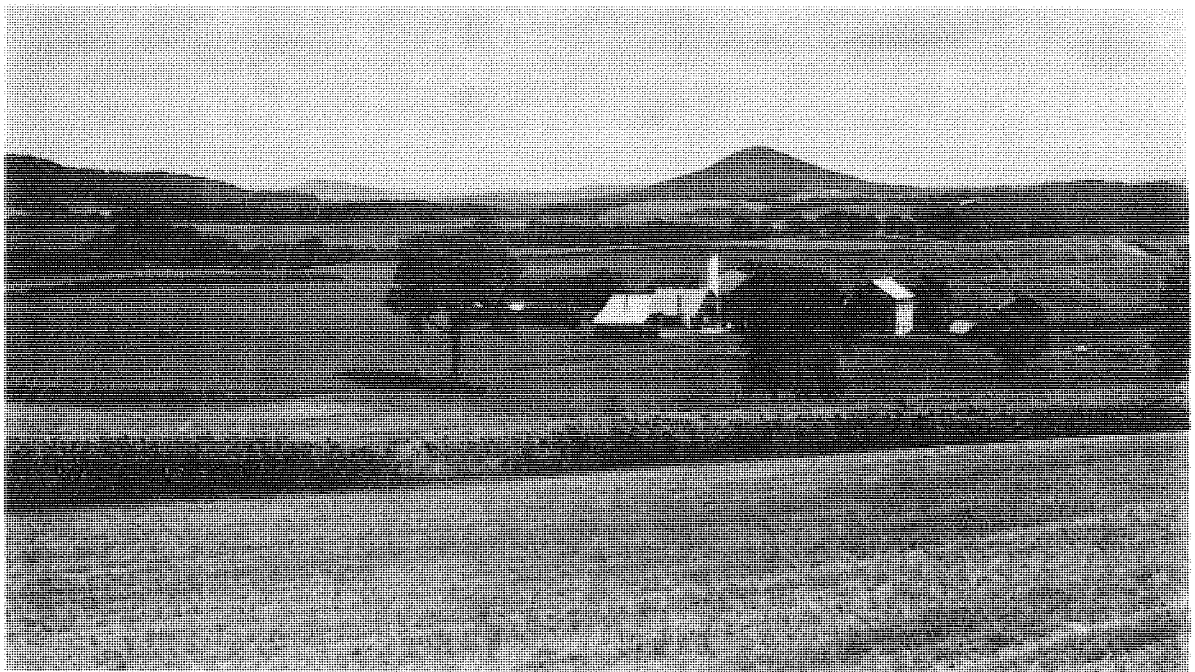
GUIDEBOOK

49th. Annual Field Conference Of Pennsylvania Geologists



Geology of an accreted terrane: the eastern Hamburg klippe and surrounding rocks, eastern Pennsylvania

October 5 and 6, 1984
Reading, Pa.

Host: Spitzenburg Hoch
Erziehunganstalt

Guidebook for the

49th ANNUAL FIELD CONFERENCE OF PENNSYLVANIA GEOLOGISTS

GEOLOGY OF AN ACCRETED TERRANE: THE EASTERN HAMBURG KLIPPE AND SURROUNDING ROCKS, EASTERN PENNSYLVANIA

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INTRODUCTION

The 49th Field Conference of Pennsylvania Geologists will have the opportunity to see Cambro-Ordovician rocks of the eastern Hamburg klippe, nearby Ordovician rocks of Shochary Ridge and the Martinsburg Formation, as well as Silurian and Devonian rocks in the Valley and Ridge north of Reading, Pennsylvania.

A new stratigraphic and structural framework has been developed by Lash in the eastern Hamburg klippe and he will lead the field trip for Stops 1 through 6. These rocks are found in two major thrust sheets. The lower thrust sheet contains a complex sequence of graywackes and shales that probably were deposited in a convergent margin setting. The upper thrust sheet contains slope carbonates and clastics that were probably deposited in a passive continental margin setting.

To the north are two parautochthonous clastic sequences, the rocks of Shochary Ridge and the Martinsburg Formation, that are in fault contact with the far-travelled klippe. Lyttle will lead the field trip for Stops 7 and 8, and Epstein for Stop 9, in these rocks. Here the emphasis will be on the differences between the two sequences and their complex tectonic history.

Overlying all of these rocks are fluvial and shelf deposits of Silurian and Devonian age of the Valley and Ridge. Epstein will conclude the trip in these rocks at Stop 10. In general, the writing of the various secitons of this guidebook was done by the person leading the field trip in the group of rocks being discussed.

The area of our field trip is unusually interesting for the following reasons:

1) The rocks in the eastern Hamburg klippe contain a wealth of sedimentological and structural information which suggests that these rocks were part of an accretionary complex at a convergent margin. This is often difficult to discern in an old orogenic belt such as in the central Appalachains.

2) The area is a transition zone in the external part of the central Appalachians where Paleozoic structures range in age from predominantly Taconic or older in the south to Alleghanian in the north. On a local scale there are many exceptions to this generalization, but taking a larger view it is certainly true. Applying the model of thrust faults younging toward the foreland, amply documented in other areas of the central Appalachians (Perry, 1979; Nickelson, 1978), and using the cross section provided on Plate 1, one can see that most of the thrusts originating in the Piedmont to the south of the area are the highest in the section, whereas the important thrusts affecting the rocks of the Valley and Ridge to the north originate much deeper structurally.

3) The structures in the area show a very large, regional warp. In the south part the geologic units dip predominantly to the south, in the central part, which includes most of the Hamburg klippe, dips are sub-horizontal, and in the northern part the dips are predominantly to the north. This large scale first-order fold may result from a deeply buried major ramp. It is based on this large fold that we have interpreted such a ramp on the cross section of Plate 1.

4) The area around Wyomissing may represent a culmination or broad warping parallel to the regional strike which has folded major thrust sheets of both Taconic and Alleghanian age, causing erosion of most of the Proterozoic gneisses that once formed a continuous stack of thrust sheets connecting the main belt of the Reading Prong and the Little South Mountain outlier. This may be crudely analogous to the classic area of the Assynt culmination in the highlands of Scotland. There, the large scale warping allows a view

beneath the oldest and uppermost thrust fault, the Moine thrust, through a series of lower and progressively younger thrusts to the Sole thrust. Whether or not these culminations are the result of buried lateral ramps or unrelated post-thrusting folds, whose axes are at a high angle to the strike of the thrust faults, these large regional warps allow us to see an excellent cross section of imbricately stacked and folded thrust sheets.

5) The eastern termination of the Hamburg klippe is remarkably coincident with the eastern termination of the rocks of the Shochary Ridge area (see Plate 1). We feel that this relation is not an accident, but instead indicates that these rocks have very closely linked sedimentologic and tectonic histories.

6) A major controversy that still exists after nearly a century of debate concerns the stratigraphic subdivision of the Martinsburg Formation, as well as the relations between the Martinsburg and the rocks of Shochary Ridge. The arguments have been based on faunal and structural evidence, much of it accumulated in the area of this field trip. In general, those workers who have studied the Martinsburg west of the Lehigh River have divided it into two parts: a lower slate unit and an upper sandstone unit (Stose, 1930; Willard, 1943; Wright and Stephens, 1978). In the Delaware Valley many geologists favor a tripartite subdivision: two slate belts separated by a middle sandstone-bearing unit (Behre, 1933; Drake and Epstein, 1967; Lash, 1978; Lyttle and Drake, 1979). On this trip we will demonstrate some of the evidence for a three-member Martinsburg and discuss the nature of the prevailing arguments.

7) The Upper Silurian through lower Middle Devonian shelf sequence is much thinner in this area than it is farther east in the Delaware Valley. Thinning of individual units and some facies changes indicates the existence of a paleopositive area to the southwest (the "Harrisburg axis" or "Auburn Promontory"), so that at Harrisburg, Pennsylvania, the entire sequence is missing.

ACKNOWLEDGEMENTS

One of us (GGL) wishes to acknowledge the Donors of the Petroleum Research Fund (PRF #14030-B2), administered by the American Chemical Company, for partial support of this work as well as for support of Fredonia State College students, Paul Bembia, Jeff Filock, and Dave Pivnik.

HISTORICAL PERSPECTIVE

Early work in the region lumped all of the pre-Silurian clastic rocks shown on Plate 1 in one unit, only distinguishing between slates and graywackes. Rogers (1858) called these rocks the Matinal Series and was the first to divide them into two parts, an upper shale and a lower black slate. Continuing with the bipartite division, Lesley (1892) correlated the upper thick-bedded and the lower thin-bedded members with the Utica and Hudson River Slates. Peck (1908) recognized three members, a lower slate, a middle sandstone, and an upper slate. In 1910, Stose proposed the correlation of these rocks in Pennsylvania with the Martinsburg Formation of West Virginia. Like many of the southern workers he preferred the bipartite division of the Martinsburg, a lower slate and an upper sandstone. Behre (1933) reinstituted Peck's three-fold division and recognized important differences between the upper and lower slates. Stose (1930) responded by suggesting that Behre's upper slate was a structural repetition of the lower slate across a regional syncline. Following Stose's lead, Willard and Cleaves (1939) named the upper sandstone member the Shochary Sandstone.

In the 1960's detailed mapping of 7 1/2 minute quadrangles by geologists of the U. S. Geological Survey, including A. A. Drake, Jr., J. M. Aaron, D. C. Alvord, R. E. Davis, and Jack B. Epstein, supported Behre's three-member Martinsburg in eastern Pennsylvania and northern New Jersey. Drake and Epstein (1967) proposed the names Bushkill, Ramseyburg, and Pen Argyl for the lower, middle, and upper members respectively. This tripartite classification of the Martinsburg, also recognized by us and used on Plate 1, has recently been challenged by Wright and Stevens (1978) and Wright and others (1979). Using biostratigraphic arguments, they prefer the bipartite classification and structural explanations of Stose.

However, detailed mapping in eastern Pennsylvania clearly shows that the Bushkill, Ramseyburg, and Pen Argyl Members are a continuous sequence—the Pen Argyl is seen to stratigraphically overlie the Ramseyburg wherever there are adequate exposures at the contact. Where the Ramseyburg structurally overlies the Pen Argyl, it can be shown that the sequence is overturned. The thick-bedded slates of the Pen Argyl are definitely not repeated south of the Ramseyburg outcrop belt, a fact long known to the slate quarrymen of the area. They recognize the difference between the Pen Argyl and Bushkill belts, and named them the "soft slate" and "hard slate" belts, respectively. For a number of reasons, we are unconvinced that the graptolite distribution patterns of Wright and others (1979) prove that the upper and lower members are the same age. An alternate explanation is that graptolites suffer from facies control as do all other paleontologic groups, and we are dealing with recurrent faunas or overlapping ranges in the two slate members (see "Graptolite zones in the Ordovician clastic rocks" under the section on the Martinsburg Formation). The most recent geologic map of Pennsylvania (Berg and others, 1980) neatly avoids the issue by showing the three belts on the map, with northern and southern belts apparently repeated by folding, but also showing slate units above <u>and</u> below the graywacke-rich rocks (the Ramseyburg Member) in the explanation!

This report divides the rocks of Shochary Ridge into two units, the Shochary Sandstone and the New Tripoli Formation (Lyttle, Lash and Epstein, 1985; Lyttle and Epstein, 1985). No previous workers have made a distinction between this group of rocks and the Martinsburg Formation. Although the ages of both groups of rocks may, for the most part, be equivalent, there are definitely easily recognizable differences in lithology and bedding characteristics that can be mapped in the field. In addition, the rocks of Shochary Ridge are entirely separated from all surrounding units, including the Martinsburg, by thrust faults (see Plate 1).

Rogers (1858) was the first to recognize the peculiar nature of the rocks that constitute the Hamburg klippe. Subsequent studies by Kay (1941) and Stose (1946,1950a, b) pointed out that these rocks, anomalous in age and lithology with respect to surrounding rocks of the Pennsylvania Great Valley, are similar to the Taconic sequence of New York and New England. Stose (1946) referred to the gray, black, red, and green shale, graywacke, limestone, chert, and volcanic rocks that extend between the Lehigh and Susquehanna Rivers as the Hamburg klippe (fig. 1), and maintained that these rocks were thrust from an eastern ocean

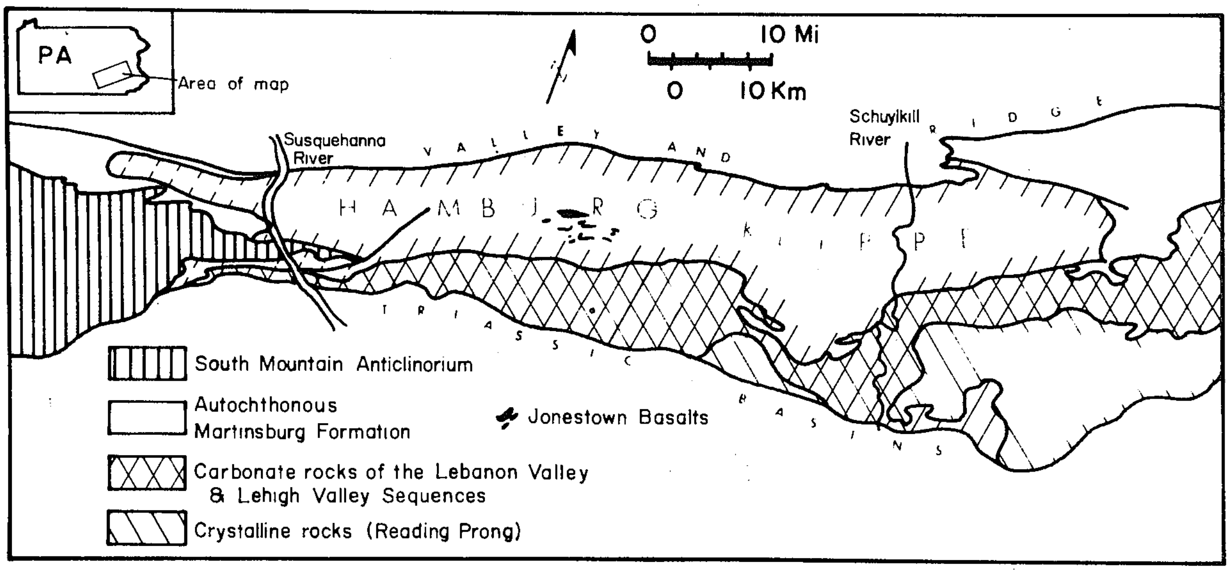


Figure 1. Generalized geologic map of the Hamburg klippe and surrounding areas. The portion of the field conference devoted to the klippe is at the eastern end of the allochthon (modified from Root and MacLachlan, 1978).

westward onto the Martinsburg Formation and underlying older carbonate rocks.

Despite the arguments of Stose and Kay, Gray and Willard (1955) still believed that these rocks were facies equivalents of the Martinsburg Formation and were not allochthonous. Behre (1933), concerned primarily with the commercial slate belt to the northeast, worked along Maiden Creek north of Lenhartsville, but apparently did not recognize the allochthonous nature of the rocks in that area. Likewise, McBride (1962) in his classic study of the Martinsburg throughout the central Appalachians did not separate the klippe rocks from the Martinsburg.

In the late 1960's and early 1970's, a number of geologists reviewed the problem of the Hamburg klippe. In particular, studies by Myers (1969) and Alterman (1972) showed that there are, indeed, allochthonous elements within the Pennsylvania Great Valley. Studies of the conodont fauna of allochthonous carbonate rocks by Bergström and others (1972) showed that not only are the allochthonous rocks older than the surrounding Martinsburg, but that they also contain a Balto-Scandic fauna rather than the North American Midcontinent fauna typical of the partly coeval and parautochthonous Beekmantown Group (see section "Conodonts from the Greenwich slice of the Hamburg klippe near Greenawald, Pennsylvania"). In addition, graptolite-based biostratigraphic studies by Wright and Stephens (1978) and Wright and others (1979) illustrated that the allochthonous graywacke and shale is significantly older than the lithologically similar Martinsburg.

The internal stratigraphy and structure of the klippe rocks is not fully understood and is still controversial. Indeed, mapping in the klippe (Wood and MacLachlan, 1978; Alterman, 1972; Carswell and others, 1968; Dyson, 1967; Lash, 1980) has emphasized the problems of tracing individual lithologic units along strike. Platt and others (1972) suggested that instead of the allochthon being a single plate it may be a jumble of smaller parts that were emplaced into the Martinsburg basin by gravity sliding. Recent investigations (e.g., Lash, 1980; Stephens and others, 1982), however, have illustrated that certain lithologic associations can be defined within the allochthon suggesting that an internal stratigraphy does exist. For example, red and aqua-green shale and mudstone together with deep-water limestone and chert is an association that can be mapped. An internal stratigraphy is consistent with the allochthon being composed of stacked thrust slices (e.g., Alterman, 1972; Root and MacLachlan, 1978; MacLachlan, 1979; Lash, 1980) rather than a chaotic melange of different parts. In fact, recent mapping by Lash (1980, work in progress) indicates that some of the slices (e.g., Greenwich slice) can be traced over the entire 125 km length of the klippe.

The northeast contact of the Hamburg klippe as shown on Plate 1 is essentially the same as proposed by Stose (1946) and confirmed much later by Alterman (1972). Though most recent workers agree on the boundaries of the klippe, there has been considerable debate about the tectonic history of these rocks, and the nature of the contact of the klippe with surrounding rocks. Like Stose and Alterman, we map the northern contact of the Hamburg klippe as a fault and strongly disagree with Wright and others (1979) who claim that the rocks of Shochary Ridge rest conformably on the klippe rocks. However, we recognize faults of at least two different ages, and the Kistler Valley fault, which marks the boundary between the klippe and the rocks of Shochary Ridge, is probably Alleghanian and not Taconic in age.

HAMBURG KLIPPE SEQUENCE

Mapping at the eastern end of the Hamburg klippe (Lash, 1980) has led to the recognition of two lithotectonic units, or sequences of rocks bounded by thrust faults. The lithotectonic units had similar depositional and tectonic histories. The tectonically lowest slice at the eastern end of the allochthon is the Greenwich slice (Lash, 1980; Lash and Drake, 1984) which consists of two major lithologic assemblages: (1) graywacke and olive-green and greenish-gray shale and (2) red, purple, and aqua-green mudstone and shale, deep-water limestone, chert, and siliceous shale. Small but conspicuous areas of boulder and pebble conglomerate and intrusive and extrusive volcanic rocks are also present within the Greenwich slice. These rocks were probably deposited in a convergent margin setting. The tectonically overlying Richmond slice consists of sandstones, siltstones, and carbonates depositied on the slope of a passive continental margin.

Greenwich Slice

The Greenwich slice includes rocks of the Windsor Township Formation (Lash, 1980; Lash and Drake, 1984), a flysch sequence that consists of medium- to coarse-grained lithic-arenites and lithic-wackes interbedded with olive-green and greenish-gray shale, and siltstone. The flysch is associated with red, purple, and aqua-green shale, micritic limestone, and chert which will be referred to as "red shale units." Minor but conspicuous boulder and pebble conglomerates appear to be more closely associated with the flysch rather than the red shale units.

Windsor Township Formation

The Windsor Township Formation is a sequence of interbedded sandstone and olive-green and greenish-gray mudstone, shale, and siltstone best exposed north of Hamburg and along the west bank of Maiden Creek north of the town of Dreibelbis in the Hamburg 7 1/2-minute quadrangle. These rocks constitute about 85 percent of the Greenwich slice.

The interbedded sandstone and shale of the Windsor Township Formation will be described using the turbidite nomenclature of Bouma (1962; fig. 2). Analysis of detailed measured sections indicates that the Windsor Township is characterized by four major facies: (1) thick sandstone; (2) turbidite-shale; (3) siltstone-shale; and (4) shale.

Thick sandstone facies.

The thick sandstone facies is characterized by thick- to very thick-bedded, locally conglomeratic sandstone and granule beds that are generally massive, poorly to non-graded, and continuous within outcrops as much as 5 m (16 ft) long (fig. 3). Most of the graded beds have coarse-tail grading (fig. 4), although rare reverse grading can be seen locally. Beds of the thick sandstone facies are typically 1 to 6 m (3-20 ft) thick. Amalgamation is common (fig. 5) which results in multistory sandstone sequences up to 9 m (30 ft) thick.

Erosional features are common in this facies and include (1) irregular and discontinuous layers of shale separating sand beds, (2) flat shale clasts (rip-up clasts) up to 1.5 m (5 ft) long within the lower parts of beds (fig. 6), and (3) small erosive channels cut into underlying hemipelagic mud (fig. 7). In addition, amalgamated sandstone beds often have well defined erosive bottom contacts (fig. 8).

Although most beds are structureless, parallel and long wave-length ripple laminations (anti-dune structures?) are present locally (fig. 9). Fluid escape structures such as pillars are rare although dish structures can be seen in a number of beds (fig. 10). Tool marks are the predominant current indicator and many beds display load casts and flame structures.

The thick sandstone facies of the Windsor Township Formation is similar to Facies B1 and B2 of the Mutti and Ricci-Lucchi (1975) classification. Beds of this facies are believed to result from a combination of fluidized and high density turbulent flow. This is supported by the presence of fluid escape structures and coarse-tail grading. Hiscott and Middleton (1979) suggested that horizontal parallel laminations, similar to those present in this facies, may result from sudden deposition or "freezing" of traction carpets at the base of high density turbidity currents. A particularly interesting feature of some of these is the presence of thin (less than 1 cm) lag deposits near the tops of the beds (fig. 4). These apparently result from reworking of sediment deposited from the head of a highly concentrated turbidity current by elutriated or entrained sediment of the tail of the current. That is, low concentration sediment-water mixtures flow over and rework the sediment leaving a well-sorted lag deposit at the top of the main deposit.

Turbidite-shale facies.

The turbidite-shale facies of the Windsor Township Formation consists of a classic coarse- to fine-grained proximal sequence of turbidite beds (fig. 11) that are generally less

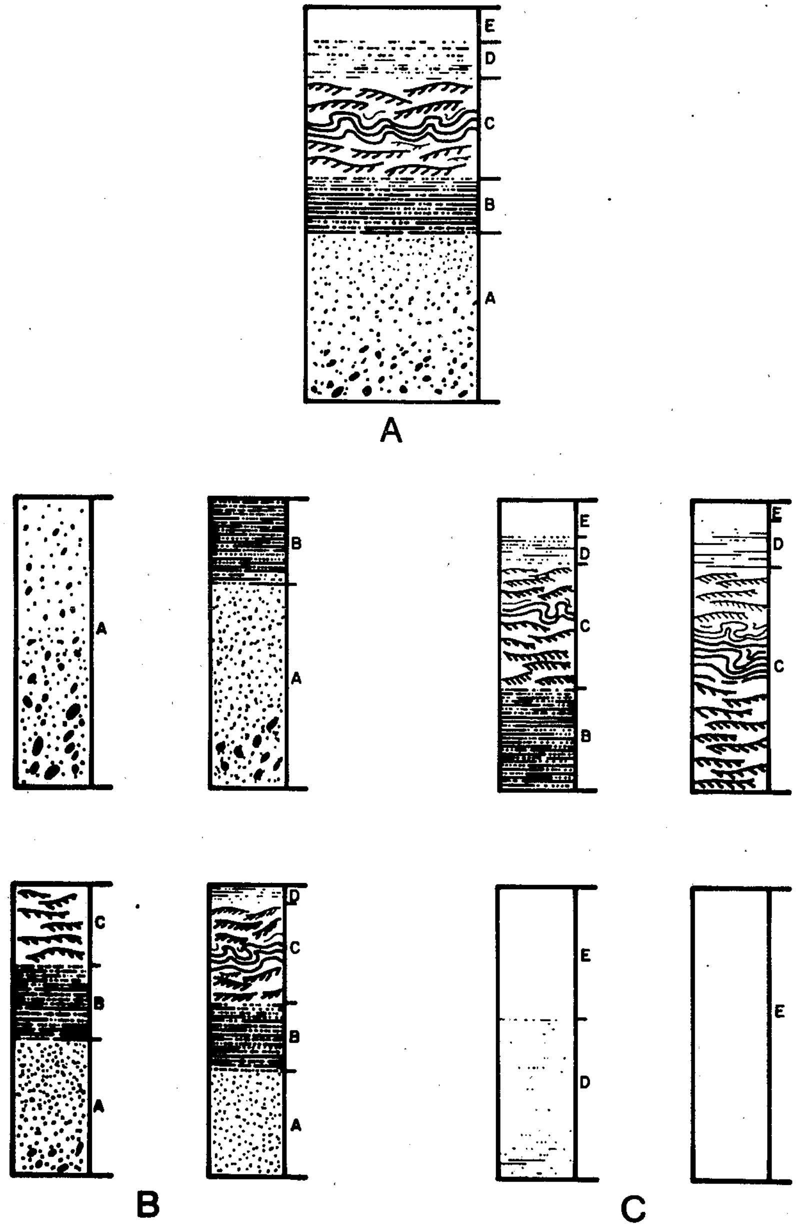


Figure 2. Bouma sequence and variations of it (from Bouma, 1962)
A. complete sequence
B. top-truncated sequence
C. base cut-out sequence
Sedimentary intervals:
a-graded interval
b-lower interval of parallel lamination
c-interval of current ripple lamination
d-upper interval of parallel lamination
e-pelitic interval

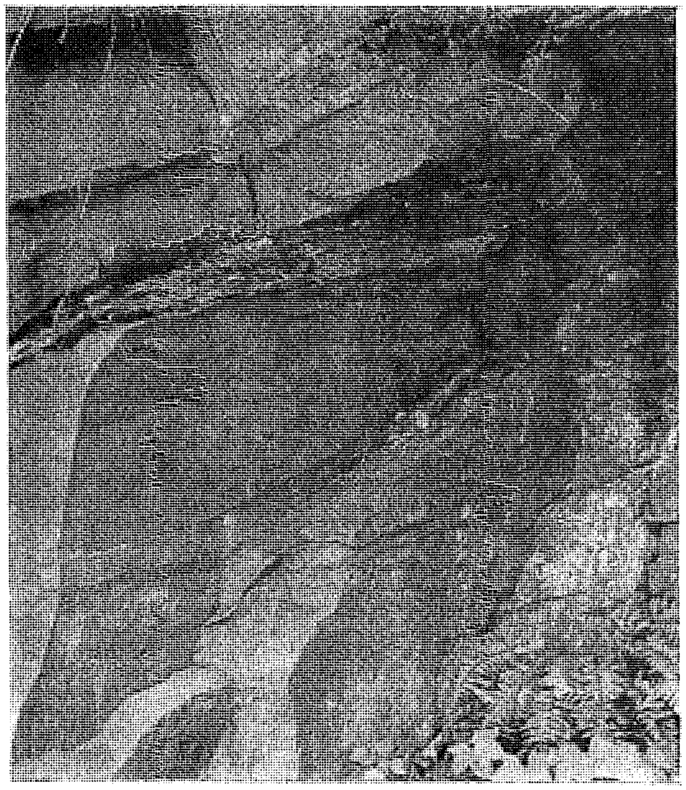


Figure 3. Thick-sandstone facies beds. Note changes in thickness of the interbedded shale.

