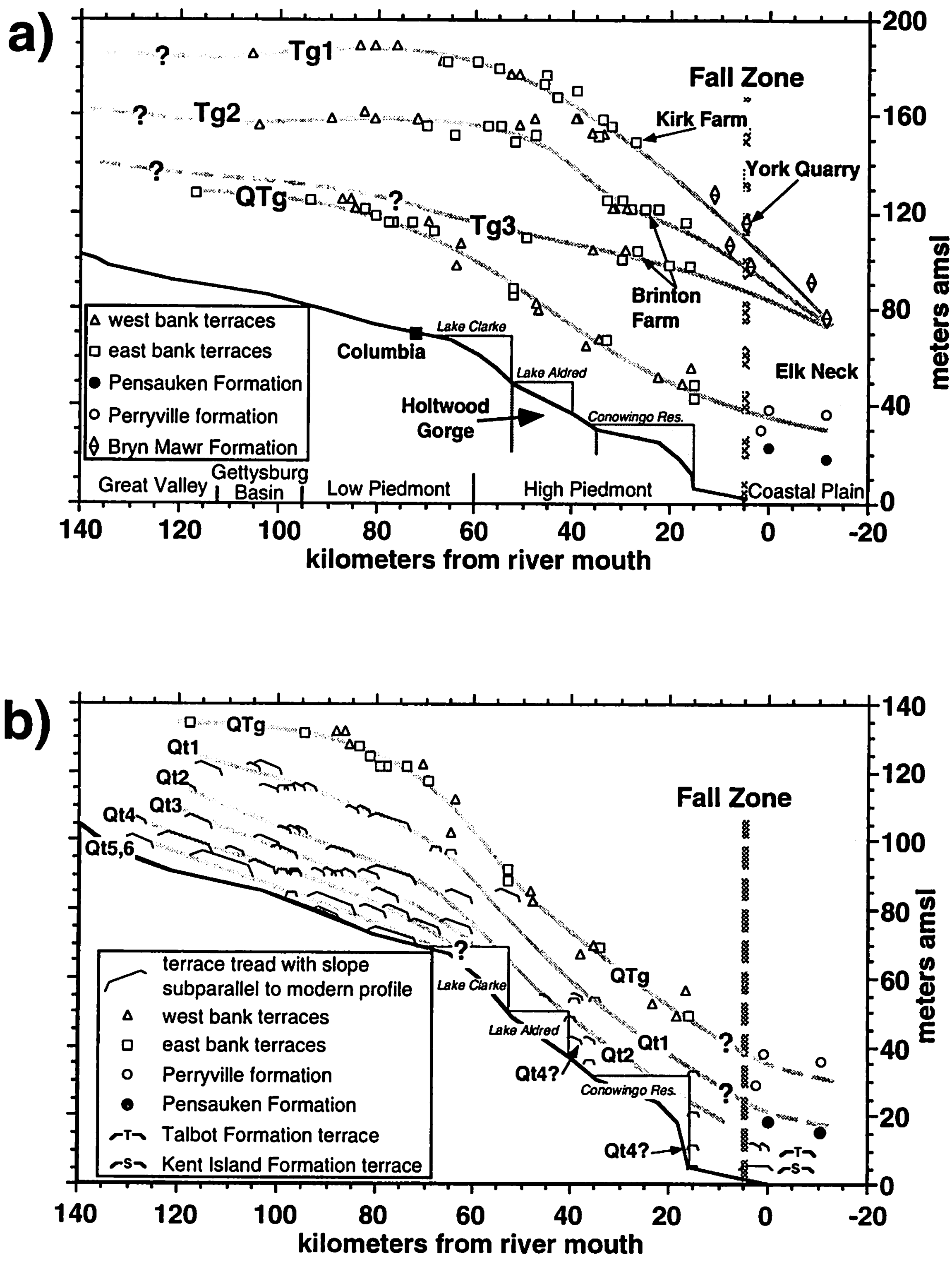


Figure 52. Correlation of **(a)** upland terraces and **(b)** lower terraces along the lower Susquehanna River from the head of Chesapeake Bay to Harrisburg, Pennsylvania. Profiles were produced by projection to a vertical plane located in the center of the Susquehanna River.

erosion rates, we propose that the oldest strath terrace (Tg1) correlates to the major period of Bryn Mawr Formation deposition represented by phase 2. Specifically, we propose that terrace Tg1 may represent strath genesis before and during Calvert Formation marine deposition (~20-15 ma), Tg2 represents strath genesis during Choptank and/or St. Marys Formation marine deposition (~14-11 ma) and Tg3 represents strath genesis during Bethany, Manokin, and Eastover Formations fluvial-deltaic and marine deposition (~9-7 ma).

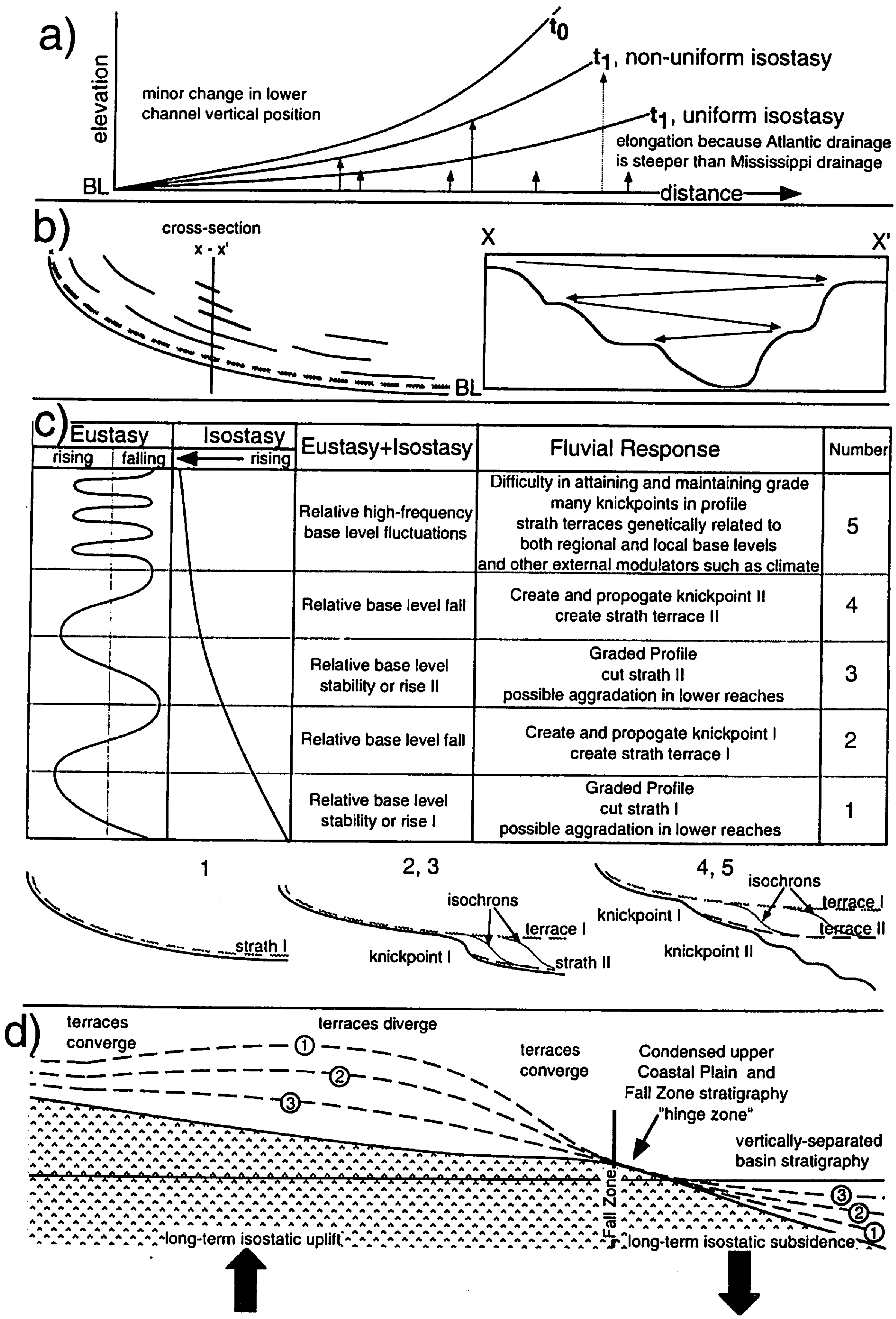
The heterolithic gravels of lower terrace QTg define a single strath consistently 35 to 45 m above and parallel to the modern channel (Figure 52b). A genetic relation of QTg to the Perryville and Pensauken Formations and a Pliocene to early Pleistocene age is interpretted from petrography and elevation of QTg at the river mouth. The presence of exotic clasts and boulders in QTg, the Pensauken Formation, and possibly in the Perryville formation (Tables 9 and 10), leads to an interpretation that terrace QTg may also genetically relate to late Pliocene - early Pleistocene glaciation of the Susquehanna basin. Pre-Illinoian glacio-lacustrine beds in the Ridge and Valley retain a paleomagnetically reversed polarity (Gardner and others, 1993; in press) indicating the presence of early Pleistocene or earlier glaciation of the Susquehanna Basin. Thus, the elevation of QTg at the river mouth, its near ubiquitous occurrence at the mouth of tributary streams, and the presence of local Piedmont and exotic lithologies are interpreted to indicate deposition along a profile similar to present, in response to increased rates of sedimention associated with early glaciation(s).