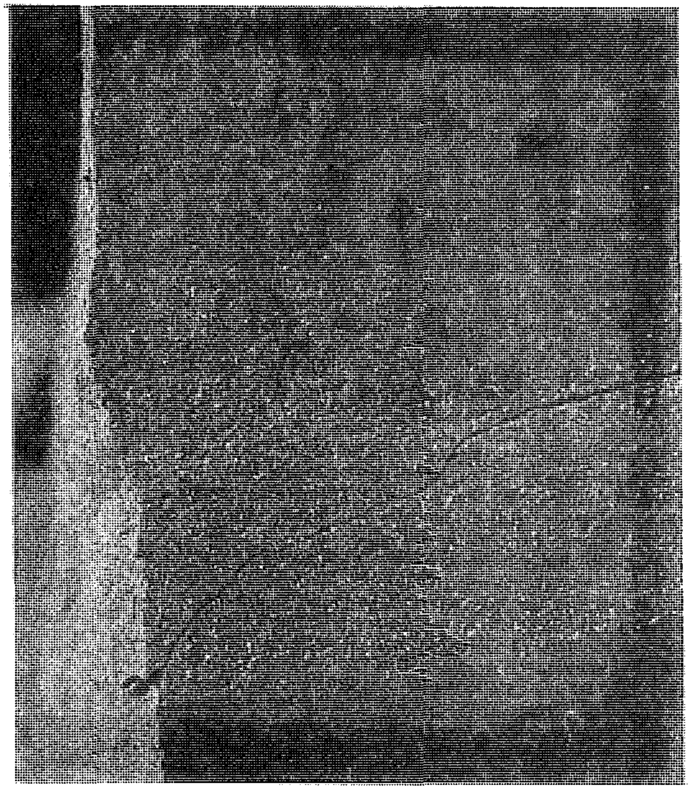


Figure 4. Coarse-tail grading at the bottom of a thick sandstone facies bed. Black arrow is 1 cm long. Most of the grading within this 3.4 m thick bed takes place within the lower 9 cm of the bed.

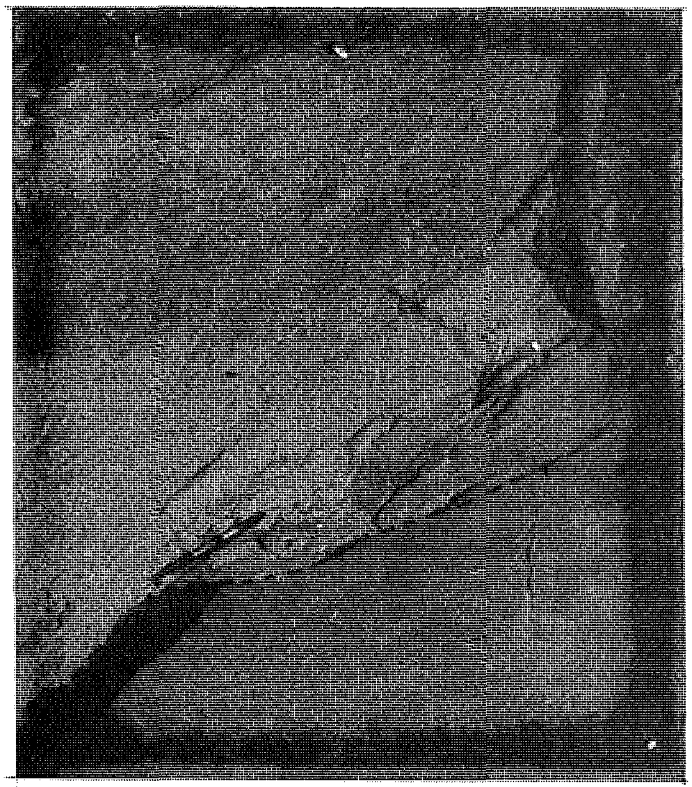


Figure 5. Amalgamated thick sandstone facies beds. The upper bed has cut into the lower bed (dashed line).

Figure 6. Rip-up clast (approximately 0.8 m long) near the bottom of a thick sandstone facies bed.

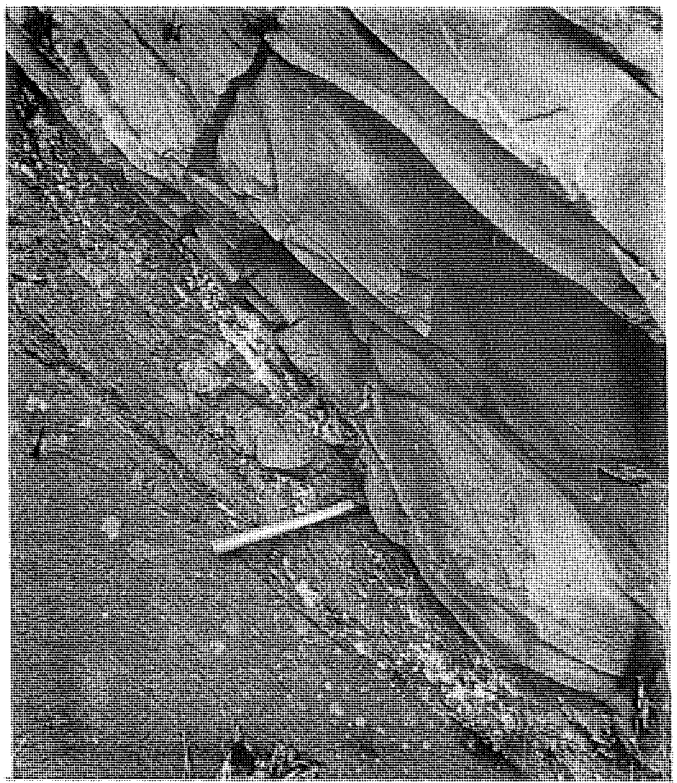


Figure 7. Erosive bottom contact of a thick sandstone facies bed and underlying mudstone.

Figure 9. Ripple laminations near the bottom of a thick-sandstone facies bed.

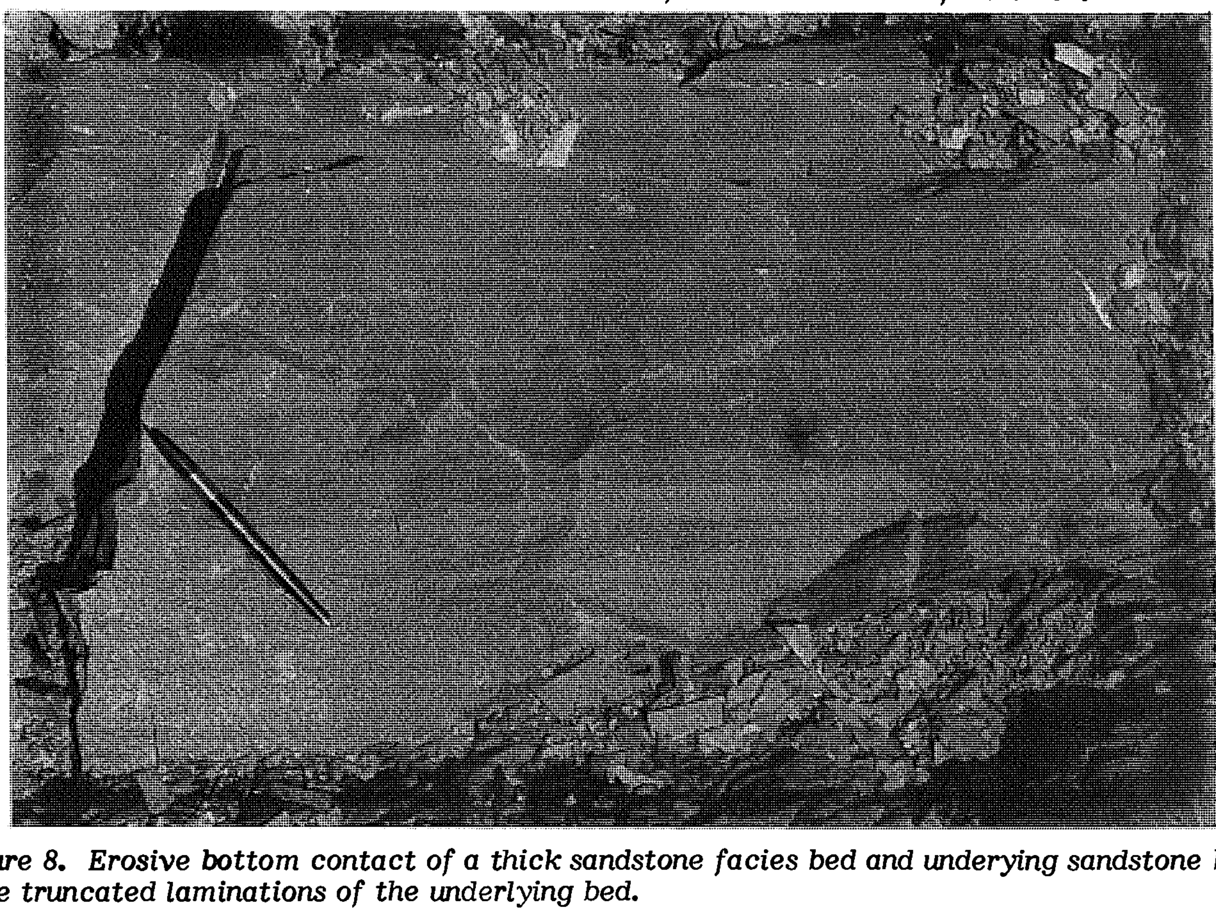


Figure 8. Erosive bottom contact of a thick sandstone facies bed and underying sandstone bed. Note truncated laminations of the underlying bed.

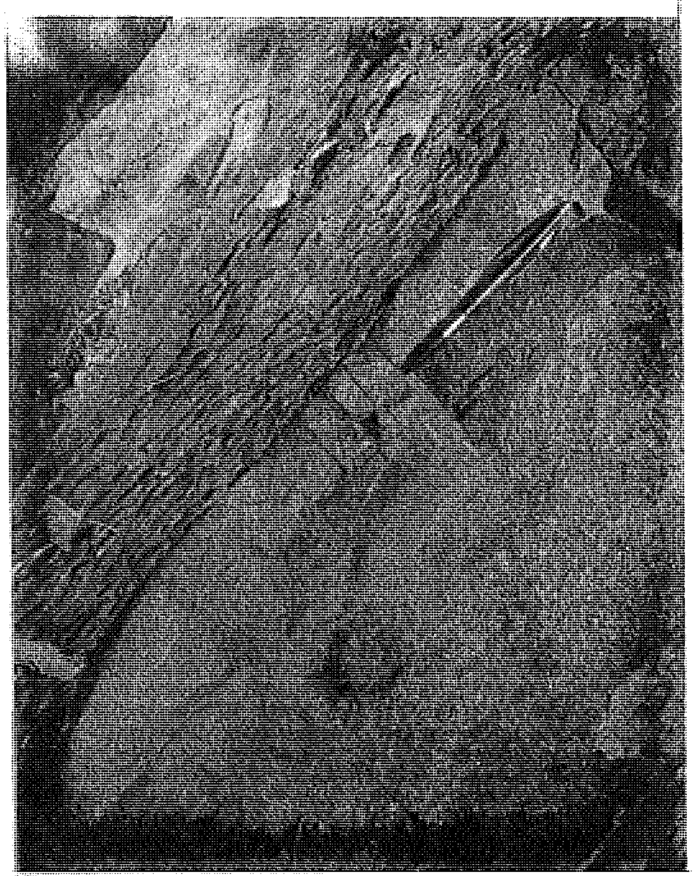


Figure 10. Dish structure at the bottom of a thick sandstone facies bed.

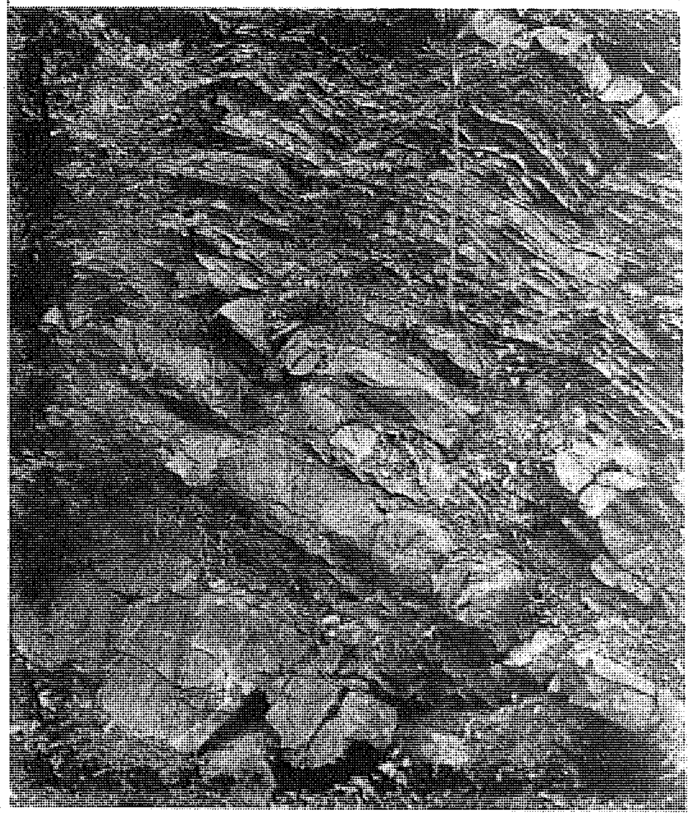


Figure 11. Turbidite-shale facies beds (lower left part of the photograph).

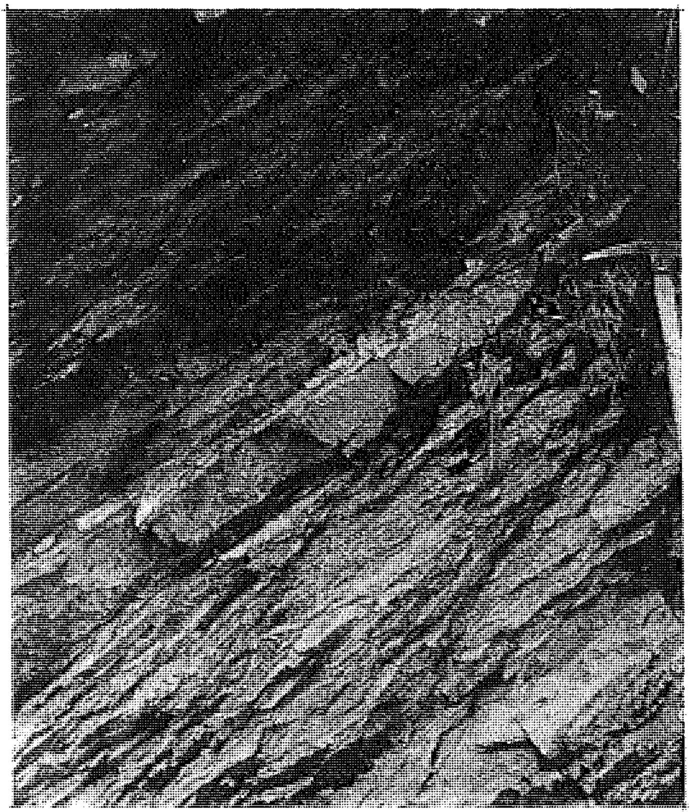


Figure 12. Discontinuous siltstone-shale facies beds.

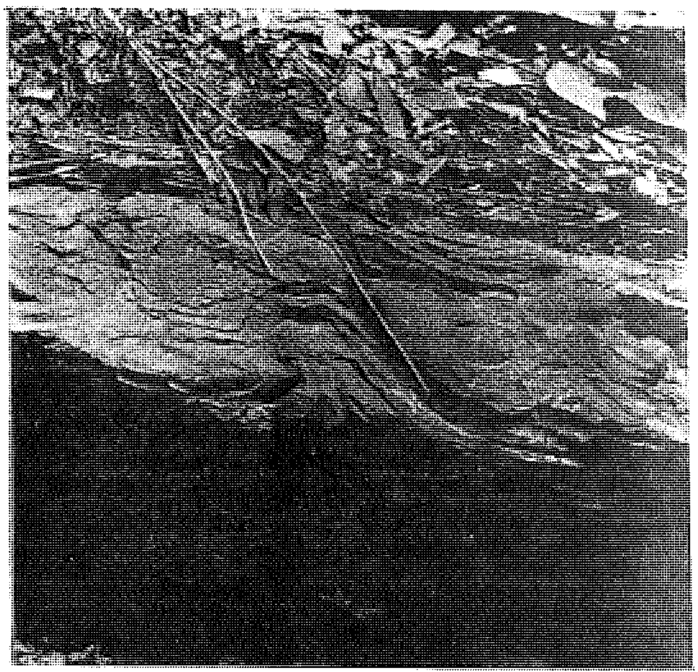


Figure 13. Tc(d)e Bouma sequence characteristic of siltstone-shale facies beds. Note pen at left for scale.

than 1 m thick. Individual turbidite beds are characterized by Ta/e and Ta/c/e and lesser proportions of Tabcde and Tab/e partial Bouma sequences (refer to fig. 2 for explanation). Many of the beds illustrate coarse-tail grading, and typical sole marks include scour and tool marks (groove, bounce, and prod marks) with superimposed load features. A minor proportion of these beds are amalgamated, and shale clasts are locally abundant near the bottoms of some of the beds. Beds of this facies are generally continuous within an outcrop although locally they can be seen to vary in thickness and even to pinch out (fig. 12).

The turbidite-shale facies is identical to Facies C of Mutti and Ricci-Lucchi (1975). The prime depositional mechanism of these deposits is both high and low concentration turbidite currents. Some of the Ta/e partial Bouma sequences are capped by lag deposits that are suggestive of reworking of top surfaces of a concentrated flow by the "entrained layer" (Middleton and Hampton, 1976). The major differences between this facies and the thick sandstone facies is the difference in thickness and the presence of obvious normal grading and identifiable partial Bouma sequences.

Siltstone-shale facies.

The siltstone-shale facies is characterized by monotonous sequences of interbedded siltstone and shale. Individual sequences range up to 3 m (10 ft) thick and are generally continuous within outcrop width of up to 15 m (50 ft). The majority of siltstone beds can be classified as Tc/d/e and Td/e partial Bouma sequences (fig. 13) although lesser proportions of Tb/c/d/e sequences are also present. Ripple laminations, climbing ripples, and parallel horizontal laminations are most common primary structures. The majority of climbing ripples present resemble Type C of Jopling and Walker (1968) (fig. 14). Bases of these beds are sharp and may display small flute and groove casts and load casts. Penecontemporaneous deformation structures, such as convoluted laminations and sedimentary boudinage, are common (fig. 15).

The siltstone-shale facies of the Windsor Township Formation is similar to Facies D1 and D2 of Mutti and Ricci-Lucchi (1975). It is the least common turbidite facies of the Windsor Township and constitutes only 15 percent of the sections measured to date. Low density turbidity currents are apparently responsible for transportation of these deposits. The siltstone-shale facies is distinguished from the other facies by the presence of partial (base-missing) Bouma sequences and the fine grain size.

Shale facies.

The shale facies of the Windsor Township Formation is arbitrarily defined as sequences with shale intervals greater than 1 m (3 ft) thick that overlie sandstone and silt beds. These rocks are characterized by poorly cleaved to fissile light-olive-gray (5Y 5/2) to olive-gray (5Y 3/2) and grayish-olive (10Y 4/2) silty shale, shale, and mudstone. Lesser medium-dark-gray (N4) to dark-greenish-gray (5GY 4/1) silicified shale and mudstone are interbedded with the olive and gray shale. The silty shale is typically laminated. Associated with the pelitic rocks are well sorted, parallel- to cross-laminated, locally discontinuous siltstone beds that resemble Tc/e and Tde partial Bouma sequences.

The shale facies is identical to Facies G of Mutti and Ricci-Lucchi (1975). The predominant depositional mechanism is hemipelagic sedimentation and/or gravity-driven sediment clouds such as nepheloid layers or submarine channel-related sediment plumes influenced by bottom currents. The shale deposits are similar to terrigenous deposits that accumulate on continental rise areas (Bouma and Hollister, 1973). Modern terrigeneous deposits are typically green or greenish-gray to black in color and composed of more than 30 percent continentally derived sand and silt-size sediment. The thin discontinuous silt layers are similar to modern contour current deposits (Stow, 1979; Chough and Hesse, 1982).

The lateral and vertical relations of the different facies described above form the basis for paleoenvironmental interpretation of these rocks and development of a submarine fan model. Details of the fan morphology which are based upon comparison of the sedimentologic characteristics of the Windsor Township

Figure 14. Climbing ripples within siltstone shale facies beds.

Formation and modern and ancient submarine fan models and sequences are discussed in a later section.

Red shale units.

Lenticular and tabular bodies of red shale interbedded with micritic limestone and chert constitute about 15 percent of the Greenwich slice. The average thickness of these units is 40 to 60 m (130-200 ft), although one sequence which may have been thickened by faulting is about 175 m (575 ft) thick.

The red shale units consist predominantly of dusky-red (5R 3/4) to grayish-red (10R 4/2) and grayish-purple (5P 4/2) mudstone and shale to micaceous shale and minor thinly-laminated siltstone. Red shale is locally interbedded with grayish-yellow (5Y 8/4) and pale-green (10G 6/2) to light-green (5G 7/4) mudstone to micaceous shale.

Sequences of grayish-black (N2) to medium-dark-gray (N4), light-gray (N7) weathering thin-

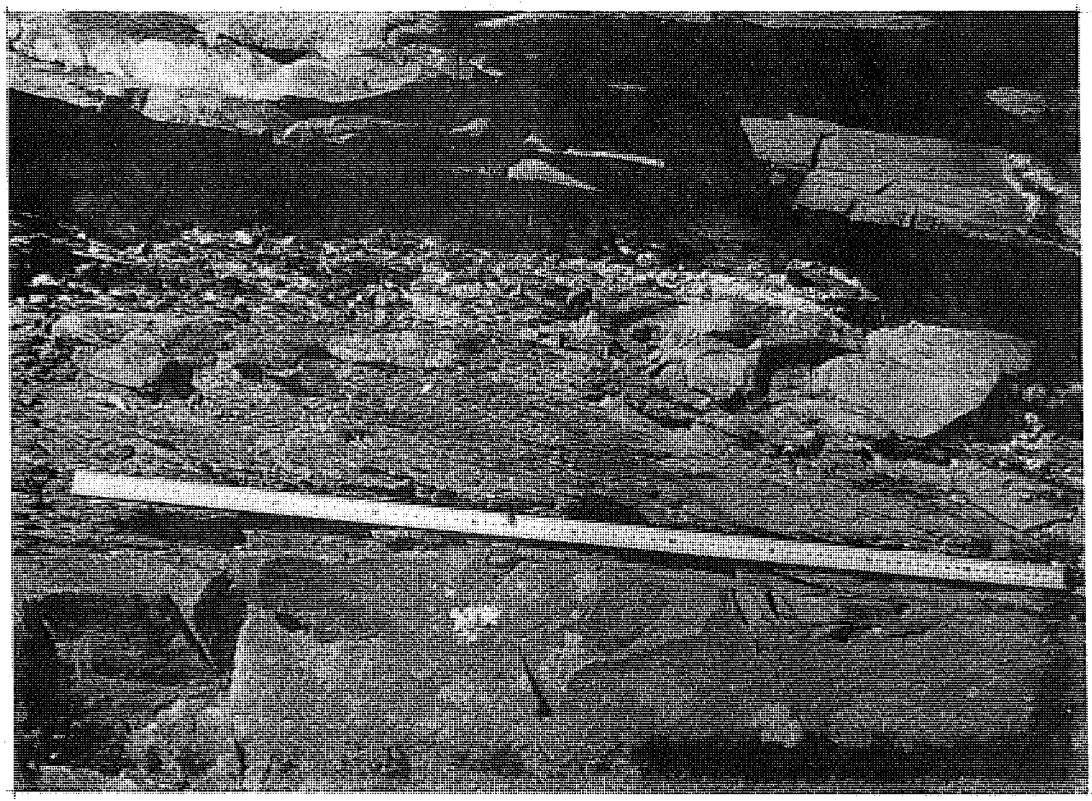


Figure 15. Disrupted (penecontemporaneously) siltstone-shale facies bed.

Figure 16. Starved calcisiltite ripple megaclast surrounded by red and green shale. White scale at right is approximately 12 cm.

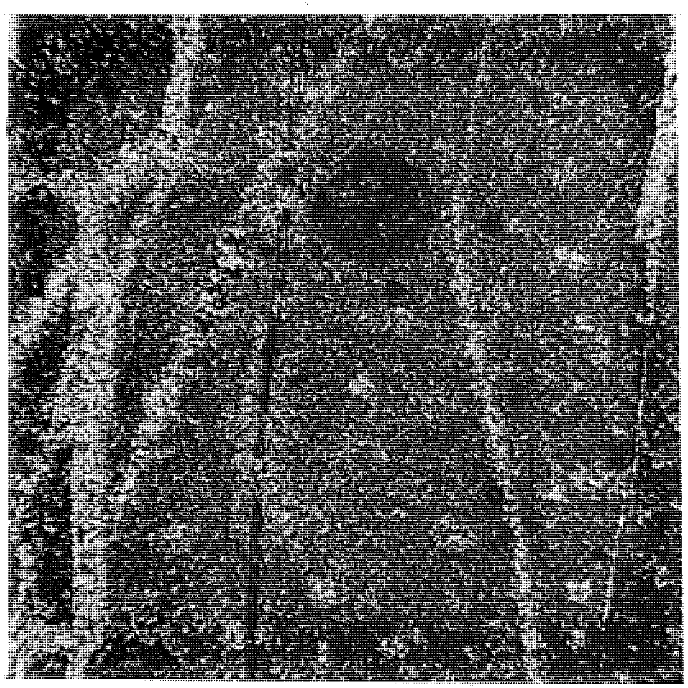


Figure 17. Radiolaria ghosts (white circles) in chert (56x). This sample is cut by chalcedony veins.

bedded calcilutite to calcisiltite are present locally within sequences of red and green shale. Sedimentary structures are limited to millimeter-scale, parallel- and ripple-laminations and climbing ripples. Locally the limestones are severely contorted.

The presence of parallel- and ripple-laminations is suggestive of deposition by dilute turbidity currents. Additional evidence for this is the presence of more than one size class of conodonts within these beds (J. E. Repetski, pers. comm., 1983), suggesting mixing of different size conodonts during turbidity current transportation and arguing against transportation by a tractive bottom current, which would promote a higher degree of sorting. Finally, the Balto-Scandic conodont fauna in these limestones (Bergstrom and others, 1972) is suggestive of sedimentation in a deep, cold water environment (Bergstrom, 1972; Bergstrom and others, 1972; Sweet and Bergstrom, 1972).

A limestone slide deposit or megaclast approximately 25 m (80 ft) long surrounded by red and green shale has been described by Epstein and others (1972) from an outcrop about 1.5 km east of Lenhartsville on Old Rt. 22 in the Kutztown quadrangle. The megaclast is composed of starved (fig. 16) and climbing ripples of light-gray (N7) calcisiltite surrounded by laminated dark-gray (N3) to medium-dark-gray (N4) calcareous shale. The megaclast and numerous lithologically similar clasts apparently slid into the red and green shale that surrounds them.

Discontinuous laminations of well-sorted quartz siltstone are interlayered with the red and green shale. Petrographic examination of these deposits indicates a predominance of very angular quartz and minor feldspar grains associated with variable amounts of dolomite (or ankerite) rhombohedra. In some samples millimeter-scale laminations of dolomite rhombs alternate with quartzose laminations.

Chert constitutes a minor but important rock type of the red shale units. Individual beds range from 1 to 35 cm (0.4-14 in) thick and are continuous to discontinuous within outcrop width. The most common variety is a medium-gray (N5) chert. Thin-section examination shows the presence of radiolaria, consisting of spherical mosaics (0.04 mm in diameter) of recrystallized chalcedony (fig. 17). Spines can be seen on some specimens. The matrix consists of light-brown, semi-opaque microcrystalline quartz and dolomite rhombohedra. A second variety is thin to thick beds of very-pale-orange (10YR 8/2) and very-pale-green (10G 8/2) chert. These beds are composed of dark, isotropic microcrystalline quartz, dolomite rhombs, and minor detrital quartz grains. Small (0.01 mm in diameter) spherical masses of chlorite give this variety a blotchy appearance in thin section. McBride (1962) suggested that the chlorite grains are pseudomorphs of radiolaria. A particularly interesting lithology is medium-gray (N5) siliceous shale which weathers grayish-olive-green (5GY 3/2). Radiolaria similar to those found in the medium-gray chert are also present. Some of the radiolaria are well preserved and spines can clearly be seen. In addition, sponge spicules in the form of elongate quartz grains are present.

Structures resulting from soft sediment deformation are common to rocks of the red shale units (fig. 18). Slump folds are well

Figure 18. Deformed red pelagic mudstone (dark) and dolomitized limestone and dolomitic chert (white) beds.

exposed in an outcrop of red mudstone and white dolomitic chert about 760 m (2490 ft) southwest of Albany on Rt. 143 in the Kutztown quadrangle. At this locality approximately 6 m (20 ft) of sediment have been deformed into disharmonic folds that are, in some cases, cut by an axial planar cleavage. Cleavage within the fold hinges displays flow patterns suggesting that the clay was squeezed out of the fold (Davies and Cave, 1976). Moreover, undeformed sediment can be seen draping the folds suggesting a soft sediment origin (Naylor, 1981). Many of the folds have rounded outer hinge perimeters and tight interlimb angles which is also suggestive of soft sediment folding (Naylor, 1981). Most of the boudins associated with the folds are gently tapered suggesting that they formed soon after deposition (Naylor, 1981).

The red and green shale and mudstone deposits of the Greenwich slice were originally interpreted as shallow water, continental sediments (Willard, 1939, 1943). McBride (1962), however, maintained that the association of red shale, graywacke, and radiolaria-bearing chert suggested a deep marine environment. The red shale of the Greenwich slice is an ancient analogue of red and brown clays presently accumulating in the deepest parts of the oceans and are considered to be pelagic or eupelagic deposits (Berger and von Rad, 1972). Modern pelagic red clay deposits are generally composed of wind-blown, continent-derived silt-size quartz and feldspar grains, clay minerals, zeolites, volcanic and altered volcanic debris, and residue from dissolved planktonic tests (Bouma and Hollister, 1973). Berger and von Rad (1972) defined pelagic or eupelagic clays as those pelagic sediments that contain less than 30 percent $CaCO_3$ and siliceous fossils. The normal accumulation rate of these deposits is 0.1 to 1.0 mm/1000 yrs although rates as high as 7.0 (Cocos Plate, DSDP Site 487, Watkins and others, 1981) to 15 mm/1000 yrs (nothern Pacific, Opdyke and Foster, 1970) have been reported.

Geochemical studies of the red and light-green pelagic mudstone of the Greenwich slice (Wright and Feeley, 1979) indicate that they are similar in composition, particularly total iron content, except for Fe^{+3}/Fe^{+2} ratios (compare columns 1 and 2, Table 1). An additional and equally important difference is the Mn content. The light-green mudstone and shale contains an average of five times less Mn as the red mudstone. The relatively low Fe^{+3}/Fe^{+2} ratio and low Mn content of the light-green mudstone are consistent with deposition and/or diagenetic alteration under reducing conditions. Reducing conditions evidenced by the light-green mudstone and shale probably occurred as a result of (1) relatively rapid sedimentation and trapping of marine organic material such that slow oxidation did not occur and/or (2) deposition in an area of coastal upwelling (Kennett, 1982). Both mechanisms were probably influential during sedimentation and early diagenesis of the light-green shale. For example, thin layers of light-green shale associated with thick sequences of red shale may represent beds deposited at at significantly greater rate than associated red shale. The relatively high Ca content (Table 1) may also indicate a minor volcanic contribution. The common association of light-green shale and olive-green hemipelagic siltstone is suggestive of sedimentation under the influence of coastal upwelling. These areas close to the continental margin are characterized by increased biological productivity and, therefore, increased organic content (Kennett, 1982). The high organic content would tend to promote reducing conditions. Indeed, the average composition of the light-green mudstone and shale is similar to that of "nearshore" muds (Table 1, column 3). The high Fe^{+3}/Fe^{+2} ratio and high Mn content of the red shale, on the other hand, are consistent with slow settling of fine-grained sediment in an oxygenated water column. The red and brown color of lithologically similar Holocene pelagic sediments is the result of amorphous or poorly crystalline coatings of iron oxide forming on slowly settled sediment particles (Kennett, 1982).

Red mudstone and shale is deposited in a number of different tectono-depositional environments. These include (1) active oceanic spreading centers, (2) marginal basins, (3) volcanic-influenced deep sea, and (4) deep sea. Comparison of Greenwich slice red shale deposits to basal sediments of the Cocos Plate (column 4, Table 1) indicates significant differences in Al, Mn, and total Fe that do not favor sedimentation near an active spreading

Table 1. Comparative Geochemistry of Greenwich Slice Pelagic Deposits from Various Tectonic Settings

	1	2	3	4	5	6	7	8
Si	29.9	30.2	31.0	-	18.8	24.5	23.4	27.3
Al	10.1	9.0	8.9	3.7	5.8	8.5	7.9	11.1
Fe^{+3}	4.0	1.1	2.4	-	-	-	5.6	-
Fe^{+2}	1.0	2.8	1.8	-	-	-	3.3	-
Mn	0.3	0.06	-	5.6	1.6	0.2	0.8	0.6
Mg	1.3	1.7	1.5	-	2.4	2.6	4.5	1.4
Ca	0.7	1.4	0.4	-	7.1	6.2	4.0	0.5
Na	1.3	1.2	0.8	-	5.9	1.8	1.5	0.9
K	3.4	2.5	3.0	-	0.9	0.3	1.4	3.3
Fe (total)	5.0	4.8	5.3	14.8	11.7	8.6	10.0	6.8

1. *Red Shale from Greenwich Slice; from Wright and Feely, 1979.*
2. *Green Shale from Greenwich Slice; from Wright and Feely, 1979.*
3. *"Nearshore muds"; El Wakeel and Riley, 1961.*
4. *Basal Cocos Plate; Leggett, 1981.*
5. *Lau Basin; Bertine, 1974.*
6. *Marianas Back Arc Basin (DSDP Site 454); Desprairias, 1981.*
7. *Volcanic sediment from Pacific Plate (area in front of Marianas Arc); El Wakeel and Riley, 1961.*
8. *Average red shale (eastern Atlantic Ocean); El Wakeel and Riley, 1961.*

center. Red and brown mudstone deposits have been described from marginal basins associated with volcanic arcs (Karig and Moore, G. F., 1976). Marginal basin deposits such as those of the Lau and Marianas basins are significantly enriched in Mg, Ca, and total Fe and depleted in Si, Al, and K relative to the Greenwich slice red shale deposits (compare columns 1, 5, and 6, Table 1). A marginal basin site of sedimentation, then, is also not favored. Early concepts of deep sea sedimentation (Murray and Renard, 1891) maintained that volcanic detritus transported by the wind was a major component of red and brown pelagic mudstone. These deposits, compositionally similar to marginal basin red and brown mud, are readily differentiated from red shale of the Greenwich slice (compare columns 1 and 7, Table 1). The composition of the Greenwich slice red shale deposits is most similar to red mudstone recovered from the eastern Atlantic Ocean (compare columns 1 and 8, Table 1) which was derived from eastern North America. The red shale of the Greenwich slice, like the eastern Atlantic pelagic deposits, shows virtually no evidence of volcanic debris. This, as will be seen later, is consistent with petrographic characteristics of associated sandstone deposits of the Windsor Township Formation.

It is inherently dangerous to speculate on depths of sedimentation of ancient pelagic deposits. Red and brown clay is currently accumulating at depths greater than 4,000 to 5,000 m where calcareous skeletal debris has been dissolved below the calcite compensation depth (CCD). Although the present CCD is found at a mean depth of 4.5 km (Kennett, 1982) it is likely that it fluctuated in the past in response to variations in eustatic sea level, climate, and surface productivity. Little is known of the CCD in Paleozoic time if it even existed. Indeed, Worsley and others (1983) suggested that the CCD formed approximately 220 m.y. ago in response to the advent of calcareous plankton. In any event, the assemblage of red and green shale and radiolaria-bearing chert and siliceous shale of the Greenwich slice is suggestive of sedimentation below the CCD. However, the micritic limestone associated with the red shale argues for deposition above CCD. Carbonate dissolution in the deep sea will affect such carbonate turbidites very little, so perhaps they are fortuitously preserved.(Hesse, 1975). A conservative estimate for the depth of sedimentation of these deposits based upon comparison to lithologically similar sediment accumulating in modern oceans, then, is 4.0 km.

In summation, the red and green shale of the Greenwich slice represents the slow accumulation of wind-blown, land-derived clay and fine silt at depths probably in excess of 4.0 km (2.5 mi) in an open ocean. Dilute turbidity currents transported fine-grained carbonate probably derived from the North American shelf into this setting. The fact that all of the carbonate beds are Early Ordovician in age suggests that external events such as the proposed Early Ordovician eustatic sea level rise (Barnes, 1982) were responsible for deposition of carbonates at this time. Indeed, numerous workers (Kier and Pilkey, 1971; Neuman and Land, 1975; Schlager and Ginsburg, 1981) maintain that increases in sea level lead to increased productivity on shelves which results in increased carbonate slope and off-slope sedimentation rates. Pelagic accumulations of radiolaria may have supplied much of the silica for the chert and siliceous shale beds.

Boulder conglomerate.

Polymict boulder conglomerate crops out along Rte 737 approximately 2 km (1.2 mi) north of Kutztown in a discontinuous exposure approximately 1,450 m (4760 ft) long. The conglomerate contains angular clasts and megaclasts of graywacke (figs. 19 and 20) and rounded clasts of dark-greenish-gray (5GY 4/1) mudstone and shale, grayish-black (N2) to dark-gray (N4) parallel- and ripple-laminated calcilutite and calcisiltite, dusky-red (5R 3/4) and grayish-red (10R 4/2) shale and mudstone and pale-green (10G 6/2) shale and mudstone, all of which are lithologies common to the Greenwich slice. The clasts are in a matrix of medium-dark-gray (N4) to dark-greenish-gray (5GY 4/1) and grayish-olive (10Y 4/2) to pale-green (10G 6/2) shale and mudstone. Graywacke clasts are dominant and range from less than 2 cm (0.8 in) to 10 m (33 ft) or more in length. Some of the larger clasts are crudely imbricated.

Detailed study of exposures along Rt 737 indicates that the conglomerate is zoned with respect to clast lithology and size. Moving north along Rt 737, the first 305 m (1000 ft) of the conglomerate contains clasts of shale, mudstone, siltstone, and limestone. The largest clast is about 0.3 m (1 ft) long. These rocks grade into 915 m (3000 ft) of conglomerate characterized by extremely large clasts or megaclasts of graywacke and fewerr smaller clasts of graywacke, mudstone, and limestone. Some of the clasts are so large that their full size cannot be seen. They are probably about 1 to 10 m (3-30 ft) long. The next 30 m (100 ft) of conglomerate contains small (up to 30 cm (12 in) calcilutite clasts almost to the exclusion of all other types. The matrix for these clasts is the distinctive pale-green (10G 6/2) mudstone that is found in the red shale units. The remainder of the outcrop contains small (up to 75 cm (30 in) angular clasts of graywacke, shale, and limestone.

As stated above, all clasts and matrix of the boulder conglomerate are Greenwich slice lithologies. In particular, the graywacke clasts are texturally and mineralogically different from the graywacke of the Ramseyburg Member (Drake and Epstein, 1967) of the Martinsburg Formation to the south of the allochthon. In addition, the pelitic matrix of the boulder conglomerate is not the dark-gray claystone-slate that is typical of the Bushkill and Ramseyburg Members of the Martinsburg but is like that of the Windsor Township Formation. This distinction becomes very important in determining whether or not the boulder conglomerate represents a wildflysch deposit. Alterman (1972) first recognized this conglomerate and described it as a "wildflysch-

Figure 19. Angular graywacke megaclast surrounded by shale. Ruler at lower left of clast is 1m long.

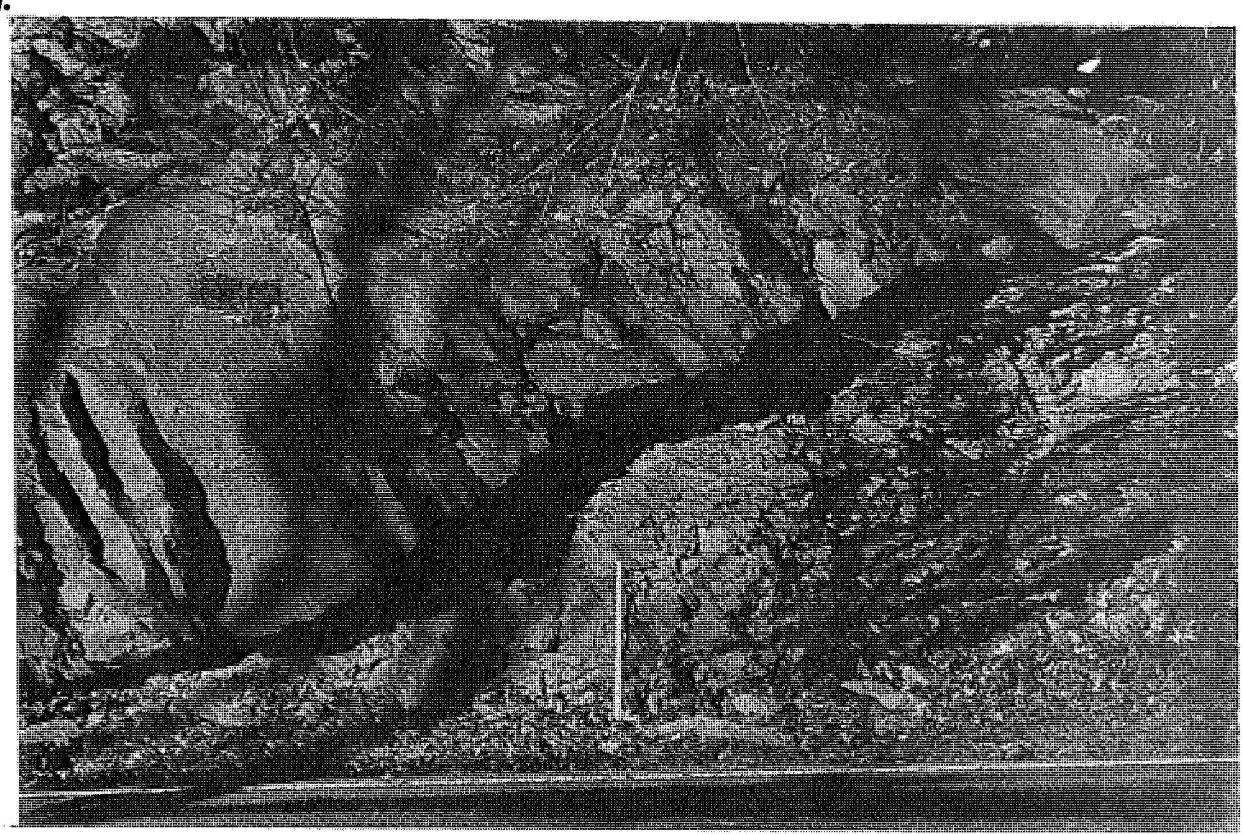


Figure 20. Irregular graywacke megaclast surrounded by cleaved mudstone. Scale is 1m long.

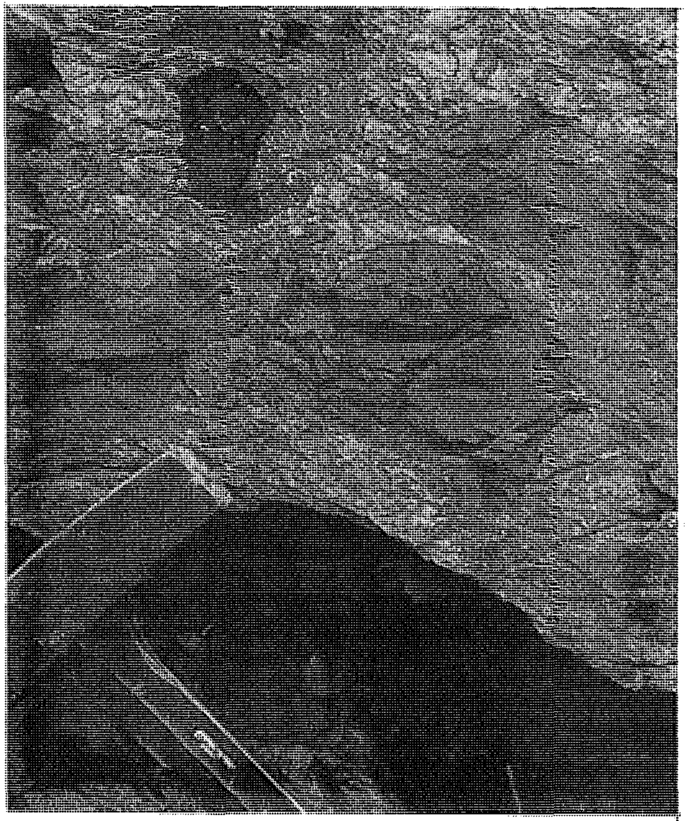


Figure 21A. Siliceous mudstone clasts within deformed mudstone matrix.

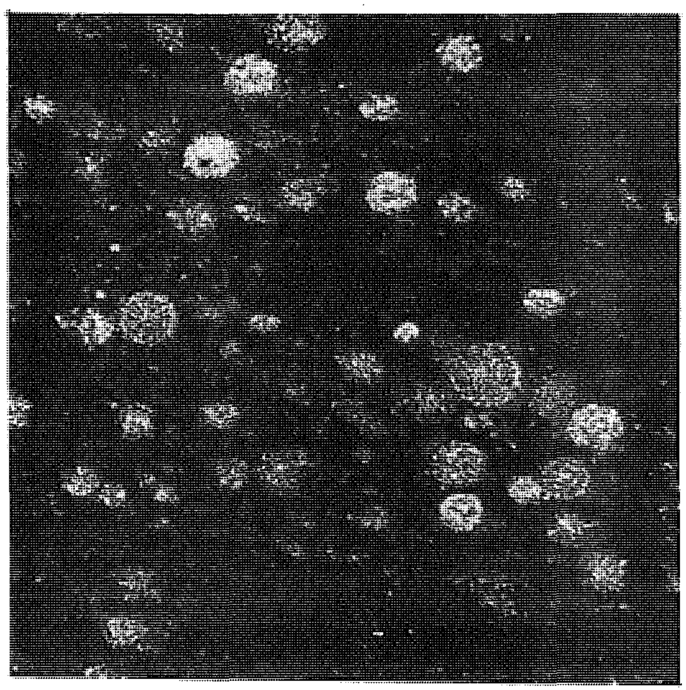


Figure 21B. Radiolaria ghosts (recrystallized chalcedony) within a mud matrix, sample from a siliceous mudstone clast (56x).

type" conglomerate which may have heralded the approach of the allochthon in much the same way as the wildflysch described by Bird (1969) and Zen (1972) heralded advance of the allochthonous slices of the Taconic Range of New York and New England. However, the characteristics of the boulder conglomerate suggest that it is a slumped level in the sense of Elter and Trevason (1973) with no input from sources outside the basin. The boulder conglomerate, therefore, should be considered as an intrabasin slump deposit and not a wildflysch marking the sole of the Hamburg klippe.