The texture, composition (Tables 9 and 10), longitudinal correlations (Figure 52b), and wellpreserved nature of terraces Qt1 - Qt6 are interpreted to indicate a Pleistocene age and genetic relation to increased sedimention rates associated with glaciation. Relative weathering and soil profile development characteristics (Marchand, 1978; Levine and Ciolkosz, 1983; Cunningham and Ciolkosz, 1984; Ridge and others, 1990) assigns terraces Qt1 to Qt6 to pre-Illinoian, Illinoian, pre-(late) Wisconsinan, late Wisconsinan (2), and Holocene age, respectively.

TERRACE GENESIS

The model for terrace genesis along the lower Susquehanna River requires a graded river with a fixed base level to attain and maintain a characteristic longitudinal profile on an isostatically dominated margin (Figure 53a). In terms of the movement of mass through the fluvial system, a graded river is defined as one that will transport all of the mass moving vertically up through the bed of the stream, as well as that supplied by hillslopes in the drainage basin, while maintaining a relatively fixed position in space (Leopold and Bull, 1979; Knox, 1975). Given the dynamic feedback between uplift of mass and fluvial erosion on an isostatically dominated passive margin with a crustal root, a river can attain and maintain a graded fluvial profile if base level remains stable at the river mouth over graded time periods (Schumm and Lichty, 1965). As the crustal root is consumed and isostatic uplift diminishes, the profile will flatten, more quickly for uniform erosion and Airy isostasy, less quickly for non-uniform erosion and spatially variable and/or flexural isostasy.

External modulating factors such as climate and base-level fluctuations cause the stream to adjust its profile resulting in the creation of strath terraces (Figure 53b). Arguably, the dominant modulator for the lower Susquehanna River, because of its proximity to the coast, is relative base level change. In this system, relative base level reflects a complex interaction between eustasy and passive margin isostasy. Erosionally driven continental isostasy results in progressive, vertical uplift over graded time periods. Over this same time span, eustatic rise and fall occurs at variable amplitudes and frequencies (Figure 53c). A prolonged period of relative base level stability is achieved by relatively low-frequency eustatic rise acting in concert with continental isostatic rise (Figure 53c). The Susquehanna River is able to attain and maintain a graded profile, cutting straths as it sweeps laterally during these periods. Periods of relative base-level fall ensue at the eustatic maxima and subsequent fall which act counter to the steady, continental isostatic rise. Strath terraces are generated downstream of an upstream-propagating fluvial knickpoint as the channel incises attempting to adjust to the new, lower base level. Thus strath genesis spans a time range commensurate with a period of Coastal Plain deposition, whereas the strath terrace is created during hiatuses in Coastal Plain deposition.



Eustasy (rising / falling)	Isostasy (← rising)	Eustasy+Isostasy	Fluvial Response	Number
		Relative high-frequency base level fluctuations	Difficulty in attaining and maintaining grade many knickpoints in profile strath terraces genetically related to both regional and local base levels and other external modulators such as climate	5
		Relative base level fall	Create and propogate knickpoint II create strath terrace II	4
		Relative base level stability or rise II	Graded Profile cut strath II possible aggradation in lower reaches	3
		Relative base level fall	Create and propogate knickpoint I create strath terrace I	2
		Relative base level stability or rise I	Graded Profile cut strath I possible aggradation in lower reaches	1

Figure 53. Terrace genesis model for streams on the U.S. Atlantic passive margin. BL = base level.

EFFECTS OF FLEXURAL ISOSTASY

Over graded and cyclic time, surficial processes erode mass from the continent and deposit it offshore. Long-term mass removal results in isostatic uplift of the continent while offshore deposition results in isostatic basin subsidence (Figure 53d). Given that the Atlantic margin is underlain by old, cold, and relatively rigid lithosphere (Beaumont, 1978; Steckler and Watts, 1981; Karner and Watts, 1982; Watts and Thorne, 1984; Pazzaglia and Gardner, 1994), we propose that a flexural isostatic response should be manifest in lower Susquehanna River terrace longitudinal profiles (Figure 52). Upland terraces Tg1 and Tg2 exhibit a broad, convex-up shape and overall gradients slightly less than the modern channel. In contrast, terrace Tg3 exhibits a relatively straight profile with a gradient of 0.0002, five times less steep than the modern channel (Figure 52). The upland terraces converge downstream into the Bryn Mawr Formation at the river mouth, diverge through the Piedmont, converge again north of the Piedmont, and finally trend parallel to the modern profile greater than 100 km from the river mouth.

Current, convex-up, upland-terrace, longitudinal profiles (Figure 52) stand in contrast to their original concave-up profiles (Figure 46, curve 2). The present upland terrace convexity reflects progressive and cumulative flexural upwarping of the passive margin landward of the Fall Zone where long-term differential crustal movement is accommodated. The condensed stratigraphic section on the Fall Zone preserved in the multiple phases of vertically-inset Bryn Mawr Formation deposits (Figures 50 and 51) attests to eustatic rather than isostatic-dominated base-level fluctuations. As flexural isostatic uplift increases landward, terraces diverge away from the condensed section at the river mouth (Figures 52a and 53d). At some point upstream, the flexural effects decay to simple Airy vertical isostasy and the terraces converge again and then finally trend parallel to the graded fluvial profile (Figures 52a and 53d).

Geodynamic model

A simple geodynamic model simulates flexural deformation based on the assumptions that (1) the U.S. Atlantic passive margin is in isostatic equilibrium (Karner and Watts, 1982), (2) original time line geometry can be reconstructed from geologic and eustatic data (Figure 6), and (3) the passive-margin lithosphere can be simulated as a uniformly thick, perfectly elastic plate. The plate is assumed to lack horizontal stresses and responds flexurally to strike-averaged, vertically applied, line loads. Initial geometry of a time line must be known because current time line elevation with respect to modern mean sea level ($E_{TL}(x)$) is the sum of original land-surface or depositional gradient ($E_{TL0}(x)$), the change in eustatic sea level (DSL), and isostatic deformation (I(x)) since the time of its creation:

$$E_{TL}(x) = E_{TL0}(x) + DSL + I(x). \qquad (1)$$

Slope and change in eustatic sea level can be obtained by regional geomorphic and stratigraphic relationships and from published sources. Values for isostatic deformation will be generated by the geodynamic model.

The model (Pazzaglia and Gardner, 1994) is composed of 17 equally spaced, 50-km-wide cells (Figure 54), aligned parallel to the lower Susquehanna River. Sediment loads in the Salisbury Embayment and Baltimore Canyon Trough (q(b)), determined from known cross-sectional areas (An) obtained from isopach maps (Poag and Sevon, 1989), and sediment densities (rs) (Scholle, 1977), are applied in cells 1 through 12 (Figure 54). Because the analytic solution does not allow lateral variations in flexural rigidity, if sediment for a given time line is not present seaward of the ECMA, loads equal in magnitude to the most seaward positioned load are applied to minimize the effects of a strong oceanic lithosphere. Erosional unloading on the continent (q(c)) determined by the product of rock density (rr), and a cross sectional area defined by Dx of 50 km, and Dy equal to the product of erosion rate (e) and time line age (tn) is applied in cells 13 through 17. Geochemical mass balance studies of saprolite production rates suggest average Piedmont denudation rates ranging from about 5 to 50 m/m.y. (Cleaves and others, 1970; 1974; Cleaves, 1989, 1993; Pavich, 1985, 1989; Pavich and others, 1989).