The great size of some of the clasts of the boulder conglomerate makes it analogous to modern submarine"blocky slide" deposits described by Jacobi (in press). On 3.5 and 12 KHz echograms, blocky slide deposits are characterized by irregularly spaced hyperbolae with vertex elevations that denote local relief on top of the flows. This is especially common near the "erosional" scarps which mark the source of the blocky material. Jacobi (1976) suggested that the hyperbolae associated with these deposits may represent large olistoliths present in the flows.

A second example of a conglomerate can be found south of Kempton in a borrow pit on the east side of Rt 737 in the Kutztown quadrangle. This exposure was described by Alterman (1972) who maintained that this conglomerate was another wildflysch deposit at the sole of the allochthon. This deposit consists of pebbles and boulders up to 2.5 m (8.2 ft) long of weathered dark-gray silicified shale and chert. The matrix for the clasts is olive-gray (5Y 3/2) to light-olive-brown (5Y 5/6) shale. Of particular interest is the presence of radiolaria-bearing chert and siliceous shale pebbles (fig. 21). The pebbles are light bluish gray (5B 7/1) to medium bluish gray (5B 5/1) on fresh surface and olive gray (5Y 3/2) to greenish gray (5GY 6/1) on weathered surface. Radiolaria in these pebbles average 0.04 mm in diameter. Some of the larger and better preserved radiolaria still have spines attached to them. The chert and silicified shale pebbles are all part of the Greenwich slice and are not exotic.

Sediment slide deposits similar to those of the Greenwich slice are not restricted to any particular type of tectonic environment. They have been described from many areas (for example, Jacobi, 1976; in press). Although deposits of active convergent margins have not been examined in as much detail as those of passive margins, it is apparent that a great number of slide deposits are present at these margins (Piper and others, 1973; von Huene, 1974; Coulbourn and Moberly, 1977; among others).

Age of the Greenwich slice

One of the early reasons for proposing the existence of allochthonous rocks within the Pennsylvania Great Valley was the recognition of anomalously old graptolites within pelitic rocks. Recent graptolite studies by Wright and Stephens (1978) and Wright and others (1979) and conodont studies by Bergstrom and others (1972), Repetski (1979), Lash (1980, unpub. data), and Lash and Drake (1984) have contributed immensely to the overall biostratigraphic understanding of the Greenwich slice. As discussed in the preceeding section the Greenwich slice can be separated into two major lithologies: (1) coarse-grained turbidite and other mass flow deposits and hemipelagic olive-green mudstone and (2) red and green pelagic mudstone, deep-water limestone, and radiolaria-bearing chert and siliceous shale. Synthesis of results of the biostratigraphic studies cited above along with the unpublished data indicates that it is also possible to separate these lithologies on the basis of biostratigraphic characteristics (fig. 22).

Wright and others (1978, 1979), Lash (1980), and Lash and Drake (1984) noted the presence of Nemagraptus gracilis Zone graptolite species and possibly Glyptograptus teretiusculus Zone (Wright and others, 1978, 1979) types within interbedded graywacke and olive-green siltstone and shale. Graptolites collected by Wright and associated workers were classified by Riva's (1972, 1974) classification whereas graptolites collected by USGS geologists and associated workers were classified by Berry's (1962, 1970, 1971) scheme. Of particular interest is the duration of the N. gracilis Zone of each classification. Riva's N. gracilis Zone is about 8 m.y. long whereas Berry's is only 5 m.y. long and spans the lower half of Riva's zone (fig. 22). This suggests that the majority of turbidite

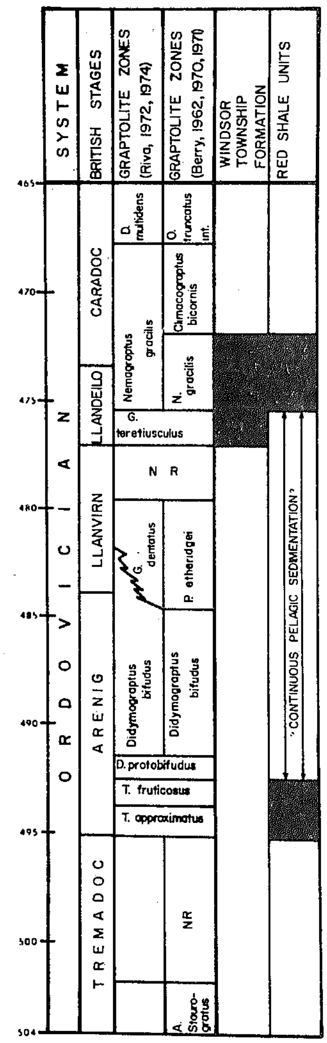


Figure 22. Age relations of rocks of the Greenwich slice. Modified from Ross and others (1982).

and hemipelagic deposits of the Greenwich slice were probably deposited during the lower half of Riva's N. gracilis Zone over a relatively short time span of about 3 to 5 m.y. (fig. 22). The age of the red shale units is somewhat more of a problem. Stephens and Wright (1982) noted the sparse presence of N. gracilis Zone graptolites in some red shale beds. However, Bergström and others (1972), Repetski (1979), and Lash (1980, unpub. data) recovered Early Ordovician (late Tremadocian-early Arenigian) conodonts of Balto-Scandic affinities from limestone interbedded with red shale suggesting that some of the red shale is Early Ordovician in age. Significantly, Perissoratis and others (1979) report that graptolites from rocks of the

Jutland klippe in New Jersey, a sequence of rocks lithologically similar to the red shale units of the Greenwich slice, range in age from Early to Middle Ordovician. The accumulated data for the pelagic red shale units, then, suggests that they range in age from Early to Middle Ordovician and were probably deposited over a 20 m.y. period (fig. 22), significantly longer than that of the somewhat younger turbidite and hemipelagic deposits noted above.

<u>Submarine fan sedimentology and morphology of Greenwich slice sediments</u>

This section describes the submarine fan characteristics of the turbidite and other mass flow deposits of the Windsor Township Formation of the Greenwich slice. Approximately 800 m (2600 ft) of flysch have been measured. Sections were chosen for width of exposure and from areas where structural control is known well enough to minimize measurement of parts of sections more than once. Bed thickness, rock type, bed contacts, internal primary structures, orientation of paleocurrent indicators, and maximum grain size were noted for each bed. A complete section was then graphically plotted to show internal organization of the different lithofacies described earlier. Depositional subenvironments could then be assigned to each lithofacies through comparison of their sedimentary attributes (rock type, facies association, cyclicity, vertical relations, etc.) with those of available fan models.

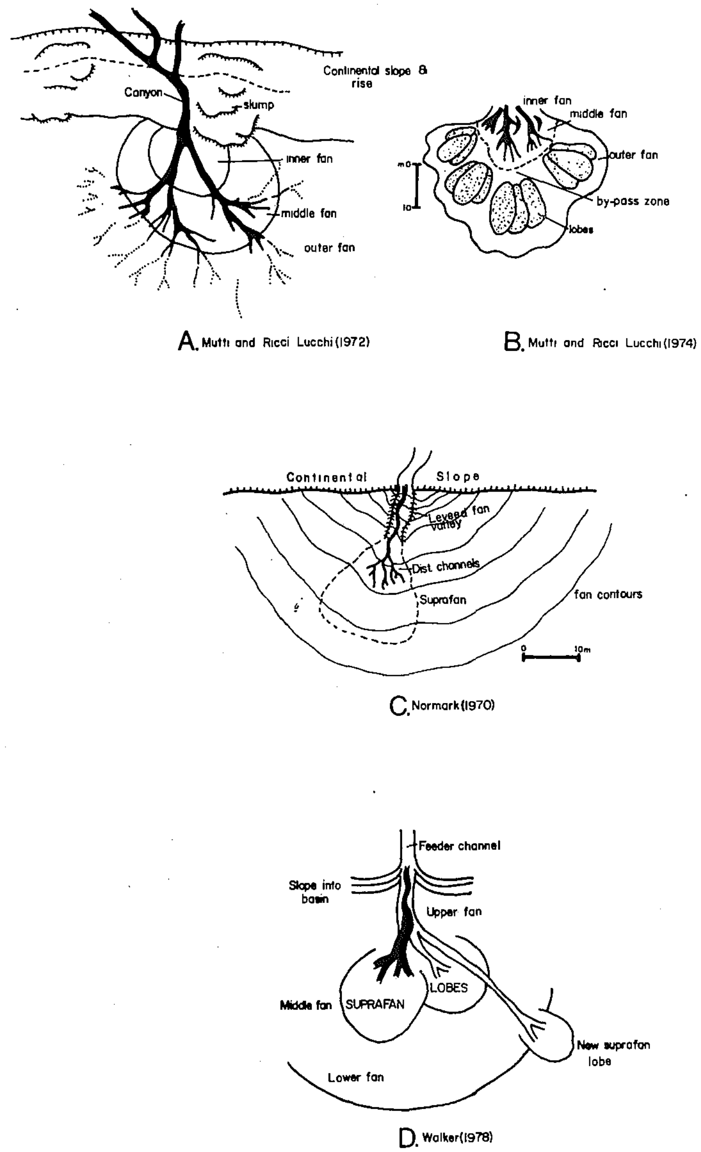


Figure 23. Major submarine fan models.

<u>Submarine fan models.</u>

Mutti and Ricci-Lucchi (1972) developed the first comprehensive model for ancient submarine fan sequences (fig. 23A). They combined data from modern and ancient submarine fan sequences (i.e., Marnoso-arenacea basin in the northern Apennines and the Hecho basin in the north-central Pyrenees) and recognized a three-fold subdivision of submarine fans; an <u>inner fan</u> characterized by a main leveed fan valley that connects upslope with a submarine canyon and splits downslope into a number of shallow, straight to sinuous, leveed <u>middle fan</u> distributary channels which feed <u>outer fan</u> depositional lobes. They recognized that each physiographic subdivision is characterized by a distinct facies association. The inner-fan association consists of large, lenticular coarse-grained sandstone and conglomerate bodies enclosed by hemipelagic sediment that results from the in-filling of major fan valleys. The middle-fan association consists of thin-bedded, fine-grained interchannel deposits enclosing lenticular, channel-fill sandstone bodies displaying fining-upward megasequences (Ricci-Lucchi, 1975) that apparently formed in response to channel abandonment. The outer-fan association consists of fine-grained, thin-bedded, fan-fringe deposits and thicker-bedded, non-channelized outer fan depositional sandstone lobes. Outer fan lobes, in particular, consist of progradational sheet-like sandstone bodies organized in thickening- and coarsening-upward megasequences (Ricci-Lucchi, 1975) that record progradation of depositional lobes formed at the mouths of middle-fan distributary channels (Mutti and Ghibaudo, 1972). Mutti and Ricci-Lucchi (1974, 1975) and Mutti (1977) refined this model by showing that in some cases outer fan depositional lobes are separated from feeder channels by a zone of sediment by-passing (fig. 23B). This later addition to the 1972 model has gained little support among most workers.

Bouma and Nilsen (1978) noted that the facies sequence developed by Mutti and Ricci-Lucchi is compatible with data from modern fans and added the concept of suprafans. The suprafan, first described by Normark (1970) from the San Lucas fan off southern California, is a depositional, convex-upward bulge located in the midfan area at the termination of the main leveed fan valley (fig. 23C). It is characterized by coarse-grained sediments, by the lack of persistent levees, and by the presence in its upper reaches of a braided and rapidly shifting distributary channel system.

Walker (1976a,b; 1978; fig. 23D) presented a facies model for ancient submarine fans based on the suprafan concept. Although inner, middle, and lower fans are still the major subdivisions, there are some important differences with the earlier models. This involves suprafan lobes. Each suprafan lobe has shallow, braided channels in its upper part and a smooth outer part that grades basinward into the outer fan and basin plain. The Mutti and Ricci-Lucchi outer fan depositional lobes are incorporated in the Walker middle fan environment and considered to be the smooth, outer part of the suprafan lobes. Massive and pebbly sandstone deposits are concentrated within the shallow, braided distributary channel system of the suprafan lobes, and fining-upward sequences are developed during channel abandonment and filling. "Proximal" classical turbidites are deposited in the outer part of the suprafan lobes and are internally organized into thickening-upward cycles.

The Mutti and Ricci-Lucchi and suprafan models form end member types of fans. Normark (1978, p. 929) noted that many modern fans have no development of suprafan features but instead have isolated, more stable, midfan channels. In addition, recent studies (Reynolds and others, 1983) have described intermediate or hybrid fans that contain aspects of both models. Normark (1978) ascribes the development of one fan type over another to the grain-size distribution of the sediment supplied to the fan. If little coarse sediment is supplied, "rapid deposition at the end of the fan valley may not occur; instead levee development by overbank deposition...results in relatively straight channels which will prograde rapidly downfan. In this case no middle fan convex-upward bulge would develop." A predominance of coarse sediment, on the other hand, would tend to result in poor levee development and a less stable distributary channel system. Canyon-fed deep sea fans, in particular, would tend to have a suprafan growth pattern while delta-fed fans, receiving large volumes of fine-grained material, would develop a more stable midfan channel system.

It is apparent from the discussion above that the type of deep-sea fan may reflects the grain-size distribution of sediment supplied to the fan. Recently, Nilsen (1980) has illustrated that it is possible to make the distinction between the two fan types in the field indicting that the distinction is real and should be of major concern when working in ancient submarine fan sequences.

Submarine fan facies.

Two submarine fan associations are recognized in the Greenwich slice: outer-fan depositional lobes and braided midfan deposits. The latter by far occupy the larger volume.

Outer-fan depositional lobes.—Outer-fan depositional lobes are non-channelized, constructional sandstone bodies organized in thickening- and coarsening-upward megasequences reflecting basinward progradation of channels (Ghibaudo, 1980; Mutti and Ghibaudo, 1972). They have been recognized in many ancient fan sequences both in outcrop (Ghibaudo, 1980; Ricci-Lucchi, 1975; Nilsen and Moore, 1979; among many others) and subsurface studies (Walker, 1978; Webb, 1981). In general, outer-fan lobes consist of thick-bedded, fine- to coarse-grained "classical" turbidite deposits associated with thinner and finer-grained beds that pass down-current and laterally into thin-bedded fan fringe deposits (Ghibaudo, 1980). Internal organization of individual lobes can be characterized by a simple thickening-upward cycle (negative megasequence of Ricci-Lucchi, 1975) but are more commonly characterized by repeated thickening-upward cycles that are reflective of complex depositional dynamics (multiple and composite cycles of Ricci-Lucchi, 1975). Minor channeling may be found on lobe tops or they may be overlain abruptly or gradationally by thin turbidite beds and hemipelagic mud reflecting sudden or gradual lobe abandonment.

Outer-fan lobes of the Greenwich slice are characterized by sequences of siltstone-shale facies and shale facies beds overlain by turbidite-shale facies beds (fig. 24). Individual lobes range from 2.5 to 16 m thick (8-50 ft). Lobe tops are either abruptly overlain by hemipelagic deposits or gradationally by thin turbidite beds. The majority of depositional lobes of the Windsor Township Formation are complex cycles that record rapid middle fan channel abandonment and/or channel thalweg shifting. Mutti and Sonnino (1981) have referred to the small-scale thickening-upward cycles observed within complex outer fan depositional lobe deposits as compensation cycles and related them to the smoothing of local topography and not necessarily to channel abandonment up slope. Small lobe-shaped bodies are built up at the end of a channel and subsequent flows smooth the local topography by flowing around these highs, thereby resulting in deposition of thin silt beds or hemipelagic mud on top of the previous lobe. Thus, lobe abandonment does not necessarily imply channel avulsion upslope.

In detail, individual lobes consist of thick- to very-thick-bedded, medium- to coarse-grained "classical" turbidites and thin-bedded, base-cut-out turbidite beds and hemipelagic shale (fig. 24). Amalgamated beds, although not common, do occur locally near the tops of some lobe sequences. Most of the beds near the top of lobe, the turbidite-shale facies, can be classified as Ta/c/e, Ta/c, and Ta partial Bouma sequences. These deposits correspond to C1

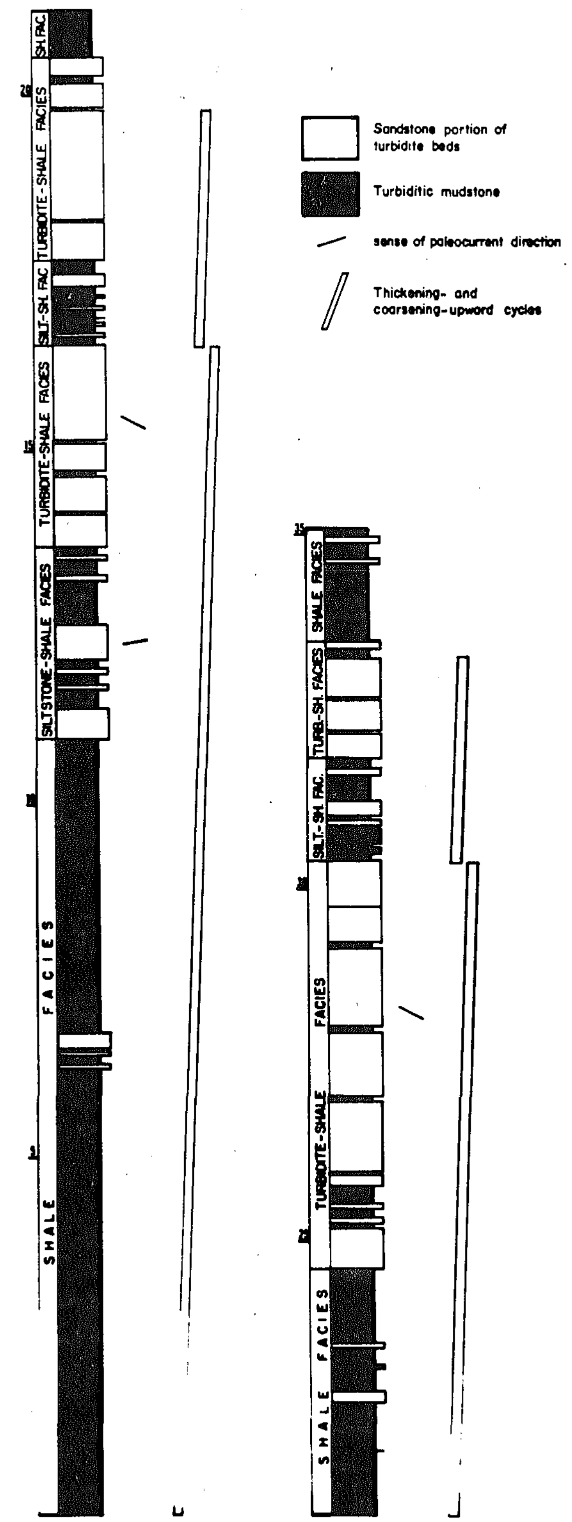


Figure 24. Example of the internal organization of depositional lobe deposits of the Windsor Township Formation. Section measured along Maiden Creek north of Dreibelbis Bridge (Hamburg quadrangle).

facies of Mutti and Ricci-Lucchi (1975) and are deposited by high-concentration turbidity currents.

The turbidite-shale facies deposits are underlain by sequences of hemipelagic mudstone (shale facies) and thin-bedded, ripple- and cross-laminated siltstone deposits (siltstone-shale facies) that are characterized by Tc(d)e, Tc/e, Tde, and lesser proportions of Tbc(d)e and Tbc/e partial Bouma sequences. The basal division (a) of the Bouma sequence, is notably absent. These deposits are comparable to Facies D beds of Mutti and Ricci-Lucchi (1975) and are apparently the result of dilute turbidity currents.

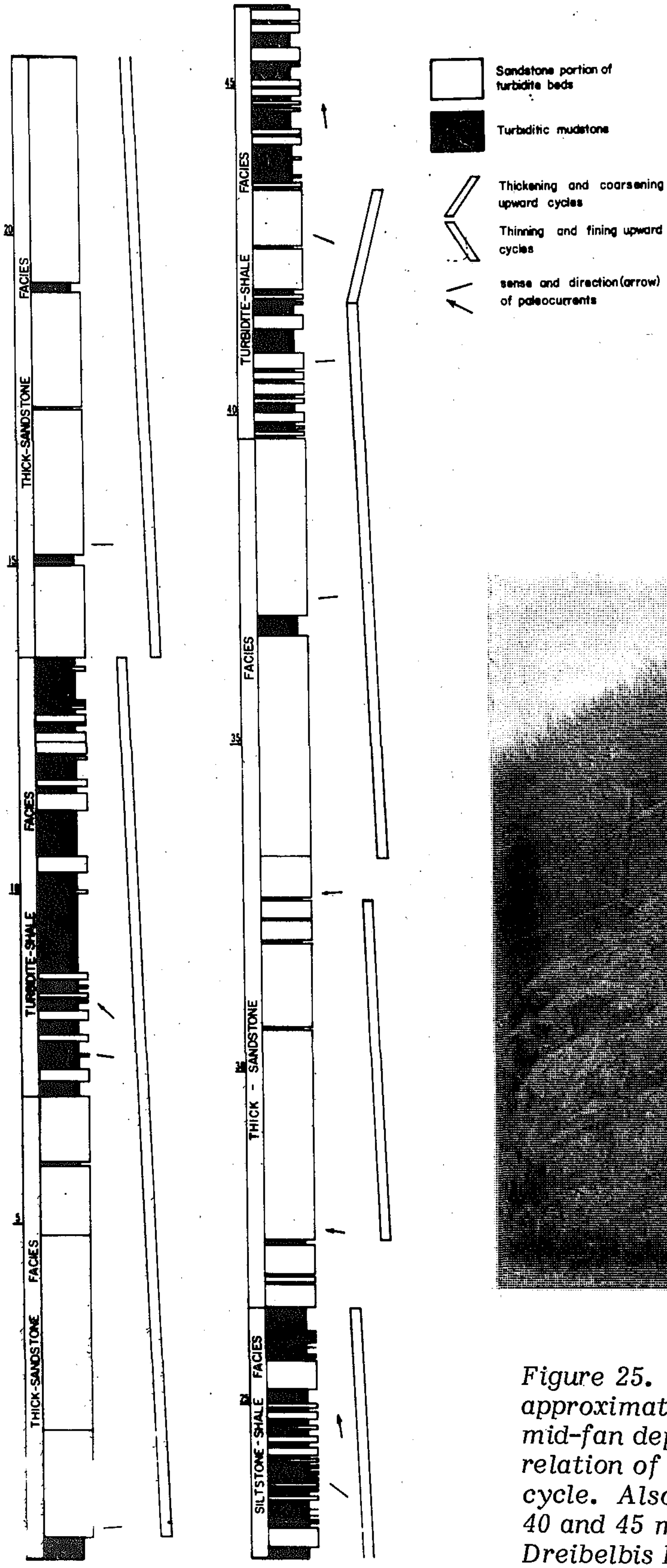
Braided midfan deposits.—Middle fan deposits are well documented from ancient submarine fan sequences (Mutti and Ricci-Lucchi, 1972; Walker, 1975). Thick-bedded, coarse-grained deposits filling intricate patterns of shallow channels have been described by Piper and Normark (1983), Ricci-Lucchi and Parea (1974), Walker (1975, 1976a,b, 1977, 1978), Surlyk (1978), and Pickering (1982).

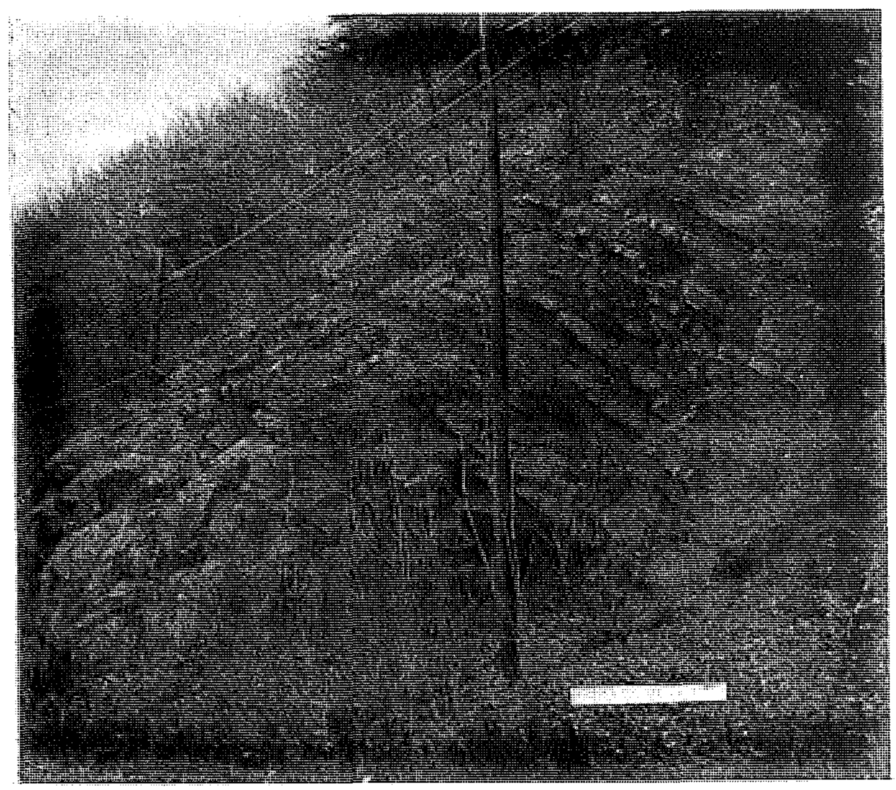
These workers observed that many channelized turbidite deposits display a thinning- and fining-upward trend which contrasts markedly with the thickening-upward trend of depositional lobe deposits described above. Most workers agree that the thinning- and fining-upward cycles termed "positive megasequences" by Ricci-Lucchi (1975) record shifting of middle fan distributary channels and as such are invaluable in deciphering ancient submarine fan sequences. Hiscott (1980), however, suggested that a more diagnostic feature of inferred channel deposits is the concentration of coarse amalgamated sandstone beds into packets separated by a sequence of turbidite-shale and/or siltstone-shale facies beds.

Middle fan distributary channel deposits of the Windsor Township Formation consist of thick-bedded, locally amalgamated sandstone beds overlain by turbidite-shale and lesser proportions of siltstone-shale facies beds (fig. 25). Three subfacies of the middle fan (Mutti, 1977) environment identified by characteristics of sandstone beds are present within the Windsor Township Formation: (1) channel axis, (2) interchannel, and (3) levee (fig. 26). Thinning-upward cycles are present generally as complex sets that range from 5 to 15 m (16 to 50 ft) thick. Beds in the lower part of the cycles consist of medium to granule size sediment. Upper contacts of these beds range from irregular surfaces with little or no interlayered shale in amalgamated sequences to regular and straight. Bottom contacts of many of these beds are erosive and locally cut down as much as 0.75 m (2.5 ft) into underlying sandstone or shale beds. A particularly interesting feature of these beds is the presence of large (1.5 m (5 ft)) shale rip-up clasts. They are generally concentrated in the lower parts of the beds, typically at the base. Mutti and Nilsen (1981) maintain that these are common to submarine channel deposits and record erosion of channel floors and walls (cut banks) and/or slumping and sliding of levee sediment into the channel. The lack of folded or faulted clasts suggests that the clasts are primarily derived from erosion of the channel floors and walls (Mutti and Nilsen, 1981).

The thick sandstone facies deposits are overlain by sequences of turbidite-shale and significantly lesser siltstone-shale facies beds resulting in thinning-upward megasequences. These beds are representative of interchannel and levee facies deposits, respectively. Of particular interest are the siltstone-shale facies beds which form bundles of Tcde and Tde turbidite beds up to 3.0 m (10 ft) thick although most are significantly thinner. These beds are interpreted as levee deposits related to channel spillover and should be considered separate from other interchannel deposits of the Windsor Township Formation. A levee environment is suggested for these deposits on the basis of: (1) lithologic similarities (i.e., grain size, primary structures) to modern levee deposits (Hesse and Chough, 1980; Damuth, 1979; Chough and Hesse, 1982); (2) vertical juxtaposition with channel deposits; (3) irregular bedding; (4) silty nature of the beds; (5) common occurrence of soft-sediment deformation. In addition, paleocurrent directions of these beds, as will be discussed below, often diverge 20 to 90^{o} from directions in associated channel sequences.

Although channel-fill thickness may not always match the original active channel depth, the thickness of a thinning- and fining-upward





A

B

Figure 25. A) Fining- and thinning-upward sequence. Scale approximately 3m long. B) Internal organization of braided mid-fan deposits of the Windsor Township Formation. Note relation of paleocurrent directions to position in a particular cycle. Also note the small thickening-upward cycle between 40 and 45 m. Section measured along Maiden Creek north of Dreibelbis Bridge (Hamburg quadrangle).

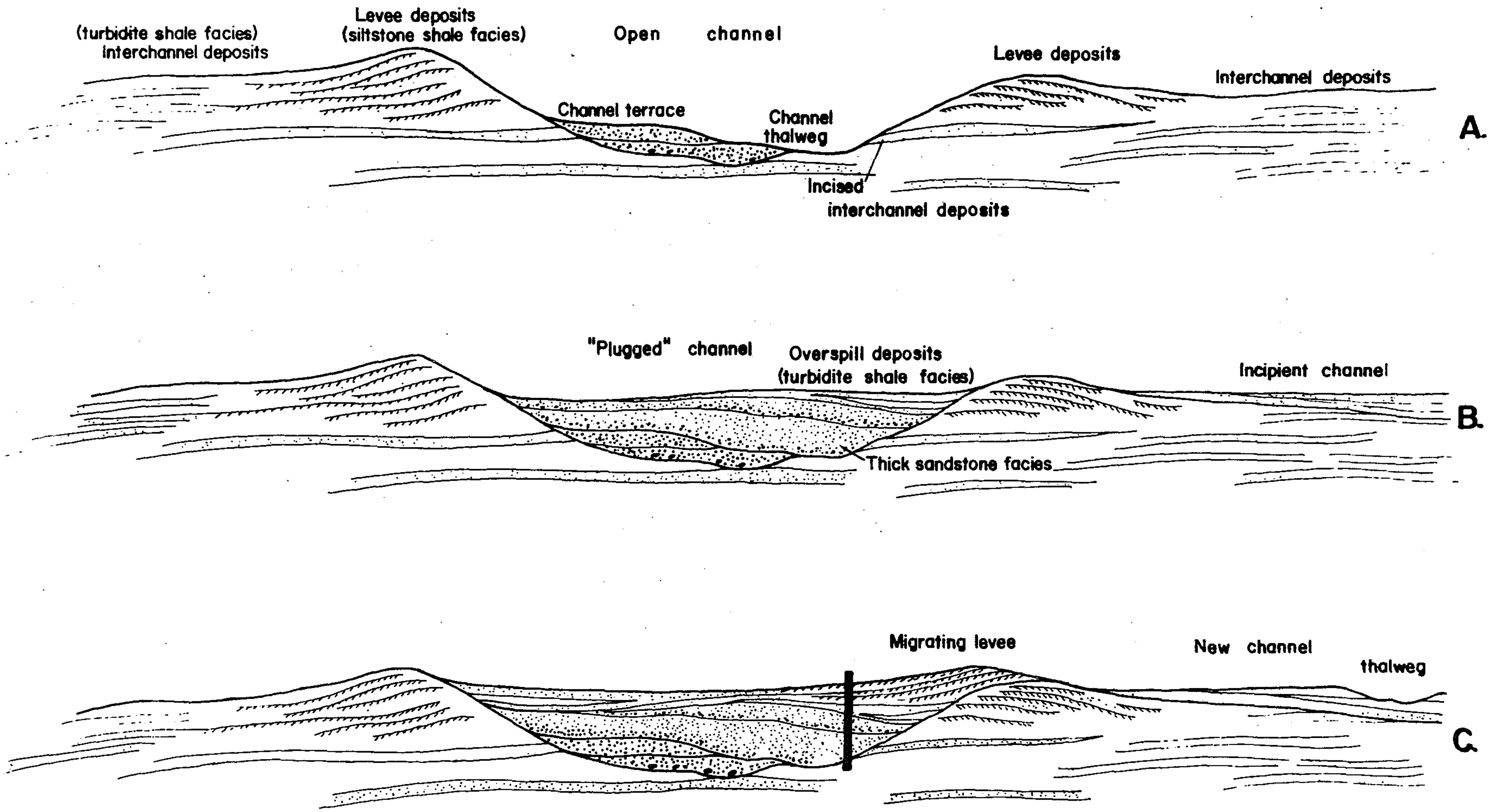


Figure 26. Conceptual model of midfan distributary channel avulsion processes. Thick black line in (C) indicates the position of a thinning-upward cycle.

A. Open channel. Sediment travels through a channel thalweg.

B. Channel plugged by thick flows that reduce the thalweg gradient. Upstream the channel has migrated (avulsion or levee breach) into the interchannel area. Sediment from the "new" channel migrates unconstrained by high banks and levees into the "old" channel.

C. The "new" channel has had enough time to incise itself into interchannel deposits and form levees. These levees may migrate over the original channel area.

sequence is probably the best approximation that can be made (Pickering, 1982). The distributary channels of the Windsor Township Formation, then, ranged from 5 to 15 m (16 to 50 ft) in depth (no allowance for compaction), which is comparable to the thickness of the outer fan depositional lobes described earlier.

The presence of thinning- and fining-upward megasequences characterized by packets of thick, amalgamated sandstone beds overlain by turbidite-shale and lesser siltstone-shale facies beds can be accounted for by short-lived activity of relatively shallow channels. In fact, this activity may be what ultimately controlled the complex internal organization of the outer-fan lobes. In any event, the thinning-upward trend of middle fan deposits is thought to result from the abandonment of a channel after plugging by a large flow or avulsion farther upslope (Walker, 1977, 1978; Hiscott, 1980; fig. 26). The thick, amalgamated beds at the bottom of the cycle represent the "plugs" whereas the thinner turbidite beds above represent spillover from a new channel adjacent to the old channel. Progressive thinning of beds above the plug bed may be due to the fact that the "new" channel has not had sufficient time to incise into the interchannel deposits so that flow is more or less unconstrained by levees or channel banks (fig. 26B). Subsequent flows continue to incise deeper into interchannel deposits forming banks that tend to reduce lateral flow of sediment. The presence of levee deposits overlying channel and interchannel deposits records lateral movement of levees formed in association with stabilization of the "new" channel (fig. 26C). The fact that the vast majority of Windsor Township channel deposits are incised into interchannel deposits and not levee sediment supports the scenario of channel evolution described above.

Locally, one encounters thickening-upward megasequences in close vertical association with the thinning- and fining-upward sequences (fig. 25). These sequences are generally less than 5 m (16 ft) thick, consist of an average of seven sand beds, and are significantly thinner than the outer-fan depositional lobe sequences. These deposits are closely associated with sequences of interchannel turbidite-shale facies beds that probably represent either crevasse channel-fill or crevasse lobe deposits and are significantly sandier (sandstone/shale=2-4/1) than those of the outer fan. There are, however, thicker (10-12 m (33 to 39 ft) negative megasequences associated with channel sequences. These indicate a very close association of channels and depositional lobes in the middle fan.

Sediment dispersal.

In general, paleocurrent indicators are not abundant in rocks of the Windsor Township Formation and it is difficult to evaluate paleoflow variations within a particular stratigraphic section. However, they are abundant locally and include groove, bounce, and prod casts, flute casts, cross- and ripple-laminations, and parting lineations. Prior to evaluation of paleocurrent indicators from a particular area, the vertical and, if possible, lateral facies associations must be understood. This was well demonstrated by Nilsen and Simoni (1973) who noted both transverse and longitudinal flow directions in the lower Tertiary Butano Sandstone of central California. Detailed turbidite facies analysis indicated to these authors that the diversity of paleoflow directions is a reflection of sequences of channelized sediment interlayered with sequences of levee and interchannel deposits that traveled outward and away from the channels. Examination of paleoflow trends within vertical sections of the Windsor Township Formation leads to two important observations: (1) Flow directions within the siltstone-shale facies and turbidite-sandstone facies beds diverge from current directions in bounding channelized thick-bedded sandstone facies by as much as 90 degrees (fig. 25). Variability is probably the result of lateral aggradation of channels. (2) Inferred outer fan depositional lobe deposits show variable flow directions.

The best evidence of flow divergence between channel fill and interchannel-levee deposits is gained from regional analysis of paleocurrent data. Graphic representation of paleoflow data without consideration of turbidite facies associations is at best difficult to interpret (fig. 27A). Interpretation of data, however, becomes much easier when a distinction of paleocurrents is made based upon

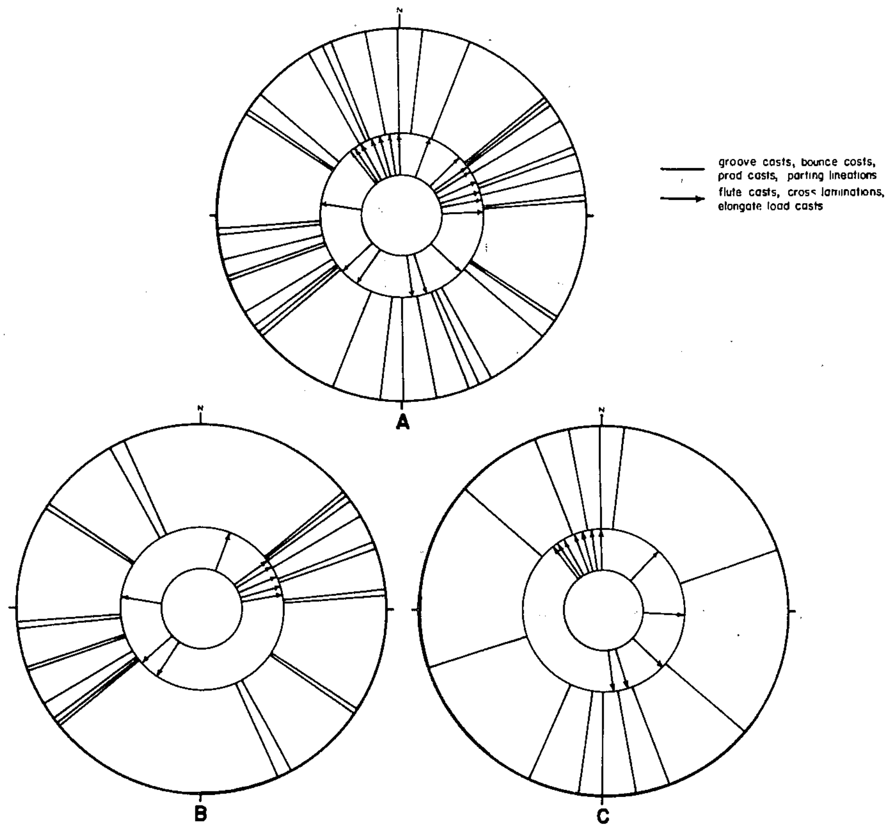


Figure 27. Paleocurrent directions in the Windsor Township Formation (n=40).
A. Undifferentiated paleocurrent directions.
B. Paleocurrent directions of channel deposits.
C. Paleocurrents of interchannel and levee (flute casts, cross laminations) deposits.

submarine fan facies interpretations (fig. 27B,C). In general, dominant channel flow directions are sub-parallel to the regional tectonic strike in both northeast and southwest directions (fig. 27B). Flow directions of interchannel and levee deposits, however, are to the northwest and southeast. In particular, levee deposits show a predominant flow direction to the north and northwest (fig. 27C). Overbank sedimentation and, in particular, levee building sedimentation is influenced by such variables as bottom and contour currents, the Coriolis effect (Menard, 1955), and subtle variations in depositional relief. The Coriolis effect, in particular, has caused levees associated with channels in the eastern Pacific Ocean to build higher on the right and therefore encourages deep-sea fan channels in this area to shift to the left with time. The predominance of northerly and northwesterly (Middle Ordovician directions; Scotese and others, 1979) flow directions in levee deposits of the Windsor Township is consistent with development of levees under the influence of the Coriolis effect in Middle Ordovician time. Another reason for north to northwesterly levee development is based upon a model to be discussed later that suggests that the Windsor Township Formation was deposited near a trench axis related to southeasterly-directed subduction.

In summation, paleocurrent analysis of the submarine fan deposits of the Greenwich slice suggests that major channels flowed to the southwest and northeast parallel to tectonic strike and that interchannel spillover, in particular levee building sedimentation, was directed predominantly to the present northwest indicating pronounced levee development on the northwest side of the channels. The presence of reversed flow directions of the channel deposits (i.e., flow to the northeast and southwest) can be explained

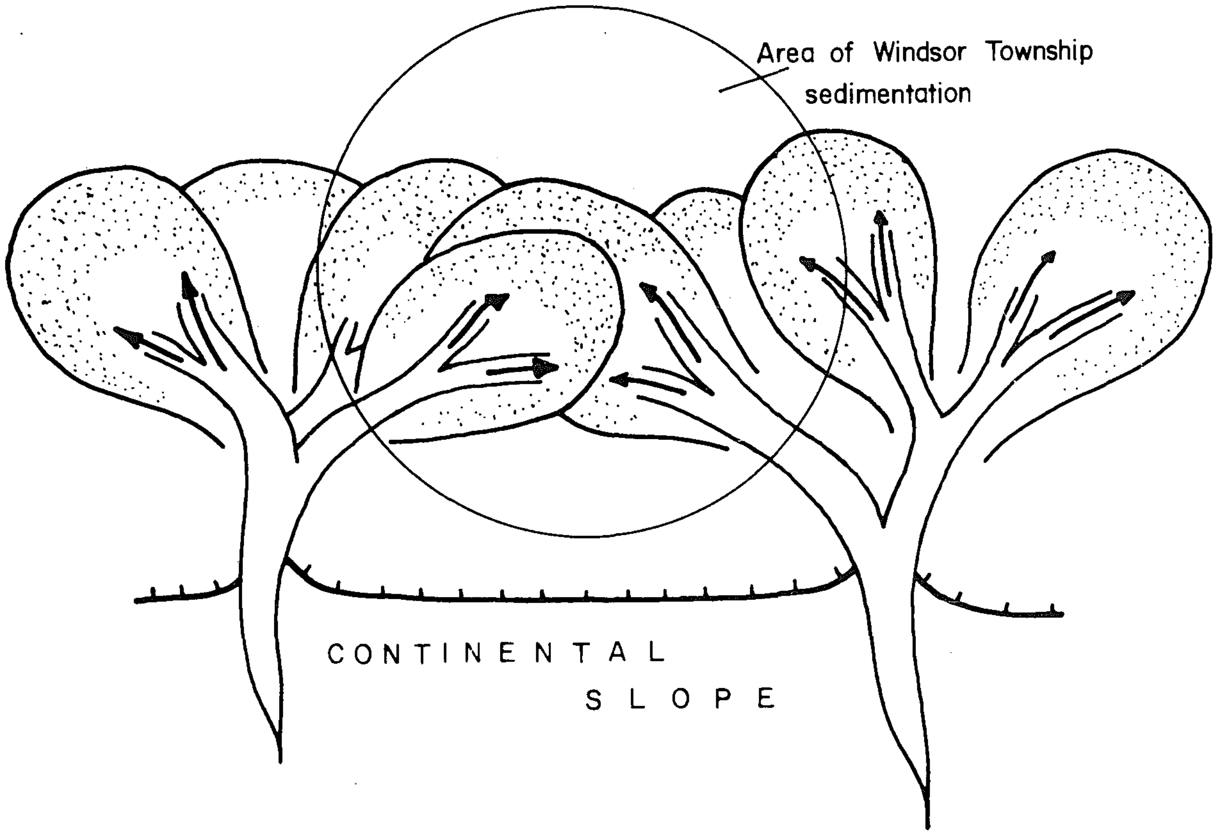


Figure 28. Coalescing fan model used to explain reversed paleocurrent directions and different sandstone source terranes (e.g., Jonestown sandstone and sandstone associated with the Werleys Conglomerate).

in two ways: (1) avulsion on an asymmetric fan (Hiscott, 1980) or (2) overlap of adjacent small fan systems (fig. 28). Petrographic data to be discussed later suggests that the second possibility is best suited to the Windsor Township Formation.

Submarine fan model for the Windsor Township Formation.

Paleoenvironmental interpretation of ancient submarine fan sequences and development of a submarine fan model is based on a combination of criteria that include: (1) analysis of vertical and lateral facies relations; (2) comparison of inferred mesotopography with both modern and ancient examples of submarine fans, and (3) synthesis of regional dispersal patterns. Ricci-Lucchi (1981), as noted earlier, argued that differences between the Italian (Mutti and Ricci-Lucchi) fan model and the suprafan model are not of terminology as suggested by Walker (1980) but are reflective of the grain-size distribution of sediment supplied to the fan. Two end member fan types reflective of this dependence upon grain-size distribution and referred to as "sand inefficient" and "sand efficient" have been proposed for modern and ancient fans (Mutti, 1979; Ricci-Lucchi, 1981). Examples of the first type are small, sand-rich fans characterized by overlapping and superposed channels and lobes and very reduced outer fan and fan fringe deposits. "Sand efficient" fans are pelite-rich, deep-sea fans in which a stable, well-leveed middle fan distributary network feeds channelless lobes.

The major sedimentary attributes of the Windsor Township Formation fan deposits are: (1) a relatively high sand supply-virtually all of the sections measured in the Windsor Township Formation have a sandstone: shale ratio greater than 2; (2) poorly developed fan fringe and channelless outer fan depositional lobe deposits; (3) superposed thinning- and thickening-upward megasequences; (4) superposed channelized, amalgamated sandstone packets separated by interchannel deposits; (5) general lack of well-defined submarine levee deposits. These characteristics are more typical of sand-rich ("inefficient") or suprafan systems. A high sand content of the sediment supplied to the Windsor Township fan(s) would tend to prevent levee development and is suggested by the lack of well defined levee deposits in measured sections. A situation such as this would allow for development of an unstable braided distributary channel system. Additionally, a poorly developed levee system explains the presence of crevasse channel-fill and crevasse lobe deposits both of which are relatively common in the Windsor Township Formation. The presence of depositional lobe megasequences associated with channelized deposits is typical of "sand inefficient" suprafan sedimentation. Local deposition of sand beyond unstable channel mouths would form a complex system of superimposed lobe and channel deposits such as that described from the Windsor Township Formation. The lack of well defined outer fan depositional lobe and fan fringe deposits adds further support to the suprafan model for the Windsor Township Formation.

Finally, the presence of reversed channel flow directions in the Windsor Township is consistent with turbidite sedimentation on a number of small coalescing fans. Normark and Hess (1980) have noted the presence of three morphologically distinct fans, the Navy, Delgada, and Monterey fans, each fed by separate submarine canyons, off the coast of California in close proximity to each other. The morphologic variations of these fans probably reflects subtle variations in grain-size distribution of sediment supplied to the fans which in turn can be related to variations in source area lithology, continental shelf transportation mechanisms, submarine canyon characteristics, shoreline morphology and current patterns. All of these factors are unrelated to sedimentation processes in the deep sea.

Sedimentary petrography

Suttner (1974) defined a sedimentary petrographic province as a compositionally distinct body of rock that forms natural units in terms of age, origin, distribution, and mineralogic composition. Mineralogic characteristics of sedimentary provinces reflect location, relief, and composition of the source area. In addition, sedimentary petrologists (Crook, 1974; Schwab, 1975) have demonstrated a close relationship between sandstone composition and tectonic setting of the source area and basin of deposition. Investigations into the framework compositions of sands from known tectonic settings have been used to provide a baseline from which to evaluate ancient sandstone (Breyer and Ehlman, 1981; Valloni and Maynard, 1981). These studies show that systematic variations in sand framework mineralogy are a function of provenance and tectonic setting. Dickinson and Suczek (1979) and Dickinson and Valloni (1980) have categorized sandstone suites in terms of sediment provenance and the various types of continental margins and ocean basins described by modern plate tectonic theory. These include: (1) opening basins like the Atlantic Ocean which are bordered by passive, rifted, or transformed orogens, (2) closing ocean basins like the Pacific Ocean which are bordered by active (arc) or orgenic continental margins, and (3) ocean basins adjacent to intraplate oceanic archipelago (Hawaiian Islands) (Table 2; Dickinson and Valloni, 1980).

Sandstone mineralogy has been shown to be a valuable tool in basin analysis and tectonic reconstruction (Graham and others, 1976; Dickinson and Suczek, 1979; Schwab, 1981; Dickinson and others, 1982; 1983) although it must be stressed that it does not argue categorically against settings other than those suggested by the detrital mineralogy of the sediment. Compositional studies must be used in a supportive role and evaluated along with all other pertinent geological information regarding a particular basin before inferences regarding conceptual plate models can be proposed.