Model results, constrained by terrace profiles, upper Coastal Plain and Fall Zone fluvial

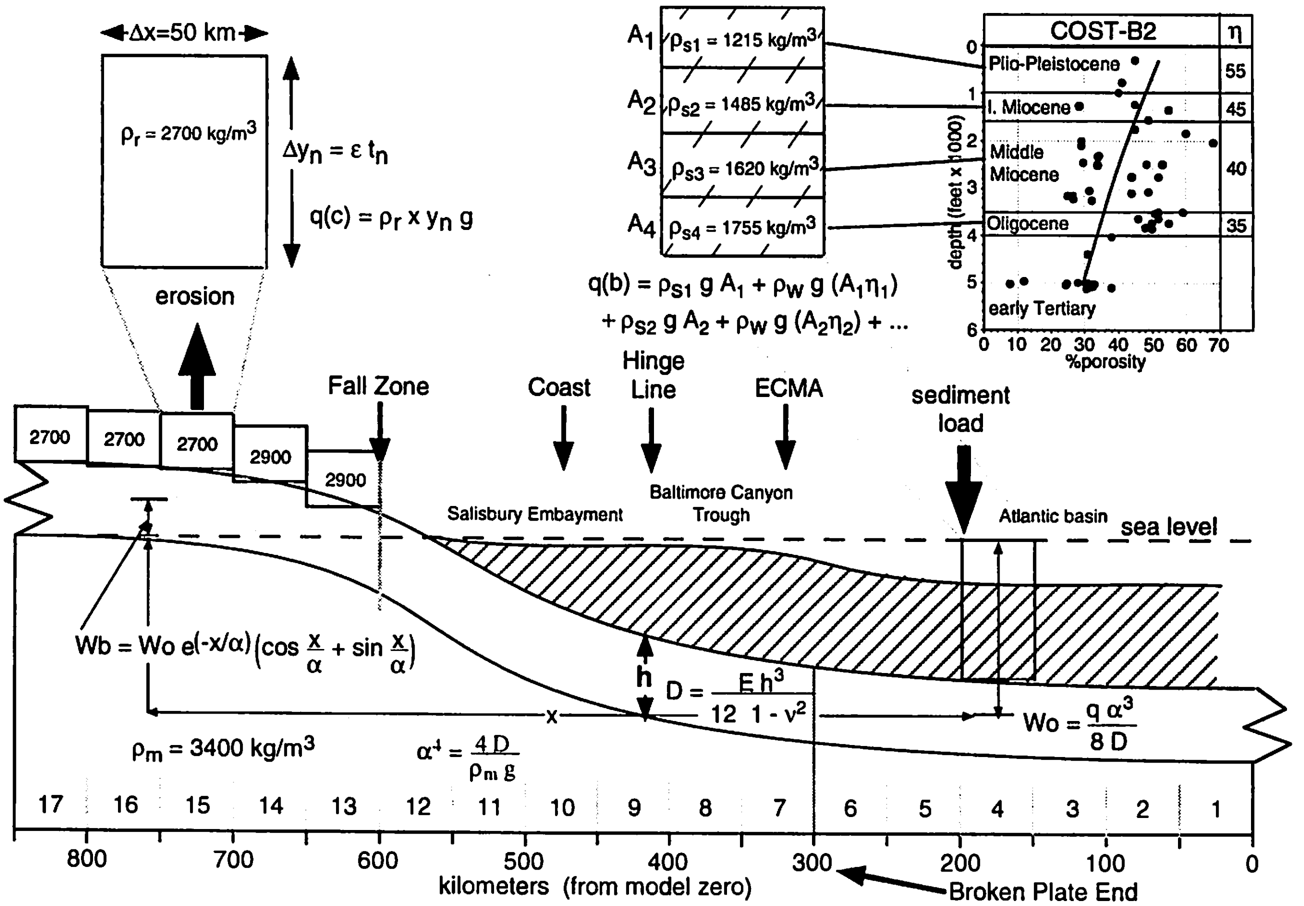


Figure 54. Geodynamic model (from Pazzaglia and Gardner, 1994).

stratigraphy, and Salisbury Embayment marine deposits show that post-Oligocene subsidence of the Baltimore Canyon Trough and Salisbury Embayment and uplift of the Appalachian Piedmont reflect flexural isostatic deformation responding to long-term continental denudation of about 10 m/m.y. and offshore sediment loading. The best correspondence between model-generated time lines and field stratigraphic data for a uniform-thickness elastic plate derives from use of an average elastic thickness of 40 km which corresponds to a flexural rigidity (D) of 4×10^{23} N m (Figure 55). These results agree with several geophysically based studies for which estimate the elastic thickness of the lithosphere underlying the U.S. passive margin ranges between 20 and 60 km (Karner and Watts, 1982; Bechtel and others, 1990).

SUMMARY AND CONCLUSIONS

Stratified and unstratified sand and gravel terrace deposits flanking the lower Susquehanna River have been remapped on the basis of petrographic characteristics and correlated with deposits of known age at the head of Chesapeake Bay. Reconstructed longitudinal terrace profiles provided data for developing a model of terrace genesis controlled by erosion, isostatic uplift, and eustatic fluctuations. This model supports flexural isostasy as the first-order late Cenozoic, passive-margin, deformation mechanism and allows for the reconstruction of the lower Susquehanna River history for the past 20 my. Total uplift of the central Appalachian Piedmont, determined by terrace age and correlation, is at least 120 m since the middle Miocene (15 ma) resulting in a long-term uplift rate of 8 m/my. Given the results presented of this paper, previous geomorphic studies describing multiple cycles of erosion and peneplain uplift (Stose, 1928, 1930) or anticlinal warping of the Piedmont (Campbell, 1929, 1933) should be modified in favor of strath terrace genesis and subsequent uplift on a passive margin dominated by eustatic, flexural isostatic, and climatic external modulating influences.

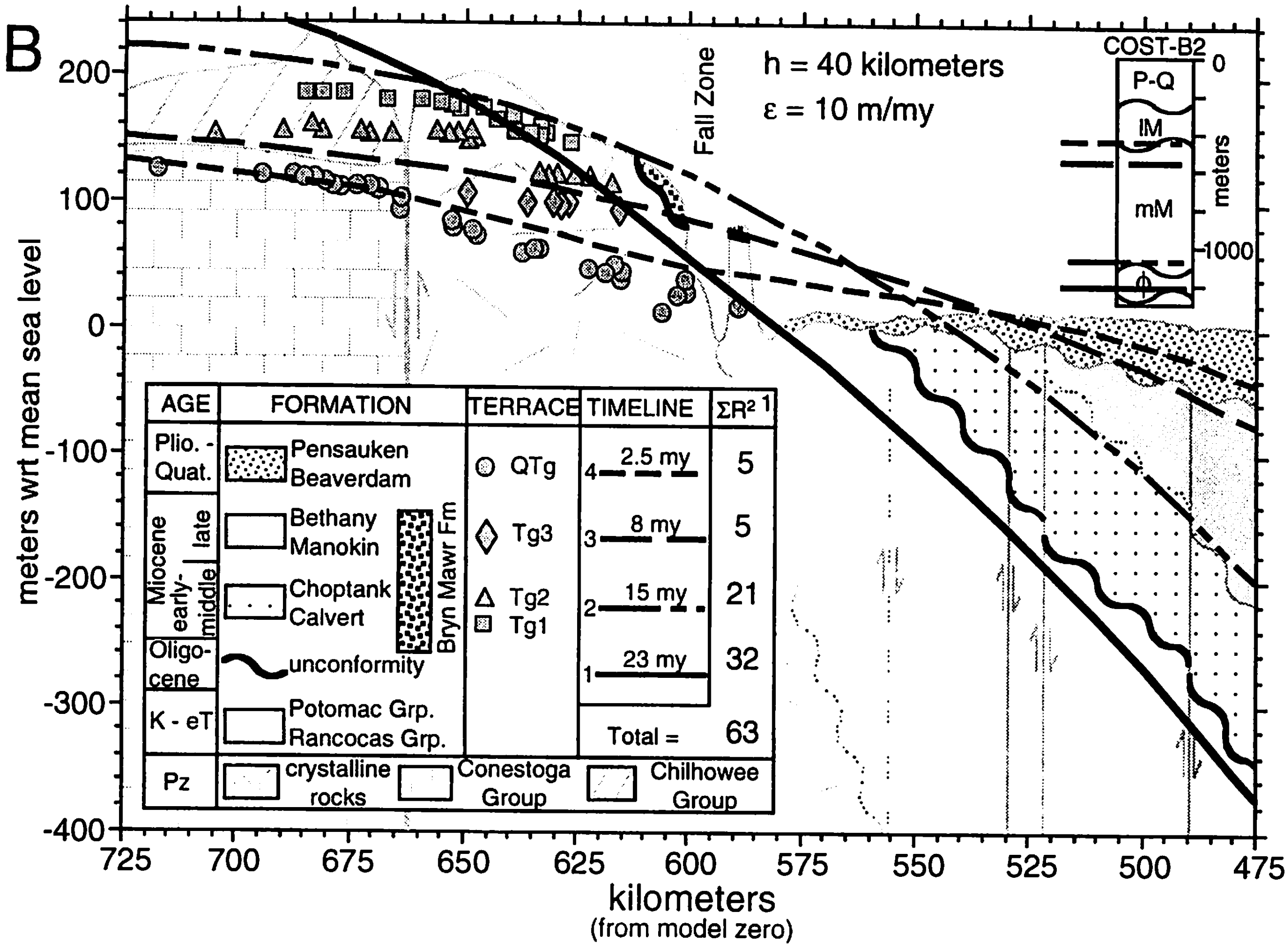


Figure 55. Model results for an erosion rate equal to 10 m/m.y. and plate elastic thickness equal to 40 km.

ACKNOWLEDGEMENTS

Acknowledgment is made to the Donors of The Petroleum Research Fund, administered by the American Chemical Society, for the support of this research. Additional support to Pazzaglia was furnished through a Shell Fellowship by The Pennsylvania State University Department of Geosciences, an Earth System Science Fellowship through EOS NASA Grant # NAS 5-30556, and a Student Research Grant of the Geological Society of America. Sincere thanks are extended to Mr. Dallas Kline and Mr. Phil Reichard of York Building Products Inc., Mr. George Coudon, Mr. Francis Kirk, Mr. Vic Brinton, Mr. Mark Abrogast, and the Holtwood Land Management Office, Sallyanne and Paul of the Otter Creek Campground, and all of the residents of Cecil and Harford Counties, Maryland and York and Lancaster Counties, Pennsylvania for their unselfish cooperation, hospitality, and support in providing access to private land. Discussions, technical support, and numerous in-field discussions with Bill Sevon, Ed Ciolkosz, Glenn Thompson, Noel Potter, Emery Cleaves, Dorothy Merritts, James Owens, Milan Pavich, Kelvin Ramsey, Rudy Slingerland, and Scott Stanford are gratefully acknowledged. This manuscript was greatly improved by a thorough review by Bill Sevon; however, we accept full responsibility for the conclusions and any remaining errors.

SURFICIAL GEOLOGY OF THE PIEDMONT IN SOUTHERN CHESTER AND LANCASTER COUNTIES, PENNSYLVANIA

W. D. Sevon
Pennsylvania Geological Survey
P.O. Box 8453, Harrisburg, PA 17105-8453

INTRODUCTION

It has long been rumored that working with bedrock in the Piedmont of Pennsylvania is very difficult because of a conspicuous lack of exposure. This rumor is true only in part. In certain parts of the Piedmont there are no natural exposures and even artificial exposures made to shallow depths are not likely to encounter bedrock. However, there are other parts of the Piedmont where outcrops are abundant. This disparity in amount of rock available for examination is directly related to the history of weathering, erosion, and deposition of surficial materials in the Piedmont.

The types and areal distribution of surficial deposits in the Piedmont of southern Chester and Lancaster Counties, Pennsylvania results from a long geological history that started during the Alleghanian orogeny, more than 260 million years ago. However, only the more recent history, the last 65 million years, is of primary importance to our understanding of the surficial materials. Following the historical discussion, the surficial deposits will be briefly described.

GEOLOGICAL HISTORY

Alleghanian orogeny to the middle Tertiary.

By the end of the Alleghanian orogeny, the Piedmont was part of a mountain range comparable in size to the modern Andes of South America (Slingerland and Furlong, 1989). This mountain range was eroded an unknown amount during 30-40 million years between the Alleghanian orogeny and the onset of Late Triassic rifting. During this time, eroded materials were transported to the west and northwest and deposited on an alluvial plain that extended to Ohio and beyond (Sevon, 1994).

Rifting between Africa and North America during the Late Triassic caused the formation of the Gettysburg-Hammer Creek-Newark basin along the north margin of the Piedmont. The Piedmont contributed most of the sediment deposited in the newly formed basin. Because the Piedmont-derived sediment was not particularly coarse grained, I conclude that the Piedmont was not a high mountain range at that time.

Sometime following the development of the Gettysburg-Hammer Creek-Newark basin, streams of significant size began to enter the basin along its northern margin. These streams, ancestors of modern drainage, included the Schuylkill River, which originated early in the history of the Hammer Creek basin and the Susquehanna River which originated late in the history of Gettysburg basin filling. It is not known when these rivers became through flowing to the newly formed Atlantic Ocean; perhaps as early as the Jurassic or as late as the Cretaceous. Once established, these rivers enlarged their drainage basins by headward erosion to their present size.

Starting in the Late Cretaceous, erosion of clastic material throughout these drainage basins decreased steadily and attained a minimum in the Late Eocene-Early Oligocene (Poag and Sevon, 1989). During this time chemical weathering and erosion was intense and a considerable thickness of saprolite developed in the Piedmont (Sevon, 1990) as well as elsewhere in Pennsylvania. Because of variations in composition between rock types (e.g., schist and gneiss) as well as within a given rock type (e.g., schist), the thickness of saprolite varied greatly in the Piedmont, particularly in areas of schist with moderate to steep dip of foliation.

Middle Miocene to Pleistocene

Climatic fluctuations that may have started as early as the Late Oligocene disrupted landscape equilibrium and erosion of clastic material commenced with vigor in the Middle Miocene. At that time the Susquehanna River, local base level for much of the Piedmont in Pennsylvania, flowed about 40 m above its present level (Pazzaglia and Gardner, this guidebook) and adjacent Piedmont landscape stood at a higher level. Braun (1989) estimates that as much as 1 km of material must have been removed from the drainage basins between the James River and Cape Cod to account for the material within the Baltimore trough. Pazzaglia and Garner (1994; this guidebook) indicate that their flexture deformation model is consistent with a long term denudation rate in the Piedmont of 10 m/m.y. Thus, perhaps only about 150$\pm$ m of material have been eroded from parts of the Piedmont during the past 15 million years while the Susquehanna River has lowered its bed only 40 m.

The visible result of this erosion was the production of the basic form of the Piedmont topography that exists today. Drainage basins were developed and enlarged. Eventually streams attained equilibrium profiles and began to create floodplains on valley bottoms as they changed from down-cutting to lateral-cutting streams. The lower parts of these drainage basins were cut into bedrock, but only weathered bedrock or saprolite was present in the headwaters of the basins.

Pleistocene

Severe climates that accompanied several glacial intervals during the Pleistocene had a pronounced effect on the Piedmont landscape. Times of three glacial intervals are known for eastern Pennsylvania: a pre-Illinoian glacial prior to 820,000 years ago (Gardner and Sadowski, 1994), an Illinoian glacial that ended about 150,000 years ago, and a late Wisconsinan glacial that ended about 20,000 years ago. A fourth much older glacial event is known in western Pennsylvania, but has not been identified in eastern Pennsylvania.

During these glacial events, the Piedmont had continuous to discontinuous permafrost. Weathering and erosion processes associated with the permafrost produced new debris through frost riving and moved unconsolidated debris from higher to lower elevations by solifluction. Some of the material stripped from uplands accumulated on sideslopes, within small tributary valleys, and on alluvial plains, while some material was removed from the Piedmont to the Baltimore trough. Pollack (1992) demonstrated that at least three episodes of deposition are recorded in the colluvial deposits that partially mantle the side slopes and tributary valleys.

The amount of upland lowering that occurred in the Piedmont during the Pleistocene was probably a few meters to a few tens of meters at the most. All told, about 10 km of material has been eroded from the Piedmont since the Alleghanian orogeny.

While the above was happening, the Susquehanna River was fluctuating between being partly filled with outwash sand and gravels and being downcut by upstream-migrating knickpoints caused by sea-level lowering. The effect on tributaries to the Susquehanna River has been minimal to date because the knickpoints have migrated only short distances up the tributaries. Below the knickpoint the valley is v-shaped and steep valley walls have abundant bedrock exposures. Upstream from the knickpoint there is a floodplain, valley walls have moderate to gentle slopes, and bedrock exposures are infrequent (lower part of the floodplain valley) to absent (upper part of the floodplain valley).