Table 2. Mean Detrital Modes (in percent) of Modern Sands from Different Tectonic Settings (from Dickinson and Valloni, 1980)

Tectonic Setting	*Quartz (Q)*	*Feldspar (F)*	*Lithic Fragments (L)*
A. Rifted Continental Margin:			
1. cratonic block	78	18	6
2. craton plus rift belt	69	26	5
3. craton plus orogen	63	26	11
B. Orogenic Continental Margin			
1. transform arc orogen	31	45	24
2. continental-margin arc	20	41	39
C. Oceanic Island Chain:			
1. ocanic island arc	11	34	55
2. intraplate archipelago	0	5	95

Petrographic methods.

Seventy sandstone samples from the Windsor Township Formation covering the entire length of the Greenwich slice were examined. All samples were cut perpendicular to bedding and stained for both plagioclase and K-feldspar. Between 400 and 400 detrital grains were counted per section. Statistical analysis indicates that the determined detrital modes have a maximum uncertainty of 4 to 4.5 percent (Van Der Plas and Tobi, 1965). Grid spacing exceeded the average grain size so that in most cases individual grains were not counted more than once. Petrographic studies of sandstones can be carried out in two ways. In the first, any polymineralic grain is treated as a rock fragment. The second way, described by Dickinson (1970), involves counting all grains greater than an arbitrary limiting size of 0.0625 mm within lithic fragments as monocrystalline grains. This tends to reduce the influence of grain size on the composition of sand (Potter, 1978, fig. 3). Because lithic fragments are extremely important to interpretation of source area the first method was employed. Sample selection was limited to medium-grained sandstones (Graham and others, 1976) thereby limiting the influence of grain size on lithic fragment content but at the same time allowing for consideration of rock fragment lithology. Standard point counting methods described by Dickinson (1970), Graham and others (1976), and Ingersoll (1978), among others, were used in the present study. All parameters counted are explained in Table 3.

Matrix was defined as all grains less than 0.03 mm in diameter (Dickinson, 1970). Only samples with less than 20 percent matrix were considered, as a greater amount of matrix increases the probability that unstable lithic fragments were broken down to form matrix thereby shifting the detrital composition toward the quartz corner of standard QFL diagrams.

Detrital fragments include quartzose grains, plagioclase and K-feldspar, and sedimentary, igneous, and metamorphic lithic fragments. Sandstone samples from the Windsor Township Formation fall within subzone 1 of Dickinson (1970) and detrital textures are generally well preserved thereby allowing for determination of detrital modes with a high degree of confidence.

Quartzose grains include monocrystalline (Qm) and polycrystalline (Qp) quartz. Polycrystalline quartz grains are considered as lithic fragments and will be discussed below. Undulose monocrystalline quartz grains predominate over non-undulose grains. Individual grains are well rounded to angular.

Feldspar grains are unaltered to moderately altered and both twinned and non-twinned

Table 3. Definition of Grain Populations for Triangular Composition Diagrams (after Dickinson and others, 1982)

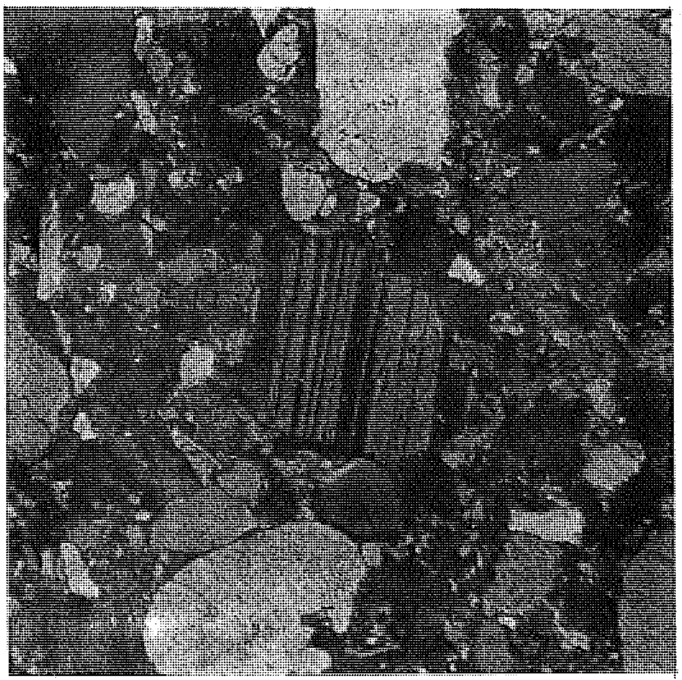
Triangular Diagram	*Uppermost Pole*	*Lower Left Pole*	*Lower Right Pole*
QFL	*Q Quartzose grains (=Qm + Qp)*	*F Feldspar grains (=P + K)*	*L Unstable aphanitic lithic (=Lv + Ls)*
QmFLt	*Qm Monocrystal-line quartz grains*	*F (same as above)*	*Lt Total aphanitic lithic fragments (=L + Qp)*
QmPK	*Qm (same as above)*	*P Plagioclase grains*	*K K-feldspar grains*

Table 4. Distribution (in percent) of A-twinned, C-twinned, and untwinned plagioclase feldspar in Windsor Township Formation sandstones samples. A minimum of 100 feldspar grains were counted.

Sample	*A-twins*	*C-twins*	*Untwinned Plagioclase*
1	*20*	*9*	*71*
2	*16*	*4*	*80*
3	*18*	*7*	*75*
4	*14*	*7*	*79*
5	*15*	*6*	*79*
6	*17*	*3*	*80*
7	*10*	*5*	*85*
8	*8*	*2*	*90*
9	*7*	*2*	*91*

varieties are present (fig. 29B). Plagioclase is generally equivalent to, or more abundant than, K-feldspar except for sandstones collected from the area around Jonestown (fig. 31C). Twinning is common to uncommon within a particular sample and zoned plagioclase grains could not be detected. Flat stage analysis of twinned plagioclase grains indicats that A-twins (Gorai, 1951) are much more abundant than C-twins. Table 4 shows the distribution of A- and C-twinned and untwinned plagioclase grains in 10 Windsor Township samples. The observations that more than 70 percent of the plagioclase grains are untwinned, and A-twins are two to five times as abundant as C-twins suggests derivation of the plagioclase from a metamorphic-plutonic source (Pittman, 1970).

A.



B.

Figure 29. Photomicrographs of detrital grains from Windsor Township sandstones (all photos at 56x): A) monocrystalline quartz (note bubble inclusion trains); B) plagioclase feldspar.

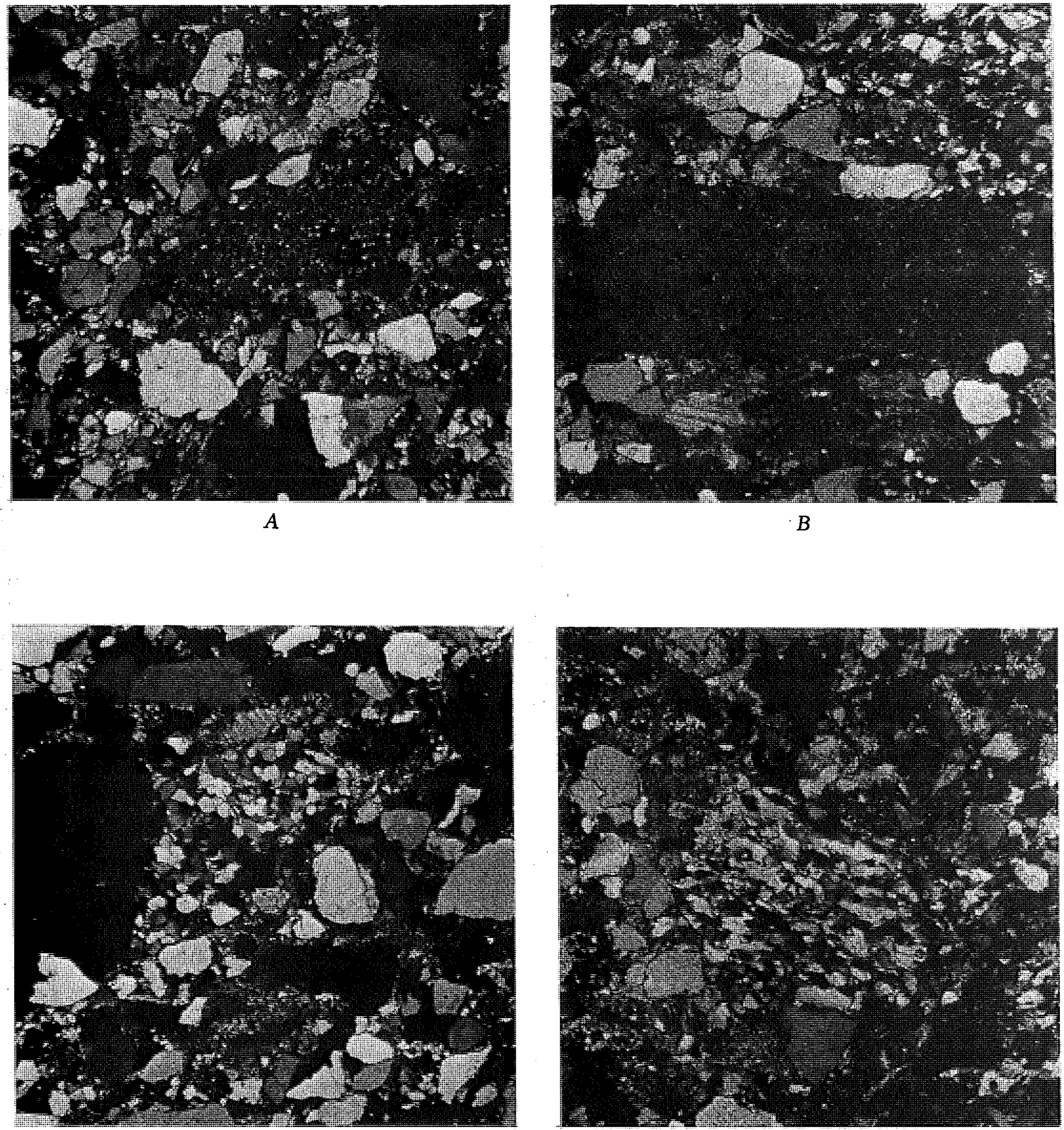
K-feldspar can be divided into grains of microcline with cross-hatch twinning and untwinned grains which are generally orthoclase as determined by the size of the optic angle. Orthoclase, locally perthitic, is the common K-feldspar. Both plagioclase and K-feldspar grains are smaller and more angular than quartz grains.

Lithic fragments constitute a highly variable proportion of the framework of the Windsor Township sandstones. Lithic grains are either fragments of older rocks or re-worked silt- and sand-size particles of penecontemporaneously deposited mud and siltstone. Lithic fragments of the Windsor Township Formation can be classified into eight major categories briefly described below in order of decreasing abundance:

1. Argillite-shale: murky, fine-grained argillaceous fragments with occasional enclosed detrital silt grains (fig. 30A);
2. Chert: microcrystalline but monomineralic silica aggregates (fig. 30B);
3. Aggregate quartz: intergrown crystalline quartz without significant planar fabric (fig. 30C, I);
4. Foliate metaquartzite: polycrystalline quartz with a planar fabric (fig. 30D);
5. Polycrystalline mica: metamorphic mica with prominant planar fabric (fig. 30E);
6. Plutonic igneous: phaneritic, intergrown quartz and K-feldspar in varying proportions (fig. 30F);
7. Quartz-mica tectonite: metamorphic mica and quartz in varying proportions and commonly, but not necessarily, with planar fabric (fig. 30G);
8. Volcanic: primarily fine-grained felsitic and microlitic fragments with poorly- to well-developed lathwork textures (figs. 30H);

   Lithic fragments not listed above include micritic, recrystallized, and peloidal limestone grains.

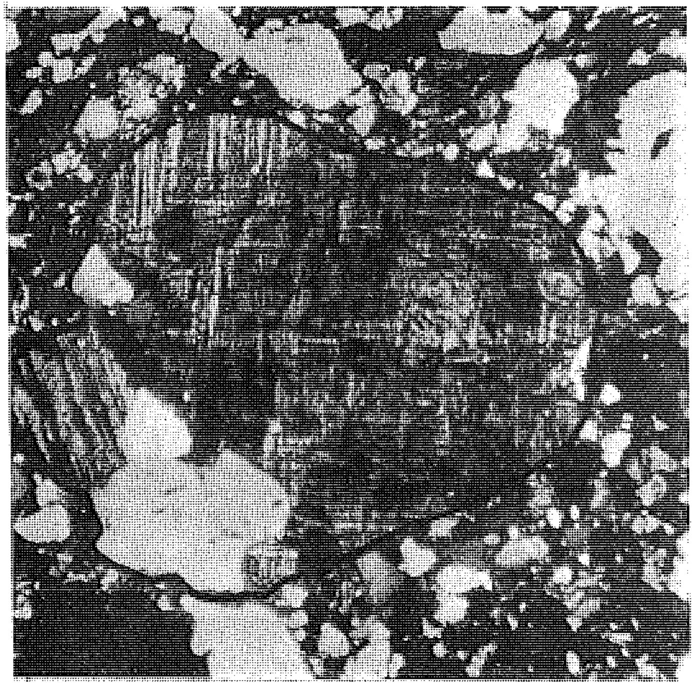
The most common polycrystalline quartz grains are chert and aggregate quartz. These types vary in abundance from sample to sample. Chert fragments are similar to some of the chert associated with the pelagic shale of the Greenwich slice. Aggregate quartz is differentiated from foliate metaquartzite by its



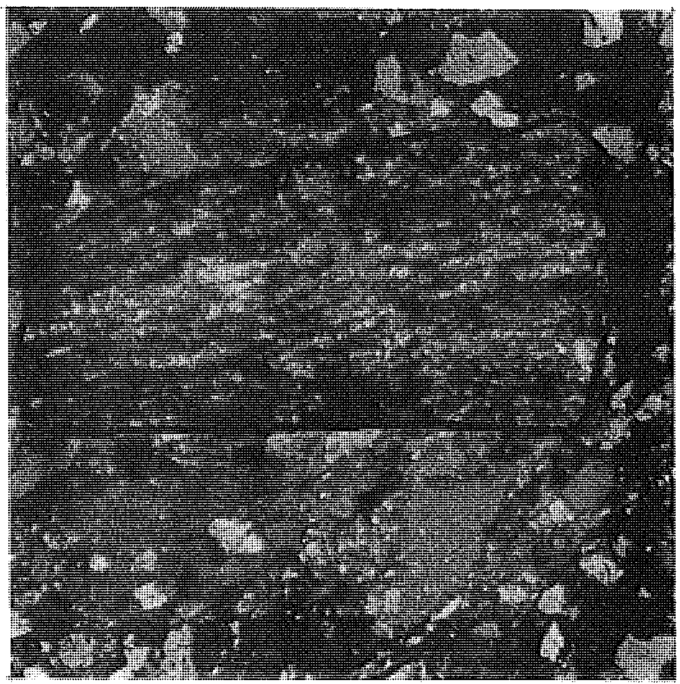
A B C D

Figure 30. Photomicrographs of lithic fragments from Windsor Township sandstones (all photos at 56x): A) argillite-shale; B) chert (Qp); C) aggregate quartz (Qp); D) foliate metaquartzite (Qp).

E



F



G

H

Figure 30 (continued). E) polycrystalline mica (outlined in black); F) plutonic igneous (note interpenetrating grain boundaries outlined in black); G) quartz-mica tectonite (outlined in black); H) volcanic lithic (outlined in black, note lathwork texture near the center of the grain).

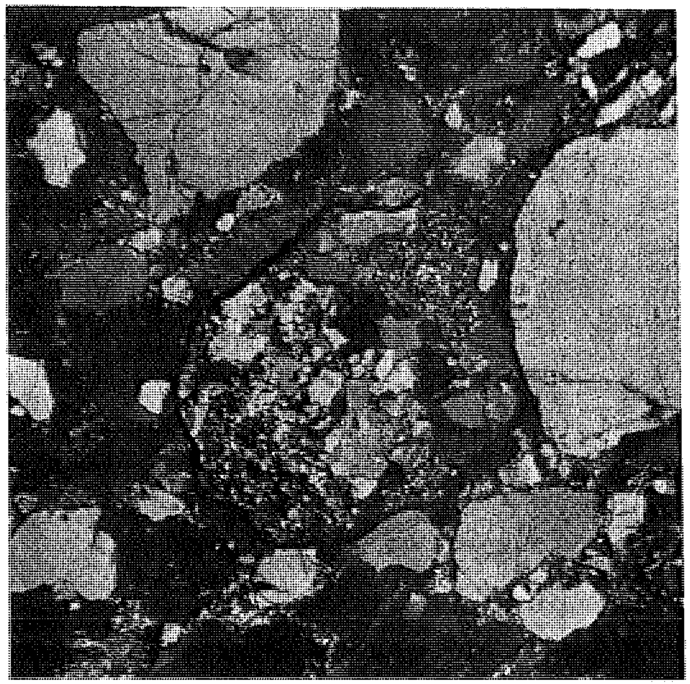


Figure 30 (continued). I) aggregate quartz (metachert)-chert fragment: the aggregate quartz (metachert) probably formed from recrystallization of chert.

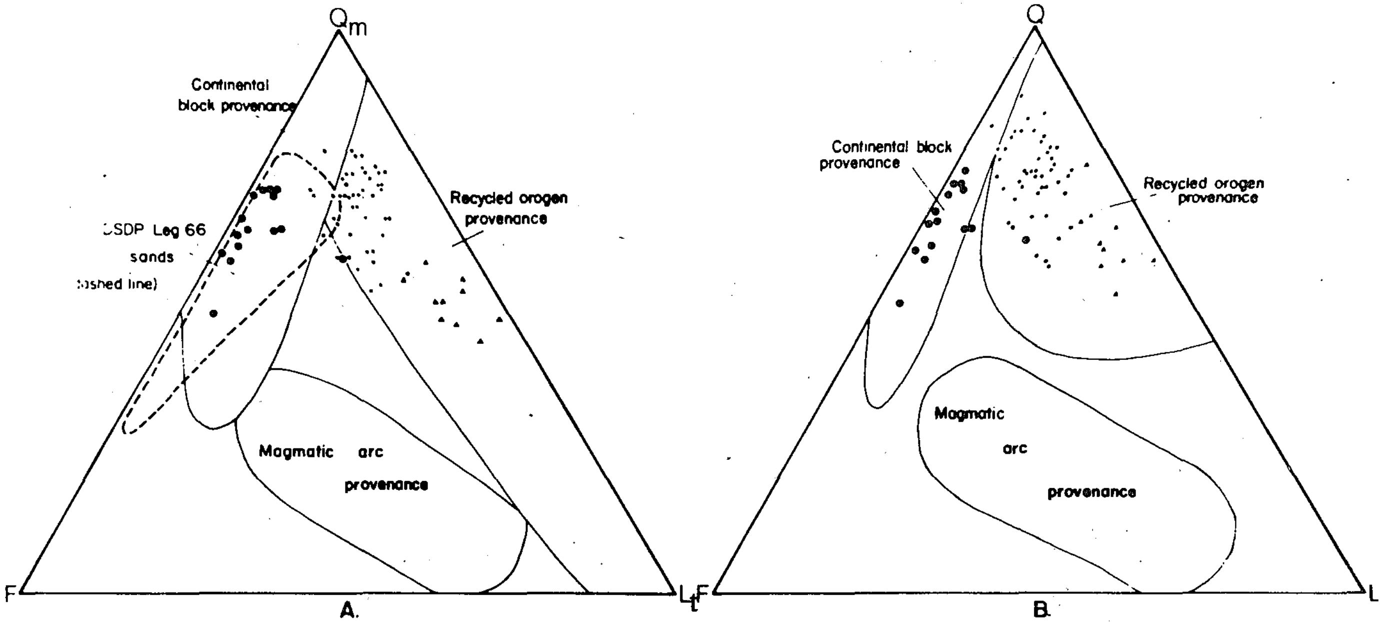
lack of tectonite fabric. Microgranular aggregates of quartz grains characterized by sutured grain contacts are probably metachert grains (Dickinson and others, 1979). Quartz-mica tectonite and foliate metaquartzite fragments form a continuum of related metamorphic types that are primarily metasedimentary. Igneous rocks are of varying occurrence. Plutonic igneous fragments are locally abundant (samples from the Johnstown area) and appear to be granite and syenite clasts. Volcanic rock fragments are rare and consist predominantly of microlitic and minor felsitic grains. Microlitic grains are characterized by laths or microlites of plagioclase crystals arranged in either disoriented or fluidal fabrics. Felsite grains display a fine-grained mosaic of quartz and feldspar that are difficult to distinguish from chert. The dominant criteria for distinction of felsite and chert grains in plane-polarized light is the difference in internal relief between quartz and feldspar and the presence of microphenocrysts. Best success, however, is obtained by studying stained thin sections. Felsite grains display yellow, pink, or mixed orange colors where chert remains unstained.

Discussion of results.

Figure 31 shows the results of the point counting of key framework components of Windsor Township sandstones. The plots indicate three dominant petrographic associations: (1) sandstones collected from near Jonestown, (2) sandstones associated with the conglomerate near Werleys Corner collected from near the eastern end of the klippe in Lehigh County, and (3) the remainder of the sandstones collected over the entire length of the Greenwich slice. The Jonestown rocks, by virtue of their high feldspar content, low P/F ratio, and low lithic fragment content (fig. 31) fall within the continental block provinance of QFL (quartz, feldspar, and lithic fragments) and QmFLt (monocrystalline quartz, feldspar, and triangular diagrams (fig. 31A,B). Sandstones associated with the conglomerate near Werleys Corner and the remaining Windsor Township sandstones fall within the recycled orogen provenance of these plots. The Werleys Corner sandstone is differentiated from the other sandstones of the Windsor Township Formation by presence of abundant lithic fragments, in particular limestone and chert fragments and sparse but conspicuous volcanic clasts.

Sandstones collected from near Jonestown are characterized by monocrystalline quartz, feldspar, and sparse lithic fragments. Rare polycrystalline quartz fragments are typically foliate metaquartzite fragments. Feldspar consists of large K-feldspar grains and generally subordinate plagioclase (fig. 31C). Lithic fragments include granitic and syenitic plutonic igneous fragments. The absence of shale clasts is interesting as these form a major component of the lithic fragment population of the other Windsor Township sandstone suites. The detrital mineralogy of the sandstones collected from near Jonestown suggest derivation from an eroded plutonic-low grade metamorphic complex characterized by granitic and syenitic igneous rocks.

The vast majority of sandstones of the Windsor Township Formation are characterized by somewhat less feldspar, higher plagioclase/total feldspar ratio, and significantly more lithic fragments than those collected from the Jonestown area (fig. 31). Rounded monocrystalline quartz grains are the



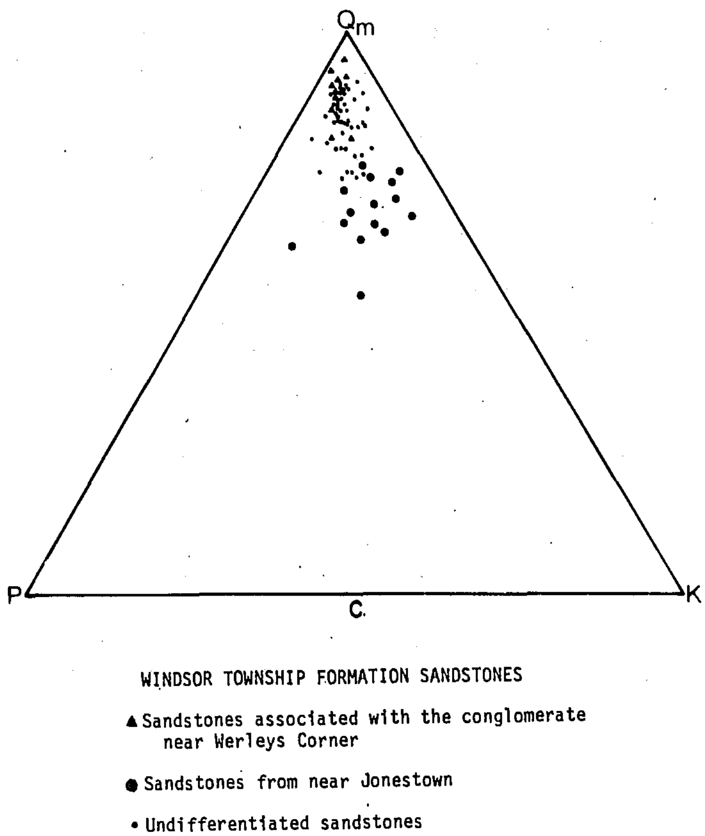


Figure 31. QmFLt, QFL, and QmPK plots of framework modes for the Windsor Township Formation.

dominant detrital component of these sandstones. Plagioclase (oligoclase) is generally more abundant than K-feldspar (fig. 31C). Orthoclase and lesser microcline constitute the K-feldspar population. Polycrystalline quartz, an important constituent of these rocks, includes variable proportions of chert, aggregate quartz, and foliate metaquartzite grains. In some cases chert fragments are cut by chalcedony veins indicating that they were deformed prior to incorporation into the sandstone. Aggregate quartz grains are probably metachert. Argillite-shale fragments probably derived from penecontemporaneous mud are the most common lithic fragments. Volcanic rock fragments are essentially non-existent in all sandstones studied except for those associated with the Werleys Corner conglomerate. Sparse volcanic fragments in these rocks are typically microlitic.