Holocene

Following the end of the Late Wisconsinan glacial, forest vegetation quickly covered the landscape and inhibited erosion. Forest burns by native North Americans may have contributed to some erosion, but land clearing and cultivation by European settlers invoked extensive erosion (Trimble, 1974). Recent changes in land use and cultivation practices have reduced the amount of present-day erosion.

SURFICIAL MATERIALS

Introduction

Much of the material at and near the surface in the Piedmont consists of loose, unconsolidated material of some type. These materials are either weathered derivatives of bedrock or weathered rock or surficial materials that have been eroded, transported, and deposited. For the purposes of mapping, I require that a mapped deposit be areally large enough to show on a 1:24,000-scale map and attain a maximum thickness (thickness is usually interpreted, not measured) of 2 meters. The boundaries of deposits shown on a map as a contact is the interpreted areal position where the deposit is no longer detectable. Thus, the contact will represent a point of minimal thickness of the mapped material.

The mapping I am doing in the Piedmont is the result of field observation, aerial photograph interpretation, and use of soils maps. No more than a few days are spent in any quadrangle obtaining ground truth. All roads are driven and good roadside exposures are examined. Contacts were interpreted on aerial photographs and drawn on topographic maps. Material interpretations are checked against soils maps, but final decisions are based on field observations. No laboratory analyses have been performed. Information from soils reports is used to supplement descriptions of materials. Figure 56 is part of a completed surficial map in the Piedmont area.

This mapping commenced in 1989 as part of cooperative project between the U. S. Geological Survey, the Maryland Geological Survey, and the Pennsylvania Geological Survey. The project was to map the surficial geology of the York 1:100,000-scale quadrangle, an area of 32 7.5-minute quadrangles (Figure 57). The project received limited Federal funding from October, 1989 to October, 1992. The Delta and Bel Air quadrangles have been mapped by the Maryland Geological Survey and are in press. Mapping for this project in Pennsylvania will be terminated upon completion of the mapping in progress (Figure 57) and the completed maps will be placed on open file late in 1994.

SURFICIAL UNITS

Rock Outcrop (●)

Rock outcrop comprises the most definitive unit on the surficial map, those areas where surficial materials are totally absent. **Rock outcrop** is shown as a blacked out area on the map. Rock will be either continuously or discontinuously exposed within the blacked out area. In discontinuous exposure, parts of the rock will be covered with either vegetation or thin colluvium or both. As a general rule, rock will be exposed extensively in the lower reaches of drainage downstream from a knickpoint at which the stream is flowing on bedrock. Upstream from the knickpoint there will be some outcrop along the valley sideslopes in the middle reaches of the drainage and generally no outcrops in the upper reaches of the drainage. Rock outcrop is mapped only where observed in the field or clearly visible on aerial photographs.

Rock (R)

Rock includes areas where outcrops are not generally present but slopes are steep to moderately steep and rock is either known or thought to be close to the surface. **Rock outcrop** is commonly associated with areas of **rock**. These areas usually are covered by woodland vegetation and have a veneer of either weathered rock material or colluvium. Bedrock will usually be present at depths less than 2 meters.

Residuum (Re)

Residuum is untransported, unconsolidated material produced by in situ weathering of bedrock. In the Piedmont of Pennsylvania, residuum is mapped only in areas of carbonate bedrock. It differs from saprolite in that none of the original rock volume, texture, or fabric is preserved. The material is mainly silt loam or silty clay loam with occasional small clasts of carbonate rock. The residuum is usually reddish in color. Thickness is variable, but depth to bedrock is generally only a few meters at most. Small outcrops of bedrock are common on sideslopes.

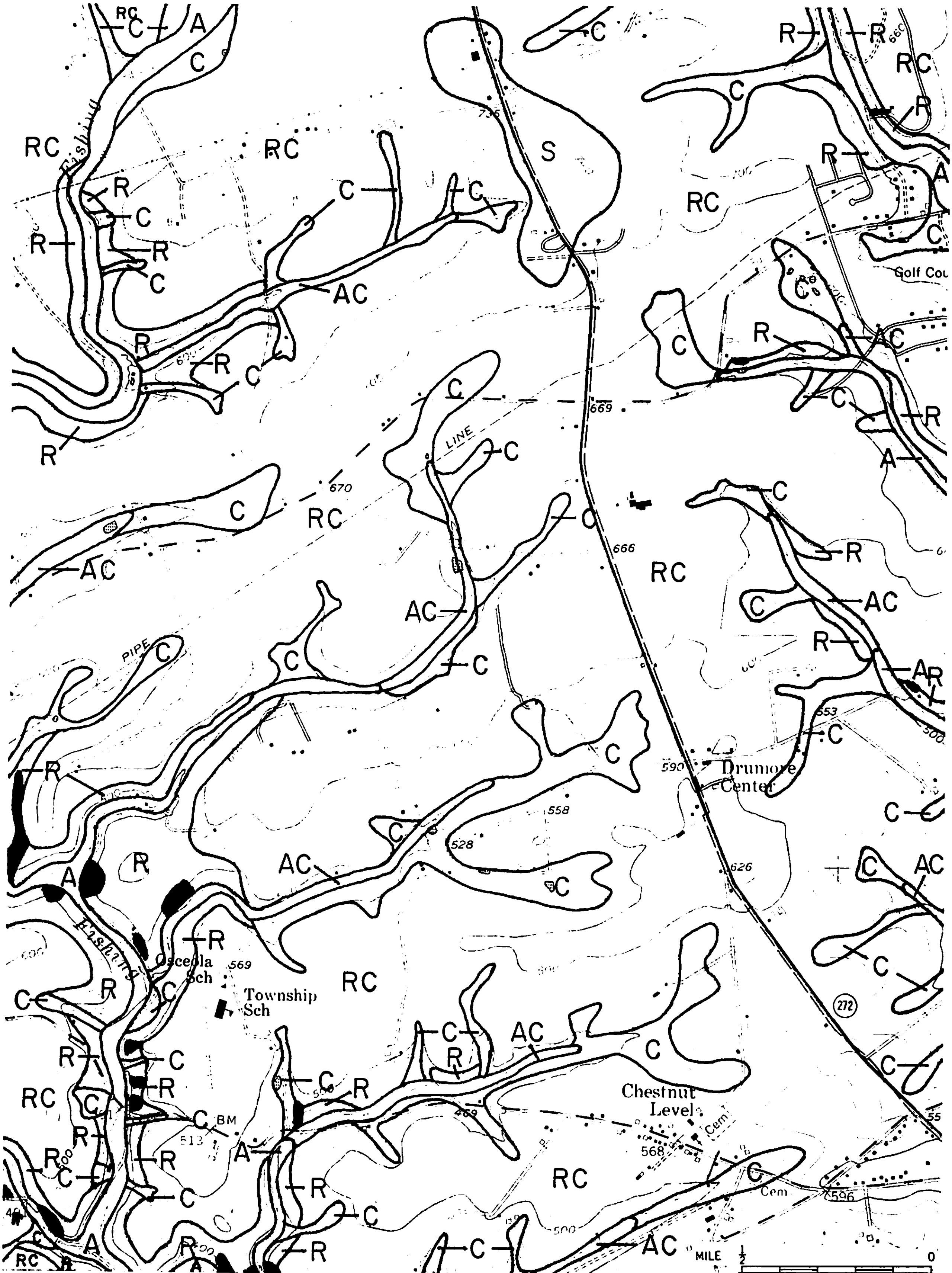


Figure 56. Part of the Wakefield 1:24,000-scale quadrangle surficial map. Map units are: Rock Outcrop ●; Rock (R); Weathered rock and colluvium (RC); Saprolite (S); Colluvium (C); Alluvium (A); and Alluvium-colluvium undivided (AC).

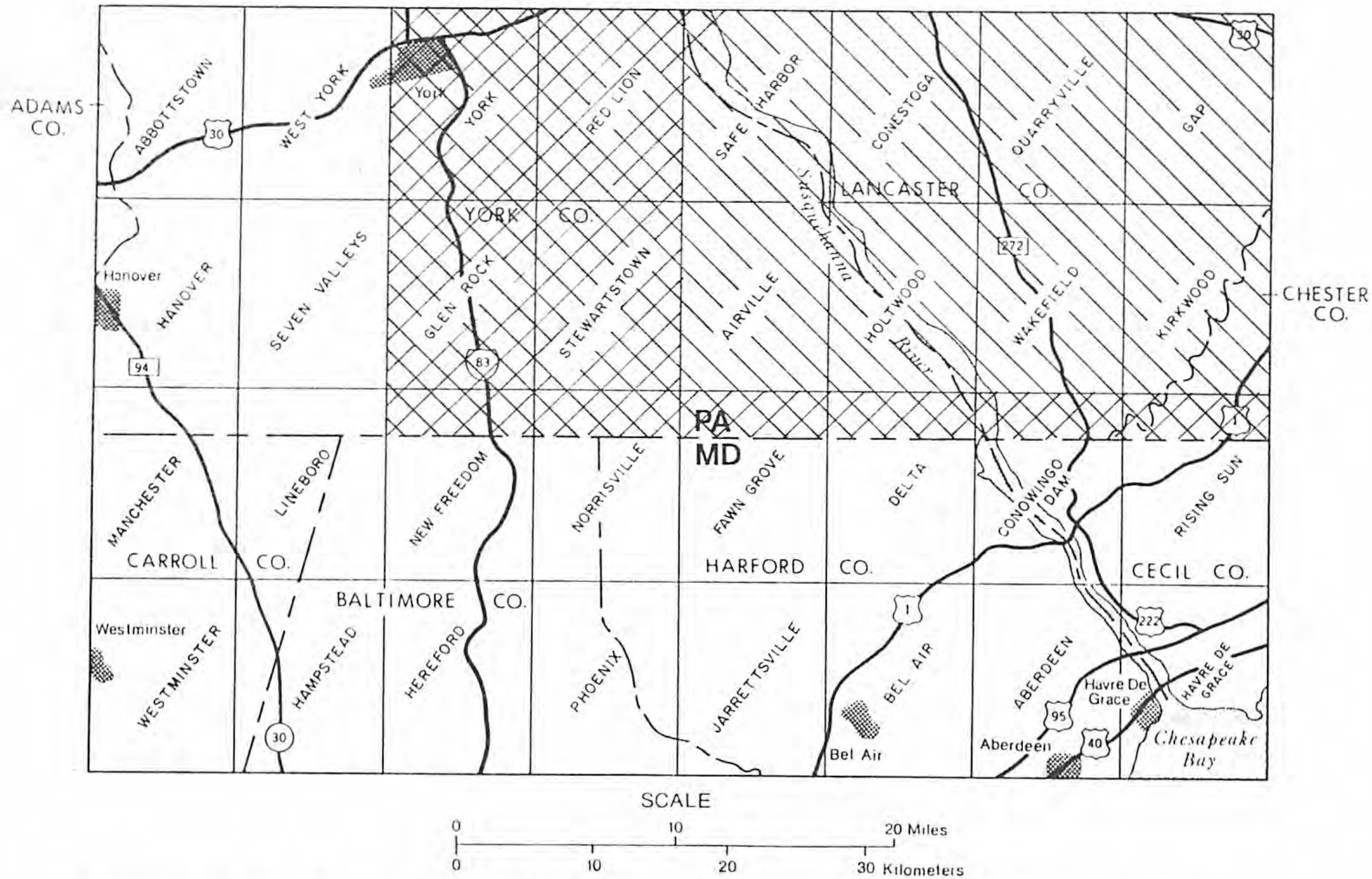


Figure 57. Map showing 1:24,000-scale quadrangles in York 1:100,000-scale quadrangle. Diagonal lines indicate quadrangles that have been completed. Cross-hatched lines indicate quadrangles in progress.

Saprolite (S)

Saprolite is unconsolidated, slightly coherent material that retains the volume, texture, and fabric of the underlying parent rock but which has compositional differences resulting from weathering. Saprolite is easily cut with a shovel or a knife. Plowed fields of saprolite have no weathered clasts, only vein quartz remains as unweathered material. Saprolite formed from schist in the Piedmont of Pennsylvania is usually redder in color than weathered schist. Depth to unweathered bedrock is several tens feet and may be more than 100 feet. Large areas of saprolite occur only on the highest uplands in areas of drainage divides. Because of the variability in degree of weathering resulting from compositional differences, small areas of saprolite often occur within larger areas of weathered rock.

Outcrops of saprolite are rare because it occurs mainly on uplands that have few or no natural or artificial cuts. Exposures in temporary backhoe pits have shown the following features of saprolite. Saprolite grades downward into weathered rock which is only partly altered compositionally and cannot be easily cut. Sometimes the saprolite is overlain by massive saprolite, a zone similar in appearance to saprolite but which lacks the fabric of the original rock. Sometimes saprolite developed from schist is overlain by "creeped saprolite," a deposit that has similar texture and fabric to the source saprolite but which has been transported by gravity-driven creep so that the fabric is more or less parallel to topographic slope rather than at the orientation of the parent fabric.

Weathered rock and colluvium (RC)

The areally most extensive mapped unit is **weathered rock** and **colluvium**. This material covers most of the uplands and low to moderate sideslopes in the Piedmont. Bedrock beneath uplands throughout the Piedmont is weathered. Depth of weathering is variable, but data from water well records indicates depths to fresh bedrock often in excess of 50 feet. Uncommon outcrops of weathered bedrock show that it consists of broken to unbroken rock that is discolored by weathering, generally to various shades light gray or yellowish gray.

Weathering causes the schist to separate along planes of foliation. Other rock types such as gneiss separate along fractures. The weathered rock is easily broken into pieces of various sizes. The pieces of broken rock may be weathered throughout, but still retain adequate coherence that a hammer is required to break a piece into smaller pieces. The ease with which weathered rock can be broken decreases with depth. The rock pieces, here called clasts for convenience and referred to as channers in soil terminology, are variable in size. Larger clasts occur where depth to unweathered bedrock is small and clasts are small when that depth is large. The size of clasts is readily observed in the spring when fields have been cultivated and rained upon once or twice.

The weathered rock may occur at the surface or may be covered with less than 2 m of colluvium. The character of the colluvium is described below. The general pattern of distribution of colluvium is as follows: thin and discontinuous patches of colluvium on uplands; generally no colluvium at the topographic shoulder, that position between the upland and the side slope where a marked change in degree of slope occurs; up to 2 m of colluvium occurring continuously to discontinuously on the sideslope; and continuous occurrence of colluvium of variable thickness at the base of the sideslope.

Colluvium (C)

Colluvium is unconsolidated material that has been transported by gravity and slope wash for some distance downslope and deposited upon the land surface. Almost all of the colluvium in the Piedmont derives from material weathered from the underlying bedrock. In some places colluvium has been transported away from the parent rock and deposited on another rock type. Colluvium is a mixture of material ranging in size from clay to small boulders. Where the clasts are platy in shape, always the case in areas of schist bedrock, there is often a crude to well-defined layering developed in the colluvium. Because colluvium is derived from local bedrock, it is difficult to distinguish from weathered bedrock on the basis of surface exposure only. Natural or artificial cuts are necessary to see the real character of colluvium. Subtle moisture differences and changes in topographic slope are useful for field and aerial photograph identification of mappable deposits of colluvium.

The colluvium can have a definite stratigraphy (Pollack, 1992). This stratigraphy represents deposition of colluvium during different time intervals. Three different colluvial layers have been identified but not all are preserved at any specific location. The oldest of the colluvial layers generally contains more fine-grained material than either of the younger layers and has been subjected to weathering that gives it a red color. The next younger layer generally has a grayish brown color with a slight hint of red. This layer appears to include a mixture of the underlying red colluvium and material derived from bedrock. The uppermost layer is grayish brown and usually contains more clasts and less fine-grained matrix than the underlying layers. Assuming that these layers represent transport and deposition during periods of glaciation, then the age of the oldest colluvium is probably pre-Illinoian, the next is Illinoian, and the youngest, Late Wisconsinan. No actual age dating of these deposits has been done.

Colluvium greater than 2 meters in thickness occurs in two places on the landscape: in the bottoms of heads of drainage and small tributary valleys that lack perennial streams; and at the base of some side slopes. The deposits are most recognizable because of changes in slope that indicate a change in character of the underlying material: colluvium has gentler slopes than adjacent rock. In addition, sometimes during the spring when fields have been cultivated colluvium will be discernible from weathered rock because of color differences and/or moisture content. Colluvium is generally sparse in areas of carbonate bedrock.