Source areas for the recycled orogen provenance include (1) subduction complexes of deformed oceanic sediments and lavas, (2) collision orogens formed along suture zones between once-separate continental blocks, and (3) foreland fold-thrust belts. The presence of abundant chert and minor but conspicuous volcanic fragments in these rocks, in particular those associated with the conglomerate near Werleys Corner, is suggestive of derivation from a subduction complex. However, the minor but conspicuous plutonic and metamorphic fragments (i.e., Jonestown) and siltstone, shale, and limestone fragments are suggestive of derivation from a collision orogen provenance (Dickinson and Suczek, 1979; Ingersoll and Suczek, 1979)

The source terrane for the Windsor Township sandstones is complex and is better considered in the context of convergent margin tectonics. Moore (1975) and Moore, Watkins, and Shipley (1981), among others, maintain that density contrasts between oceanic or pelagic sediment and overlying trench and near-trench sediment control the type and amount of sediment offscraped and accreted at the toe of the inner trench slope. In general, the greater the amount of buoyant near-trench and trench axis sediment at a particular convergent margin, the more enriched the resulting subduction complex will be in those types of deposits relative to pelagic deposits. This is typical of "clastic dominated" convergent margins (Shepard and others 1981). Furthermore, subducted pelagic and volcanic material may become underplated at depth only to move to the surface behind the accreted trench sediment (Moore, Watkins, and Shipley, 1981). A possible example of these relations can be seen in the Kodiak Islands of Alaska. Here the Uyak melange, a lithologically heterogeneous, structurally complex and metamorphosed (prehnite-pumpellyite facies, Connelly, 1978) unit composed of oceanic plate sedimentary rocks lies landward of a less-deformed broken formation composed predominantly of turbidite deposits, the Kodiak Formation, that was deposited in a trench and/or slope environment (Nilsen and Moore, 1979). Both of these units were emplaced in Late Cretaceous time (Moore, 1978).

Dickinson and Suczek (1979) maintained that the key signal of subduction complex-derived sandstone is the presence of chert grains in excess of monocrystalline quartz. However, the recognition of underplating as a viable process provides an explanation of the complex framework composition of subduction complex sandstones. For example, Dickinson and Suczek (1979) noted that most of their data were from subduction complexes composed mainly of chert, argillite, and greenstone. These sandstones may have been derived from exposed underplated rocks. On the other hand, Dickinson and Suczek maintained that erosion of subduction complexes composed primarily of turbidite and hemipelagic deposits, accreted trench and near-trench deposits, would yield detritus characterized by rounded monocrystalline quartz grains, shale clasts, and lesser proportions of hemipelagic argillite, chert, metachert, recrystallized limestone, and volcanic fragments.

Petrographic characteristics and field relations to be discussed later are consistent with derivation of the Windsor Township sandstones from a collision orogen and/or subduction complex and deposition of this sediment on oceanic crust in an active subduction trench. Sediment derived from collision orogens is generally transported longitudinally from a suture zone and shed into a closing remnant ocean basin as turbidite deposits (Dickinson and Suczek, 1979; Graham and others, 1976). An excellent modern analogue of this dispersal pattern is the Himalayan suture-Bengal Fan dispersal system. In this area sediment is shed from the Himalayan suture, transported parallel to the suture line to the apex of the Bengal Fan and deposited in the Bengal Fan and, to a lesser extent, the Java Trench. This model was originally suggested for dispersal of the Windsor Township Formation by Lash (1982a) on the basis of tectonic strike-parallel channel flow directions in these rocks. However, the high sandstone/shale ratio and virtual lack of submarine fan facies other than those reflecting the sandy middle fan tend to rule out sediment dispersal by long-distance axial flow along an inferred trench axis. If long-distance axial transport of sediment was the cause, proximal-distal relations for coeval strata, should be apparent on a regional scale (e.g., Cugach Terrane, Alaska, Nilsen and Zuffa, 1982). This is not the case for the Windsor Township. In addition, the reverse paleocurrent directions of the Windsor Township Formation are not consistent with long-distance axial transport but rather with sedimentation on a number of small coalescing fans that entered the trench at various points along its length via submarine canyons.

It is likely that submarine canyons fed by rivers that crossed or cut into different source terranes resulted in building of fans with different petrographic characteristics. The compositionally and texturally immature nature and relatively coarse-grained size of the sand suggest that the canyons funneled sediment from fluvial and littoral settings (McMillen and others, 1982) directly to the ocean floor effectively bypassing shelf and forearc basins (Underwood and others, 1980). The majority of sediment was probably derived from erosion of a subduction complex that consisted of offscraped turbidite and hemipelagic deposits and lesser proportions of pelagic deposits. The composition of the Jonestown sandstones, however, indicates erosion of a plutonic and metamorphic complex. In addition, the conspicuous lack of argillite-shale clasts in the Jonestown sandstones suggest that the sand traveled through clean, mud-free canyons probably cut into crystalline rocks. It is possible that the strike-parallel variations in sandstone composition reflect variations in source area of the sediments possibly related to strike-slip or transform faulting exposing crystalline basement and underplated oceanic sediment (Karig, 1980; Bachman and Leggett, 1981). A modern analogue of this process can be seen in that part of the Middle America Trench studied by DSDP Leg 66 (Moore and others, 1981). This part of the Central America Trench off the southern coast of Mexico is characterized by a voluminous sand supply, narrow shelf (i.e., no forearc basin) and a steep slope cut by submarine canyons. The largest submarine canyon, the Ometepec Canyon, has a drainage basin composed of Precambrian and Paleozoic metamorphic rocks and Cenozoic plutonic rocks. Sand recovered from the mouth of Ometepec Canyon is compositionally identical to the Jonestown sandstones (Bachman and Leggett, 1981; Enkeboll, 1981; fig. 31). Significantly, the amount of accreted sediment at this part of the Central America Trench is much less than that expected for about 100 m.y. of subduction (Moore, Watkins, and others, 1982). Bachman and Leggett (1981) suggested that much of this material was "eroded" by transformed faults that subsequently exposed the Trans-American plutonic complex to erosion resulting in deposition of quartzofeldspathic sediment in the trench.

In summation, petrographic studies of Windsor Township sandstones rule out dispersal models involving long-distance axial transport of sediment and supports, instead, sedimentation on a number of small fans along the strike of the basin. This substantiates the conclusions of the sedimentologic study of the Windsor Township flysch described earlier.

The majority of samples excluding those collected near Jonestown were derived from a subduction complex that included turbidite, hemipelagic, and pelagic sediment. Minor volcanic and metamorphosed pelagic sediment may have been derived from exposed underplated deposits. The sandstones collected from near Jonestown, however, were derived from a plutonic-metamorphic terrane. The presence of these deposits and possible underplated detritus (e.g., metachert and volcanic fragments) in other sandstones argues for the presence of transform faulting that "eroded" accreted sediment thereby exposing crystalline basement and underplated rocks. For example, the presence of chert, metachert, and sparse volcanic fragments in sandstone associated with the Werleys Corner conglomerate are suggestive of derivation from a drainage basin that included minor accreted and/or underplated pelagic and volcanic rocks.

Sedimentological evolution of the Greenwich slice

The lithologic, sedimentologic, and petrographic characteristics of the Greenwich slice help to constrain models involving its tectonic evolution. Recently, Lash (1980) and Lash and Drake (1984) suggested that the Greenwich slice is part of an accretionary complex related to early Paleozoic subduction in the central Appalachian orogen. This proposal is based upon consideration of depositional patterns of the Greenwich slice deposits discussed in this chapter and the meso- and microscopic fabric of these rocks to be discussed later. Recent DSDP investigations of sedimentation patterns at active convergent margins provide an excellent actualistic baseline with which to compare sedimentation patterns of the Greenwich slice. The purpose of this section is to (1) consider the significance of vertical relations of the turbidite and red shale units, (2) distinguish between sediment deposited in trench, near-trench, and ocean floor environments, (3) assess the influence of crustal convergence on sedimentation patterns of the Greenwich slice, and (4) propose a depositional model for the Greenwich slice.

Schweller and Kulm (1978), von Huene (1974), Piper (1972), Piper and others (1973), and Moore and others (1982) discussed depositional patterns at active convergent margins and Underwood and others (1980), Leggett and others (1979), and Bachman (1978) outlined sedimentation processes and facies associations in exposed convergent margin sequences. Most of these studies, particularly Schweller and Kulm (1978), have defined four major facies found in association with convergent margins: (1) fan shaped bodies of turbidite deposits, (2) horizontally layered turbidite deposits with a large axial channel(s) (trench wedge of Schweller and Kulm), (3) hemipelagic deposits (terrigenous plate of Schweller and Kulm), and (4) pelagic sediments (pelagic plate of Schweller and Kulm). These deposits generally form sequences that consist of pelagic clay or biogenic ooze covered by hemipelagic sediment and turbidite deposits (fig. 32). The transition from pelagic to hemipelagic sedimentation, i.e., dilution of the pelagic sediment by hemipelagic components (Schweller and Kulm, 1978), occurs as a site on the oceanic plate approaches and eventually reaches the trench axis, illustrating the concept of "plate stratigraphy" (Berger, 1974). The pelagic plate is composed of slowly accumulated (2-5 mm/1,000 yrs) pelagic ooze and clay deposits (Schweller and Kulm, 1978). Terrigenous hemipelagic silt- and clay-size sediment deposited from a turbid water column and/or from fine-grained turbidity currents becomes dominant as the oceanic lithosphere approaches the continental margin. Accumulation rates of hemipelagic sediment range from pelagic values to 175 m/m.y. (e.g., Aleutian Abyssal Plain, Kulm, von Huene, and others, 1973).

Seismic reflection profiles across trench wedges have occasionally illustrated the presence of a large channel or channels that disrupt the otherwise horizontal reflectors (e.g., Aleutian Trench, von Huene, 1974). Deep sea fans are found along a number of active convergent margins (e.g., Middle America Trench, Moore, Watkins and others, 1982) and consist of the same types of deposits as the trench wedge. The major difference between

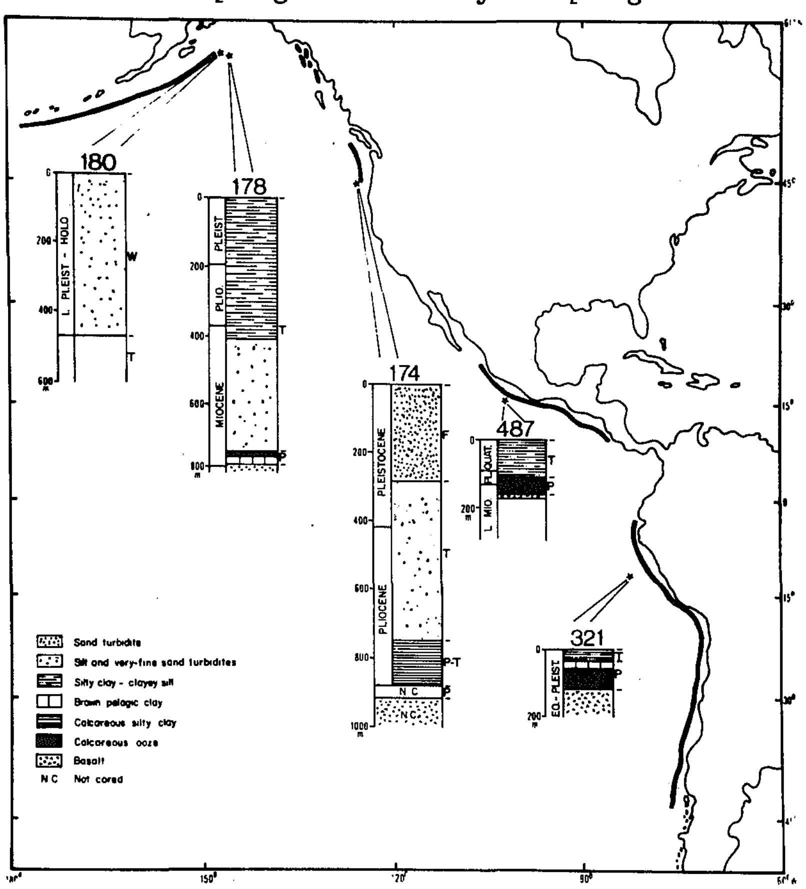


Figure 32. DSDP sites in eastern Pacific convergence zones with lithologic columns and facies. P=pelagic plate; T=terrigeneous plate; W=trench wedge; F=submarine fan. Heavy lines are trench axes. From Schweller and Kulm (1978). Site 487 from Moore and others (1982).

trench wedge and deep sea fans is the lack of a definable submarine fan morphology similar to that described by Normark (1970, 1978) associated with the trench wedge. Cone-shaped deep sea fans build outward from the base of the continental slope directly upon pelagic plate, terrigeneous plate, or trench wedge deposits. Moore and Karig (1976) maintained that recognition of coarsening-upward sequences provides the best single indicator of accreted trench and ocean deposits. Indeed, sequences similar to this have been described from exposed subduction complexes (Budnik, 1974; Leggett and others, 1982) as well as from cores recovered from active convergent margins such as the Aleutian Trench (Kulm, von Huene and others, 1973; von Huene, 1974), the Oregon/Washington Trench (Silver, 1969), the Peru-Chile Trench (Yeats, Hart, and others, 1976; Moore, Watkins and others, 1982) (fig. 32).

A number of traverses through the Greenwich slice reveal the existence of a definite vertical stratigraphic relationship between the pelagic deposits and hemipelagic-turbidite sequences. A typical sequence begins with 40 to 60 m (131 to 197 ft) of pelagic sediment that was probably deposited with no major interruptions from Early to Middle Ordovician time. The next lithology continuing up section is commonly olive-green hemipelagic clay and mudstone and interbedded thin silt turbidite (contourite?) beds. These rocks grade into thick sequences of hemipelagic mudstone and turbidite deposits that may be in excess of 1,000 m (3280 ft) thick although a complete section has never been measured. N. gracilis zone graptolite species have been collected from these rocks. The duration of sedimentation of the sequence described above is roughly 25 m.y.. Pelagic sedimentation spanned a time of approximately 20 m.y. whereas the hemipelagic-turbidite sedimentation spanned a period of approximately 5 m.y. Estimation of pelagic sedimentation rates as deduced from measured sections yields values that range from 1-9 m/m.y., well within accepted pelagic sedimentation rates. It is difficult to determine sedimentation rates of the overlying hemipelagic and turbidite deposits. The presence of load structures on the soles of many sand beds is suggestive of rapid rates of sedimentation. Indeed, the transition from pelagic to hemipelagic and turbidite sedimentation in Middle Ordovician time tells of a profound change in depositional environment and a great increase in sedimentation rate. Sedimentation rates of the turbidite deposits were probably well in excess of 100 m/m.y. and probably closer to 300 m/m.y. which is the average rate of Pleistocene trench-axis sedimentation in the Middle America Trench off Guatemala (Coulbourn and others, 1982). The coarsening-upward sequences of the Greenwich slice record movement of the oceanic plate toward the trench axis in the same way as those of the eastern Pacific margins record plate convergence (fig. 33). The generally older (Early to Middle Ordovician) pelagic sediment gave way to deposition of outer-trench slope hemipelagic mud as the oceanic plate came under the influence of the trench axis plume (Moore, Watkins, and others, 1982). The plate continued to move toward the trench axis resulting in sedimentation of coarse-grained turbidites (Middle Ordovician) on the trench floor. The complex nature of the coarsening-upward sequences of the Greenwich slice (fig. 33) attests to local complexities within and near the trench. For example, uplift of trench turbidites (Prince and others, 1974) and/or ponding of sediment related to active normal faults in the trench axis (Coulbourn, 1981).

The presence of the boulder and pebble conglomerate units in the Greenwich slice are particularly important to the sedimentological evolution of the slice. Clast lithologies of these units indicate that the Greenwich slice was the source of the clasts. Recent geophysical studies of active convergent margins have resulted in the recognition of large gravity driven slump deposits on the inner trench slope and in the trench (Jacobi, in press and refs. therein). Recently, Moore and others (1981) described large slide deposits from seismic reflection profiles across the lower inner trench slope of the Middle America Trench off southern Mexico (fig. 34). These authors have suggested that the slide deposits were derived from collapse of accreted oceanic and trench sediment and inner trench slope deposits. The clast lithologies of these deposits and their association with the coarsening-upward sequences are consistent with slumping or collapse of steep scarps formed and eventually oversteepened by thrust faults or folding of

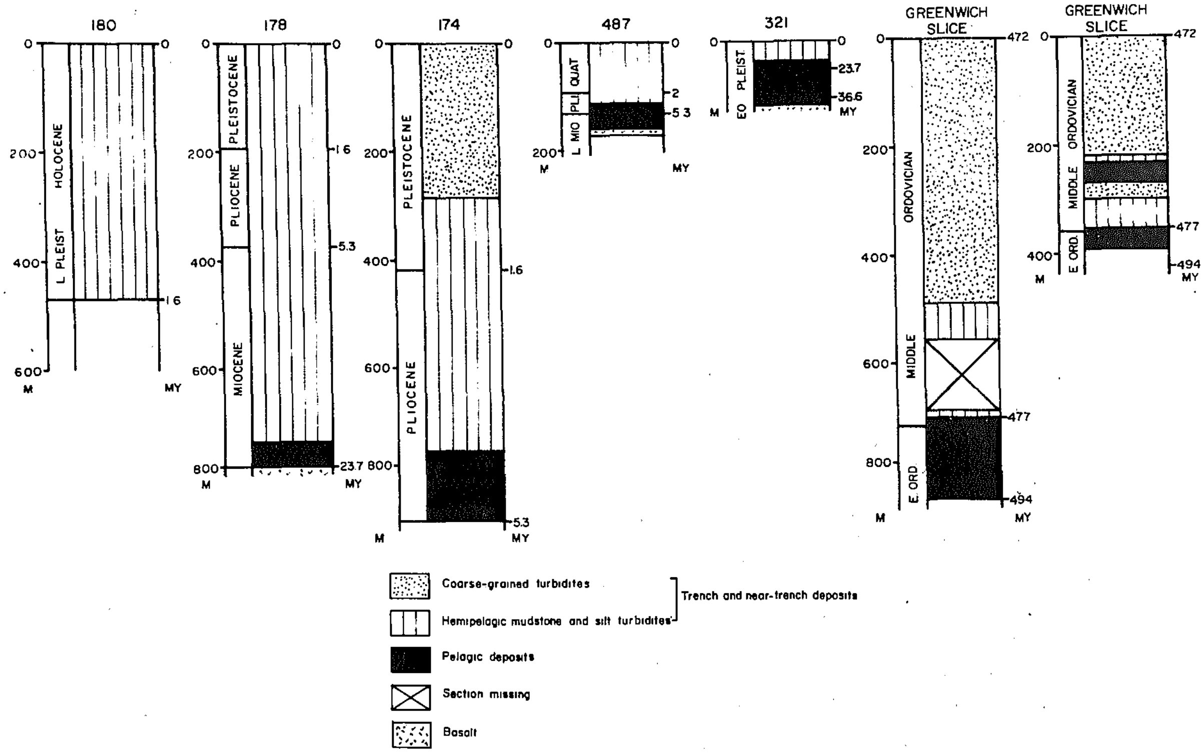


Figure 33. Comparison of coarsening-upward sequences of the Greenwich slice and those recovered from DSDP sites in eastern Pacific convergent margins. Eastern Pacific sections from Schweller and Kulm (1978) and Moore and others (1982).

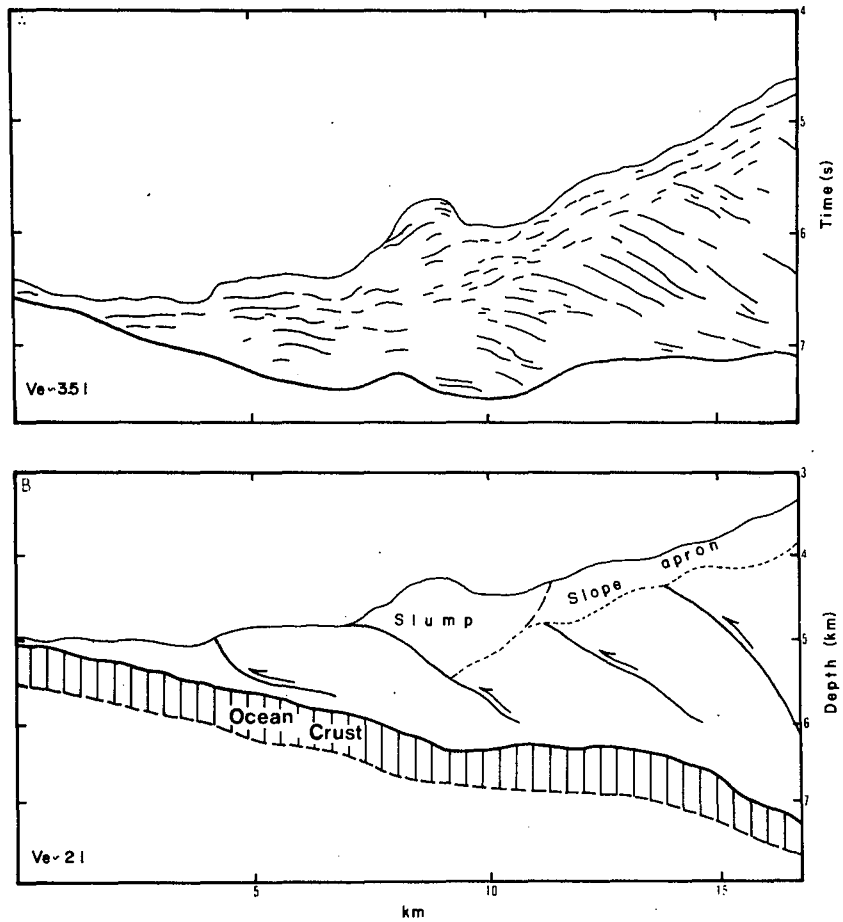


Figure 34. A) Line drawing of migrated time section MX-15 from the Middle America Trench off southern Mexico. B) Interpretation and depth conversion of line drawing emphasizing potential major faults. From Moore, Watkins, and Shipley (1981).

offscraped sediment in the inner trench slope. Continued deformation would eventually result in the incorporation of these deposits into the accretionary complex (fig. 35; Shipley and others, 1982).

Controls of sedimentation at convergent margins.