Alluvium (A)

Alluvium is material that has been transported and deposited by running water in valley bottoms. Alluvium is poorly- to well-bedded and consists of various sizes of material that are generally sorted. Materials are locally derived except for those that may be introduced from afar in larger streams that originate outside the Piedmont. Alluvium commonly has bed materials that are gravel or coarser in size and bank materials that are sand size and smaller. There is generally no soil zone developed on these alluvial soils although a grayish brown plow zone is often present just below the surface. In many places alluvium has buried soil horizons several feet below the floodplain surface. This soil is buried by material that was eroded from the landscape and deposited on the floodplain following European settlement of the area.

The upper surface of the alluvium is the floodplain. The floodplain is relatively flat and its width depends on the size of the valley. The margins of the deposit are marked by a change in slope. If the adjacent material is rock, the slope change is abrupt and usually sizable; if colluvium, subtle and low to moderate.

Alluvium and colluvium undivided (AC)

Some small valleys with perennial streams have colluvium at the base of the side slopes that grades into alluvium in the center of the valley, but the width of the valley is too small to allow both deposits to be shown at a scale of 1:24,000. In such cases the **alluvium and colluvium** are mapped as an **undivided** unit. The characteristics of the materials comprising this unit are presumably the same as described above except that the slope distinction between alluvium and colluvium is generally absent. Exposure of this undivided unit is inadequate to make definitive statements, but the unit probably includes more colluvium than alluvium. Because this unit is mapped mainly in moderately deep, steep-sided valleys where rock is close to the surface and usually relatively unweathered, alluvium and colluvium undivided consists of coarser and fresher clasts than most deposits of colluvium.

SOME COMMERCIAL ASPECTS OF THE PEACH BOTTOM SLATE: THE PROBLEM OF BEING TOO GOOD

S. W. Berkheiser, Jr.
Pennsylvania Geological Survey
P. O. Box 8453, Harrisburg, PA 17105-8453

Pennsylvania and Vermont slate production (lumped by the U. S. Bureau of Mines) accounts for about 70 percent of the volume of slate produced in the nation (Taylor, 1993). Most of Pennsylvania's slate production comes from the Lehigh-Northampton district located approximately 95 miles to the northeast of Peach Bottom. Significant slate mining has not occurred in the Peach Bottom district for some time. It is ironic that the Peach Bottom slate, which has an exceptionally high crushing strength and is unquestionably one of the best roofing slates in the world, is no longer quarried. In fact, it is this very high crushing strength (11,260 lbs/sq in) (Behre, 1933) and toughness that is in part responsible for the demise of the district.

HISTORY

According to Behre (1933), Joseph Hewes of North Carolina, a signer of the Declaration of Independence, laid one of the first slate roofs in the area at Peach Bottom. The first commercial quarrying in the district was accomplished by an Englishman by the name of William Decker, circa 1785. However, quarrying did not begin in earnest until about 1845 when Welsh immigrants discovered the district. Prior to this time, the limit of mining was the "Big Red," a clay which marked the lower limit of weathering, and slate that could be easily won using only hand tools. No explosives were used once hard rock was met, until the Welsh introduced methods brought over from the Festiniog quarries of northern Wales (Mathews, 1898). By 1850, quarrying was in full swing on both sides of the Susquehanna River (Behre, 1933). In 1858, Rogers reported 18 quarries on the west side of the river and 11 quarries on the east. Around the time of the First World War, about 20 quarries were in operation, but following the war the district experienced a steady downturn. By 1928, only one company, operating in Pennsylvania, was producing roofing granules (Behre, 1933). Since that time, minor sporadic production of roofing slates may have occurred. Filler material and roofing granules were produced in the mid 1960's and early 1970's (O'Neill, 1965; Hoover, 1971).

Marylanders apparently have been offended by the inadequate credit they receive for their production of Peach Bottom slate. Mathews (1898) lamented that the state received little credit for its share of the industry, although almost all the productive quarries were within its limits. This injustice arose from the fact that the Keystone State had better transportation routes, at that time. The shipping point for most of the quarries was Delta, Pennsylvania, which had the broad-gauge York and Peach Bottom Railroad whereas its Maryland counterpart, Cardiff, had only the narrow-gauge Baltimore and Lehigh Railroad. The injustice was compounded by the postal system (What's new here?). Delta was so much better known that residents living scarcely 100 yards from the Cardiff Post Office received their mail through the Delta Post Office.

PHYSICAL AND CHEMICAL CHARACTERISTICS

The Peach Bottom slates and associated Cardiff conglomerate form a distinctive 18-mile-long northeast-southwest-trending upland ridge in Pennsylvania and Maryland. Historically this ridge never supported significant crop development due to poor soil conditions.

The slate itself can be described as having a very dark bluish gray to dark gray color. Genth (1875) believed that the dark color of the slate was due to about 0.5 percent carbon present as graphite and partially supported this with chemical analyses (Table 11). Besides the color, perhaps the most distinguishing characteristic is what has been previously described as a very sonorous character (Dale and others, 1914). Cleavage surfaces are minutely granular and are generally wrinkled by the intersection of another cleavage. The luster is very bright and the slate is classified as an unfading mica slate (Behre, 1933; Dale and others, 1914).

Table 11. Chemical analyses of the Peach Bottom Slate.

	a		b	c	d	e	f
silicic acid	60.3	SiO_2	55.9	58.4	60.2	56.8	55.2
carbonic acid	n.d.	CO_2	n.d.	0.4	n.d.	3.4	5.3
ferric oxide	n.d.	$FeO + Fe_2O_3$	9.0	10.7	5.2	3.9	5.2
alumina	23.1	Al_2O_3	21.9	22.0	19.6	22.9	19.8
ferrous oxide	7.1		n.d.	n.d.	n.d.	n.d.	n.d.
magnesia	0.9	MgO	1.5	1.2	2.3	0.9	1.0
lime	n.d.	CaO	0.2	0.3	3.9	0.1	0.3
soda	0.5	Na_2O	4.5	n.d.	2.2	0.5	0.5
potash	3.8	K_2O	3.6	n.d.	2.9	3.6	3.4
graphite	0.6	C	1.8	0.9	n.d.	n.d.	n.d.
pyrite	0.1	FeS_2	0.1	n.d.	n.d.	n.d.	n.d.
		TiO_2	1.3	tr.	n.d.	1.2	1.1
		LOI	n.d.	n.d.	7.5	5.4	6.9
		S	n.d.	0.1	0.3	0.2	0.1
		MnO	0.6	tr.	n.d.	n.d.	n.d.

All values rounded to one decimal place. *a.* Genth, 1874. Roofing slate from Lancaster County opposite Peach Bottom, analysis by Genth, *b.* Frazer, 1880. Specimen from J. Humphrey & Co.'s Quarry, York County, analysis by McCreath, *c.* Mathews, 1898. Materials from the J. Humphrey & Co.'s Quarry, York County, analysis by Booth, Garrett, and Blair, and *d.* Behre, 1933. Slate form the J. W. Jones (Peerless) quarry, Harford County, Maryland, analysis by C. L. Lancaster, *e.* O'Neill, 1965. Grab sample from Lancaster County, analysis by Jaron, and *f.* O'Neill, 1965. Grab sample from working face, Funkhouser quarry, analysis by Jaron.

Mineralogically, the slate is composed of major mica (probably muscovite), chlorite, and quartz, minor feldspar (such as albite), kaolinite, and a trace of another feldspar such as microcline (L. Chubb personal communication, 1983). Behre (1933) also reported fine-grained andulusite, graphite, pyrite, magnetite, rutile, and zircon in thin section. More recently, chloritoid and kyanite have been verified locally.

Table 11 lists some chemical analyses from the Peach Bottom district. Generally, the Peach Bottom slates are higher in silica, alumina, and iron, but lower in carbonates, than the Lehigh-Northampton District slates. See Smith and Barnes (this guidebook) for modern chemical analysis of the Peach Bottom slates.

MINING

After the Welsh showed the English how to mine bedrock, cleavage and structure determined the mine configuration and type. The Peach Bottom district mining technique is different from that in the Lehigh-Northampton district due to the relationship between structure and cleavage. Long, narrow quarries are typical of the Peach Bottom district because a nearly vertical cleavage (apparently subparallel to bedding) is the predominate structure in what earlier workers believe is a tightly folded synclinorium (however, see Valentino: the Peach Bottom Problem, this guidebook). This nearly vertical cleavage dictates that mining of the slate is accomplished by prying loose nearly vertical slabs. Not only is this awkward and dangerous, but it produces an uneven quarry floor and is more conducive to development along strike rather than down apparent dip. By contrast, quarries in the Lehigh-Northampton District are rectangular and up to 850 feet deep. In the Peach Bottom district the abandoned "Funkhouser" quarry, on the west side of the Susquehanna River, was most recently operated by GAF Corporation for roofing granules. It was developed for about 4,500 feet along strike and shows the remains of crosscuts in the highwalls. Conventional open-pit mining was generally the norm. Tunneling was mentioned in several older quarry descriptions with reference to the tunnels providing drainage and quarry access from the valley floor level (Behre, 1933). Today, a twisted incline having a southern trend can still be observed on the east shore in the formerly productive slate ore, near the present day river level. The purpose of this incline is speculative. Approximately 1000 feet north of this incline, a drainage tunnel is still recognizable.

Although most investigators estimate the stratigraphic thickness of the Peach Bottom Slate to be approximately 1000 feet, quarry dimensions seldom exceed 200 feet in apparent stratigraphic width. The sonorous ore beds may be less than 300 feet thick.

Large waste piles remain as monuments to the determination of the Welsh immigrants who worked laboriously to carve out a new life in this county. One wonders how operations creating nearly 90 percent waste could survive in today's economy. Up to 88 percent of the slate mined in the Peach Bottom district became waste. Part of this waste problem can be attributed to the phenomena described by Mathews (1898) as "blue joints." These are sealed joint planes filled with chlorite which are difficult to identify because the chlorite mineral orientation mimics the cleavage orientation. Unfortunately these "blue joints" mostly become apparent as lines of weakness during the splitting and trimming and render the pieces useless. Dale and others (1914) estimated that losses during quarrying amounted to 25 percent and those during splitting to 50 percent. Sixty-five percent waste is generally acceptable today in the active Lehigh-Northampton District (Berkheiser, 1985).

PRODUCTS

Slate has been linked to the principle necessities of life and death, from bread boards to grave linings (Dale and others, 1914). Peach Bottom slate specimens exhibited at the Crystal Palace exposition in England in 1850 were awarded the highest premium for being the best roofing slate then known (Behre, 1933). Unfortunately, the characteristics that make this hard slate so valued as a roofing material also led to the demise of the Peach Bottom District. Its higher metamorphic grade and hardness compared to the Lehigh-Northampton slates give it exceptional roofing qualities, but limit its uses. However, somewhere in the Peach Bottom District there is probably a slate xylophone that would bring a tear to a Welshman's eye. And nearly everything else known to mankind has at one time or another been tediously fashioned from this slate, as can be witnessed at the Delta Slate Museum. But, because of the difficulty and expense of milling, Peach Bottom slate cannot economically compete with the softer more easily shaped and milled products of the Lehigh-Northampton district. In the Peach Bottom district, besides roofing slates and granules, only rough-finished grave vaults and covers, and steps and risers were ever produced in any significant quantity. Minor products included graphite filler, tombstones, and various cement and paint fillers (Behre, 1933). Knopf and Jonas (1929) noted that the slate dump of the Gorsuch Bros. Co. quarry (west side of the Susquehanna River) was used as railroad ballast when the Columbia and Port Deposit branch of the Pennsylvania Railroad had to relocate its tracks because of the Holtwood Dam construction in 1910.

Today, one of the more valuable resources of the Peach Bottom district is groundwater. Despite its hardness and appearance in outcrop the Peach Bottom Slate is, at least in some areas, capable of providing water in quantities adequate to meet even moderate industrial and public water supply needs. The median yield reported for 11 wells drilled in the Peach Bottom Slate is 18.5 gallons per minute (gpm), with a range of 8 to 140 gpm. It is reported that a well drilled 240 feet deep into the Peach Bottom Slate near the Susquehanna River was pumped at 140 gpm for 72 hours with only 12 feet of drawdown. It appears that moderate amounts of groundwater are available at relatively shallow depths in the Peach Bottom Slate. The reported well depth for twelve wells ranged from 40 to 240 feet; with a median of 115.5 feet. Groundwater from the Peach Bottom Slate tends to be low in dissolved solids, soft, and acidic. Available water quality analyses indicate that iron, manganese, and low pH may be a problem in some wells (Yannacci, 1994, written communication).

Perhaps Professor Agassiz, the great naturalist, and Mr. Humphrey, the quarry developer, best summed up the Peach Bottom slates when they toured the John Humphrey & Company quarry in the late 1800's. Professor Agassiz stated: "The Almighty might have made a more perfect fish than the trout, but he never did it." and Mr. Humphrey responded: "I say the same for the Peach Bottom slates." (Frazer, 1877).

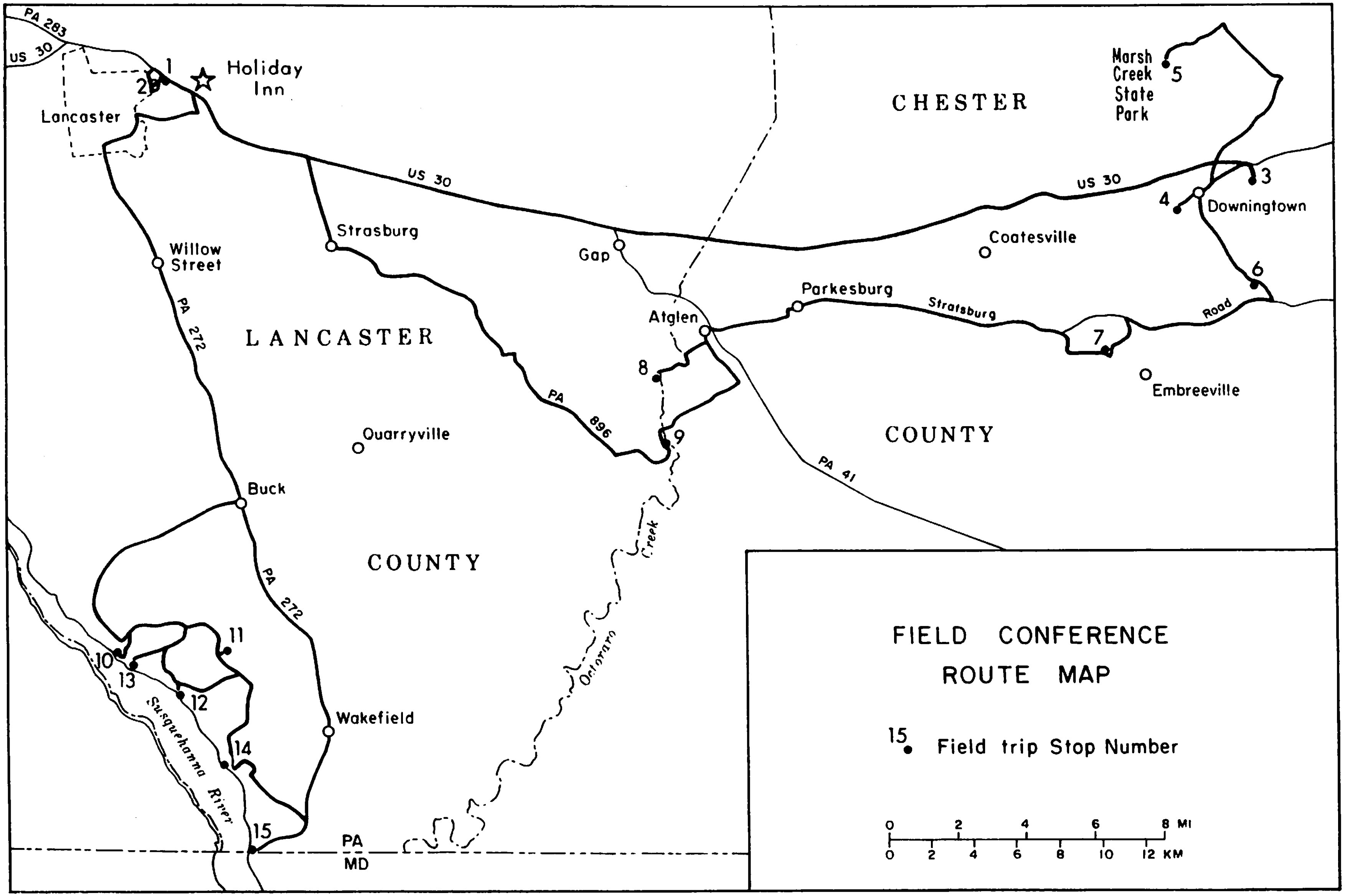


Figure 58. Field conference route map.