Depositional patterns at active subduction trenches are dependent upon (1) plate convergence rate perpendicular to the trench axis, and (2) sediment input to the trench axis. The latter is generally a function of variables such as shelf width, climate, drainage patterns, etc., all of which are generally not related to convergence rate. Sedimentation rate is perhaps the most important (Leggett, 1980b) of the two variables mentioned above. Convergence rate, however, is important in that it determines the residence time or the amount of time that trench sediment will remain in the trench until it is offscraped and deformed. To illustrate the dependence of depositional patterns on these factors Schweller and Kulm (1978) plotted convergence versus sedimentation rate for eastern Pacific trenches (fig. 36). Examination of this diagram indicates that high sedimentation and low convergence rates will result in large fans which could conceivably swamp the trench and extend onto the abyssal plain (e.g., Washington/Oregon margin). On the other hand, low sedimentation and high convergence rates will result in little or no accumulation in the trench (e.g., Marianas Trench, Northern Chile). The intermediate conditions, discontinuous and continuous trench wedges and small fans developed over a continuous trench wedge reflect increases in sedimentation rate and decreases in convergence rate. An important point to be made here is that if convergence rate is kept constant, sedimentation becomes the dominant factor. This is well demonstrated along the Barbados forearc region (Westbrook, 1982) as well as the Sunda Trench (Moore and others, 1980). The sedimentation rate of the Greenwich slice trench deposits, the Windsor Township Formation, is difficult to deduce but a reasonable value might be 250 to 300 m/m.y. Phillips and others (1976) suggested that the

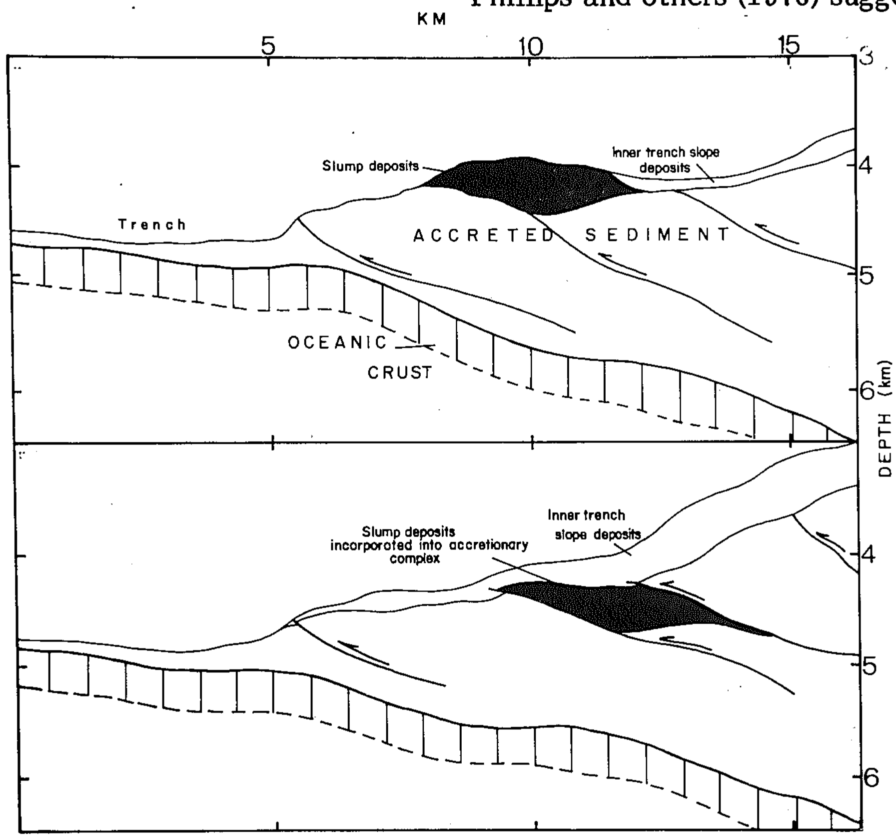


Figure 35. Model illustrating incorporation of inner trench slope slide deposits into an accretionary complex.

convergence rate of the proto-Atlantic Ocean in early Paleozoic time was less than 2 cm/yr, a low rate of convergence. This low rate of convergence and proposed moderate rate of sedimentation of the Windsor Township turbidite and hemipelagic sediment places these deposits within the field of small fans over a trench wedge on the Schweller and Kulm diagram (fig. 36).

Knowledge of the duration of sedimentation and limiting rates of plate convergence can be used to speculate on the spatial distribution of sediment at the time it was deposited. Deposition of pelagic mudstone, turbidite limestone, and radiolaria-bearing chert and siliceous shale occurred from Early to early Middle Ordovician time over a priod of about 20 m.y. Assuming a maximum plate convergence rate of 2 cm/yr the plate carrying the sediment would have moved approximately 400 km (250 mi) toward the trench axis during deposition of the pelagic sediment. The sparse presence of N. gracilis graptolite zone species collected from red shale (Stephens and others, 1982) suggests that the plate was not within the sphere of influence of near-trench hemipelagic sedimentation until Middle Ordovician time. N. gracilis and possible G. Teretiusculus Zone graptolites collected from hemipelagic and trench axis deposits suggests that near-trench and trench axis sedimentation occurred over a short span of approximately 5 m.y. prior to accretion. Hemipelagic sedimentation replaced pelagic sedimentation approximately 160 km (100 mi) from the Middle America Trench axis (Moore, Watkins and others, 1982) off southern Mexico. In addition, Coulbourn and others (1982) note that onset of hemipelagic sedimentation at Site 495 in the Middle America Trench off Guatemala occurred when it was 900 km (560 mi) seaward of its present position 22 km (14 mi) seaward of the trench. Biostratigraphic data from the Greenwich slice, however, suggest that hemipelagic sedimentation became dominant over pelagic sedimentation when the plate was quite close to the trench axis. The combined duration of the

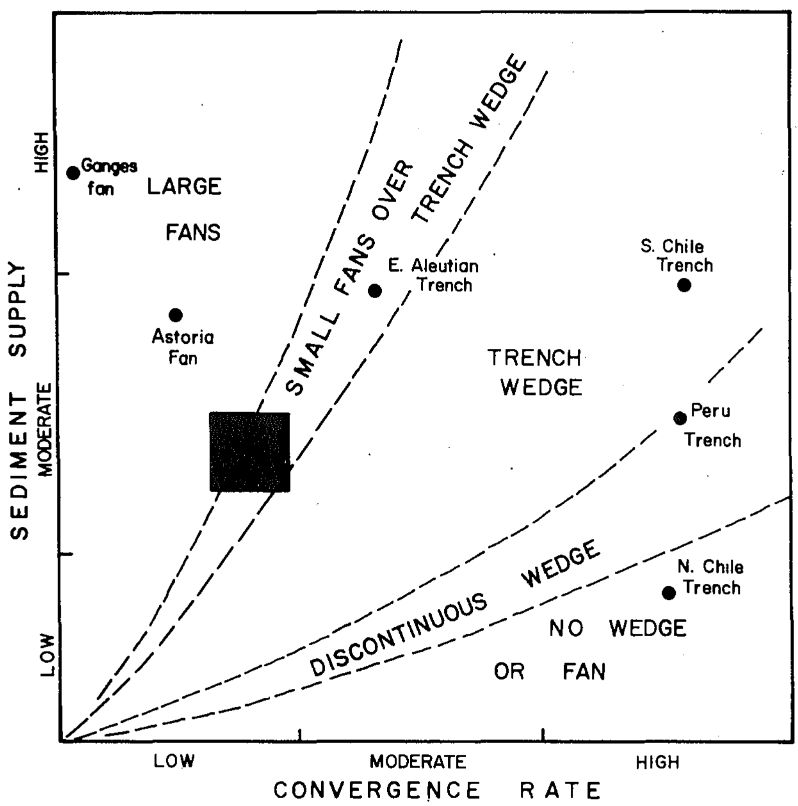


Figure 36. Convergence/sedimentation rate plot for eastern Pacific convergent margins (from Schweller and Kulm, 1978). Approximate position of the Greenwich slice is shown by the black box.

N. gracilis (Berry's zone) and G. teresciulus graptolite zones, those zones reported from the turbidite and hemipelagic deposits, is approximately 5 m.y. (fig. 22). At a maximum convergence rate of 2 cm/yr the plate would have moved approximately 100 km (60 mi) during sedimentation of these deposits. Moore, Watkins and others (1982) noted that although the first appearance of terrigeneous detritus occurred when Site 487 in the Middle America Trench off southern Mexico was 160 km (100 mi) seaward of its present position approximately 15 km (10 mi) seaward of the trench axis the greatest surge in hemipelagic sedimentation occurred when the site was approximately 40 km (25 mi) seaward of the trench. Although the biostratigraphic control required for resolution of the problem described above cannot be realized in rocks as old as the Greenwich slice and there may be problems with overlapping faunal zones the proposed scenario does fall within an actualistic framework defined by the maximum and minimum limits of hemipelagic sedimentation of the well-studied Middle America Trench.

Possible restrictions of the spatial extent of terrigeneous hemipelagic sedimentation such as that of the Greenwich slice include the degree of ocean plate downbending and vertical relief of the outer rise or peripheral bulge. Indeed, von Huene and others (1982) noted 2000 m (6500 ft) of relief between the Middle America Trench axis off Guatemala and the seaward limit of the outer trench slope. Peripheral bulges are observed seaward of ocean trenches and are characterized by a 200- to 1000-km (120 to 600 mi) wide band parallel to the trench (Hanks, 1973; Watts and Talwani, 1974). Maximum uplift of the ocean floor occurs 120 to 150 km (75 to 100 mi) from the trench axis and uplift ranges from 300 to 500 m (1000 to 1600 ft) but can be as great as a kilometer above the abyssal plain (Forsyth, 1980; Dickinson and Seely, 1979). Greater relief of the outer swell or peripheral bulge would tend to confine bottom sediment layers and turbidity currents to the "axial zone" rather than allowing them to flow onto the abyssal plain. Only in a situation of very high sedimentation rates would sediment be able to flow over the outer swell onto the abyssal plain (e.g., Washington/Oregon, Makran). Likewise, the degree of downbending of the subducting plate will exert some degree of control on sedimentation patterns in near-trench areas. Cross and Pilger (1982) maintained that low convergence rates increase the degree of downbending of the subducting plate. The proposed low convergence rate of the proto-Atlantic Ocean is consistent with a high degree of crustal downbending which would tend to limit the spatial distribution of near-trench hemipelagic sedimentation. It is likely that the hemipelagic deposits of the Greenwich slice were restricted to an area within 100 km (60 mi) of the trench axis due to a high degree of crustal downbending and the presence of an outer swell. This is consistent with the average distances of maximum uplift of modern outer swells (120-150 km (75 to 100 mi)). It is important to note that a high degree of normal faulting of the subducting plate and a greater chance of sediment ponding thereby reducing the likelihood of long-distance axial transport of trench turbidites.

In summation, the major points of the sedimentary evolution of the Greenwich slice are (fig. 37):

1. Pelagic sedimentation, dominated by eupelagic red clay and lesser proportions of calcareous turbidites, radiolarian chert, and associated rocks, occurred from at least Early Ordovician (late Tremadocian) to Middle Ordovician (N. gracilis Zone) time, a period of approximately 20 m.y. at an average rte of sedimentation of 1 to 5 m/m.y. and recorded plate movement of approximately 400 km (249 mi) (fig. 37, pt. A).

2. Normal faulting associated with migration of the plate across the outer swell and/or into the area of crustal downbending resulted in soft-sediment folding of the pelagic sediments. It was somewhat after this time that the Jonestown basalts were extruded onto hemipelagic sediments, indicating that the volcanic activity occurred less than 100 km (62 mi) from the trench axis (Lash, 1984a). In light of this it is likely that the basalt flows were related to a pressure release in the downgoing plate that resulted from normal faulting related to downflexing and extension of the

subducting lithosphere as it descended into the subduction zone (Echeverria, 1980; Lash, 1984a). The low convergence rate and resulting increase in crustal downbending and related extension is consistent with this scenario (fig. 37, pt. B).

3. The presence of an outer swell and high degree of downbending of the subducting lithosphere probably played an important role in restricting hemipelagic sedimentation to the near-trench (within about 100 km (62 mi) of the trench axis) area. Hemipelagic sedimentation did not extend much beyond approximately 100 km (60 mi) of the trench axis which probably marks the inner limit of the outer swell (Dickinson and Seely, 1979; fig. 37, pt. B).

4. By approximately N. gracilis time the plate had moved close enough to the trench axis to receive coarse clastic deposits. Minor intrusive volcanic rocks of the Jonestown volcanic rocks were intruded into trench axis deposits and were probably related to normal faulting in the trench axis (Coulbourn, 1981, 1982). Trench axis sedimentation which probably averaged 200-300 mm/1000 yrs was characterized by small fans that may have built over hemipelagic deposits or a discontinuous trench wedge. The small fans were fed by submarine canyons that received detritus from rivers whose drainage basins included plutonic-metamorphic and subduction complex rocks (fig. 37, pt. C).

5. The minor but conspicuous boulder conglomerate units resulted from gravity sliding and collapse of accreted and oversteepened trench and abyssal plain deposits and were subsequently incorporated into the subduction complex (fig. 37, pt. D).

Accretion in the Greenwich slice occurred somewhat earlier than that of the Scottish Uplands which like the Greenwich slice formed in response to closure of the proto-Atlantic Ocean (Leggett and Casey, 1983). Leggett and Casey (1983) noted that accretion in the Scottish Uplands took place over a period of 50 to 60 m.y. This is significantly longer than the

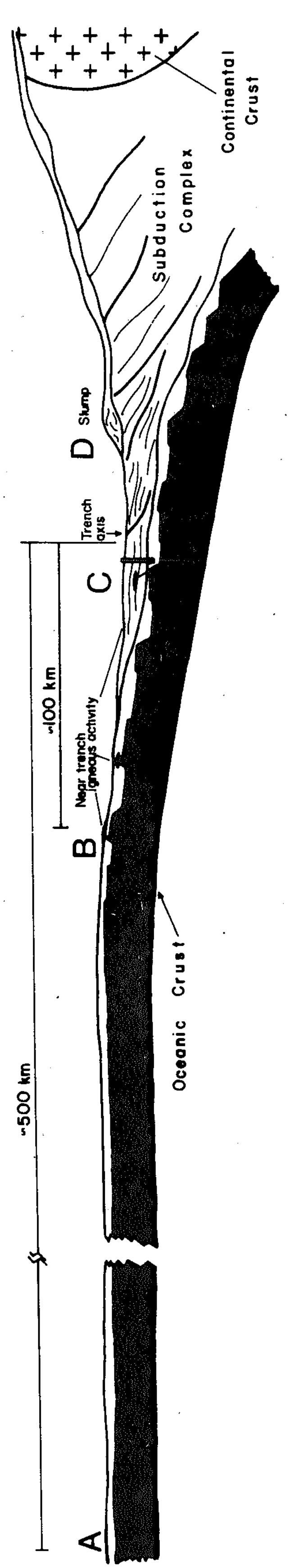


Figure 37. Tectonic evolution of the Greenwich slice. Refer to text for detailed discussion. Heavy line to the lower right of "C" represents a coarsening-upward sequence similar to those of the Greenwich slice.

25 m.y. duration of subduction and accretion recorded by the Greenwich slice. In addition, the Scottish Uplands accretionary prism is a complex of ten distinct sequences of ocean floor, near-trench, and trench axis deposits (Leggett and Casey, 1983) in which individual sequences become younger toward the bottom of the pile, a feature that forms by sequential offscraping of thick trench sequences (Karig and Sharman, 1975). The Greenwich slice, unlike the Scottish Uplands, consists of a single coarsening-upward sequence that has been fragmented and folded by later deformation. This is quite similar to the fragmented offscraped sequence on the Kodiak Islands of Alaska (S. Box, personal communication, 1983) and records a short period of subduction probably related to closure of a small basin. This is consistent with recent tectonic models of this part of the Appalachian orogen that place a small ocean basin between the North American craton and a rifted micro-continent to the southeast (Lash, 1980; Lash and Drake, 1984; Shanmugam and Lash, 1982). The Scottish Uplands accretionary prism, on the other hand, records prolonged closure of a large ocean from Early Ordovician to Middle(?) Silurian time (Leggett and Casey, 1983).

Accretion-related deformation in the Greenwich slice

Marine geophysical and geological data collected over the past ten years from active convergent margins round the world have led to the formulation of several different models for the structural evolution of forearc complexes (Grow, 1973; Hamilton, 1977; Ladd and others, 1978; Seely and others, 1974; Karig, 1974; Karig and Sharman, 1975; Shipley and others, 1980, 1982). In particular, recently recovered cores from active accretionary complexes by the Deep Sea Drilling Project (Karig and others, 1975; Moore, Biju-Duval, and others, 1982; Arthur and others, 1980; von Huene and Aubouin, 1982; among others) have greatly increased our understanding of subduction processes as well as forearc evolution. However, because coring and acoustic methods cannot resolve much of the internal structure of forearc regions it is particularly useful to study exposed subduction complexes. Many of the meso- and microscopic features of active subduction complexes are readily observed in their exposed counterparts (Cowan, 1974, 1982a; Leggett and others, 1982; Moore and Karig, 1980; among others). Previous investigations of exposed accretionary complexes such as the Kodiak Islands of Alaska (Moore and Allwardt, 1980), the Franciscan Complex of California (Cowan, 1982a), the Chrystalls Beach Complex of New Zealand (Nelson, 1982), the Barbados Ridge Complex (Speed, 1983; Speed and Larue, 1982), and Nias Island (Moore and Karig, 1980) have yielded extremely useful information that, when combined with acoustic and core data collected from active forearc regions, presents a fairly complete picture of offscraping and accretion processes.

This chapter discusses the early tectonic evolution of a proposed (Lash, 1980, Lash and Drake, 1984) early Paleozoic subduction complex in the central Appalachian orogen. The meso- and microscopic fabric of these rocks will be considered in light of recent DSDP results and other studies of sediment deformation at convergent margins in order to consider if the style of deformation illustrated by these rocks is consistent with what is known of sediment deformation in active forearc regions and current concepts of offscraping and accretion.

Rocks of the Greenwich slice were affected by three major phases of deformation (Lash, 1980), the first being the subject of this section. The second and third phases record possible Taconic and Alleghenian orogenic events that are discussed in the section "Structural geology of the Great Valley and Valley and Ridge). Taconic and Alleghenian deformation resulted in folding and faulting of the Greenwich slice and only local overprinting of fabrics (Lash, 1980). Evidence for this includes folding of zones of stratal disruption related to the first phase of deformation by second phase. These relations allow for detailed meso- and microscopic analysis of the earliest phase of deformation.

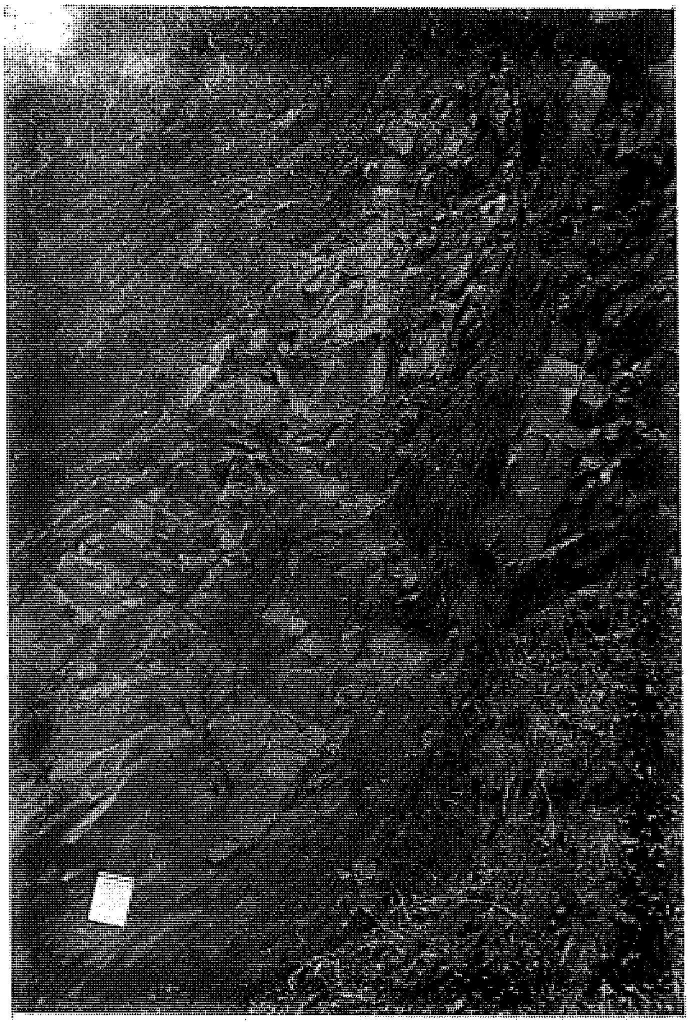
Structural features.

The Greenwich slice can be described, in part, as a broken formation in that it is an internally fragmented unit that contains no exotic blocks. Primary structures are severely disrupted in a number of horizons despite the fact that the rocks are unmetamorphosed. Zones of mesoscopically ductile deformation characterized by sequences of broken, irregular, or discontinuous sandstone beds are interlayered with non-deformed sequences (fig. 38). A complete spectrum of deformation ranges from (1) beds that show variations in thickness but have not been pulled apart (fig. 38A) to (2) beds that are pulled apart but can still be recognized as the same bed (fig. 38B) to (3) beds that are severely disrupted and are difficult to trace across the outcrop (fig. 38C). Interlayered shale is commonly cut by pervasive anastomosing, locally polished and striated scaly cleavage surfaces (fig. 39) whereas shale interlayered with non-deformed sandstone is typically cleaved by a slaty cleavage that probably formed in response to a later (Taconic or Alleghenian) deformation (see section "Structural geology of the Great Valley and Valley and Ridge"). It is particularly interesting to note that the most intensely deformed sandstone beds are invariably associated with thick (2 m+) sequences of scaly cleaved mudstone and that the severity of deformation decreases away from these zones.

Within the zones of stratal disruption sandstone beds of all thickness are found in all stages of dismemberment indicating that deformation is not controlled by bed thickness. Individual boudins range from a few centimeters to several meters long and generally show no relict sedimentary structures. This contrasts greatly with the mesoscopically non-deformed sandstone adjacent to the deformation zones that illustrate well-preserved primary structures. The long axes of boudins lie within a plane defined by scaly cleavage. Mesoscopically brittle features such as extensional shear fractures, extension fractures at boudin necks (Cowan, 1982a), and slickensided boudin surfaces (Moore and Allwardt, 1980) are difficult to find and not present in most deformation zones. A particularly interesting feature of some of the boudins is the presence of bulbous protrusions on their surfaces that resemble load casts (fig. 40).

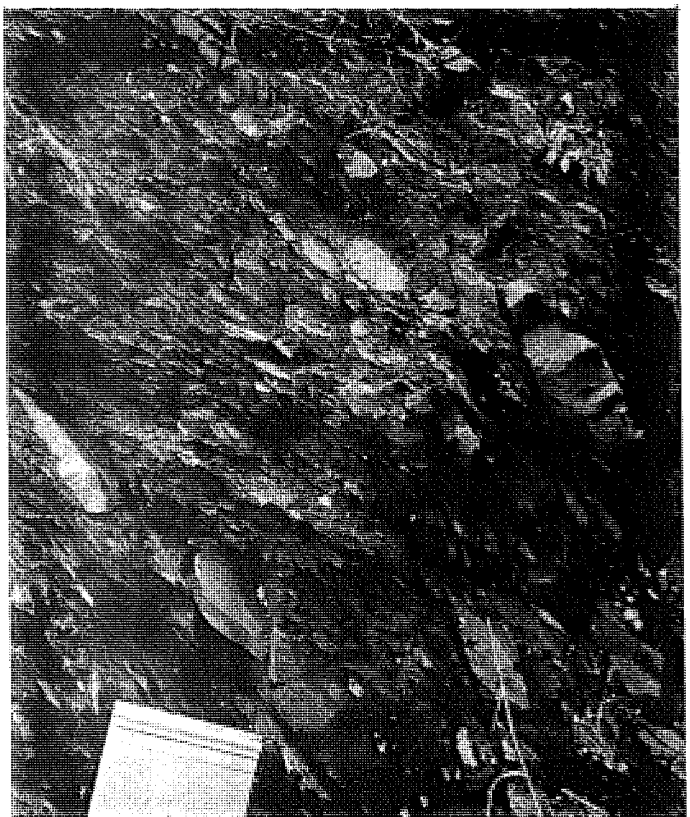
Mesoscopic fabrics similar to that of the Greenwich slice have been reported from numerous active and ancient accretionary complexes (e.g., Kodiak Islands, Moore and Allwardt, 1980; Franciscan Complex, Cowan, 1982a; Central American Trench, Lundberg and Moore, 1981; Moore, Watkins, and Shipley, 1981). However, it is possible to explain fabrics such as these by processes other than those related to subduction (Jacobi, in press). The style of deformation illustrated by the Greenwich slice is very similar to that resulting from any one of the following mechanisms: (1) sliding or slumping of incompletely consolidated sediment at passive as well as active margins (Kleist, 1974; Jacobi, in press); (2) tectonically induced deformation of consolidated sediment (Moore and Allwardt, 1980); (3) tectonically induced deformation of incompletely consolidated sediment (Cowan, 1982a). Analysis of mesoscopic structures alone, then, may not yield an unequivocal deformation history of a complex unit such as the Greenwich slice.

Analysis of microscopic fabric, when combined with results of mesoscopic investigations allows for better assessment of the physical nature of sediment during deformation (for example, see Cowan, 1982a). Petrographic and Scanning Electron Microscopic methods were employed to examine the deformed mudstone of the Greenwich slice. These investigations indicate a preferred orientation of platy grains within or along the margins of most scaly cleavage planes (fig. 41). Domains of preferred orientation pass into areas of more randomly oriented grains which probably record a detrital fabric (fig. 41). Significantly, there is little or no change in the grain size of individual phyllosilicate minerals away from the zones of reorientation suggesting that there has been little or no recrystallization related to scaly cleavage formation. Many scaly cleavage surfaces are characterized by microstriations and microsteps (fig. 42). Microfractures filled with sparry calcite (fig. 43) subparallel to scaly cleavage are common and resemble "closed" or healed fractures described by Arthur and others (1980) and Carson and others (1982). Moreover, in some samples scaly cleavage can be seen being truncated by fractures (fig. 44) suggesting that fracturing took place after scaly cleavage formation.



A

B



C

Figure 38. Stratal disruption of Greenwich slice sandstone: A) pinch-and-swell; B) boudinage where individual beds can still be traced across the outcrop; C) intense boudinage where individual beds are difficult to trace across the outcrop.

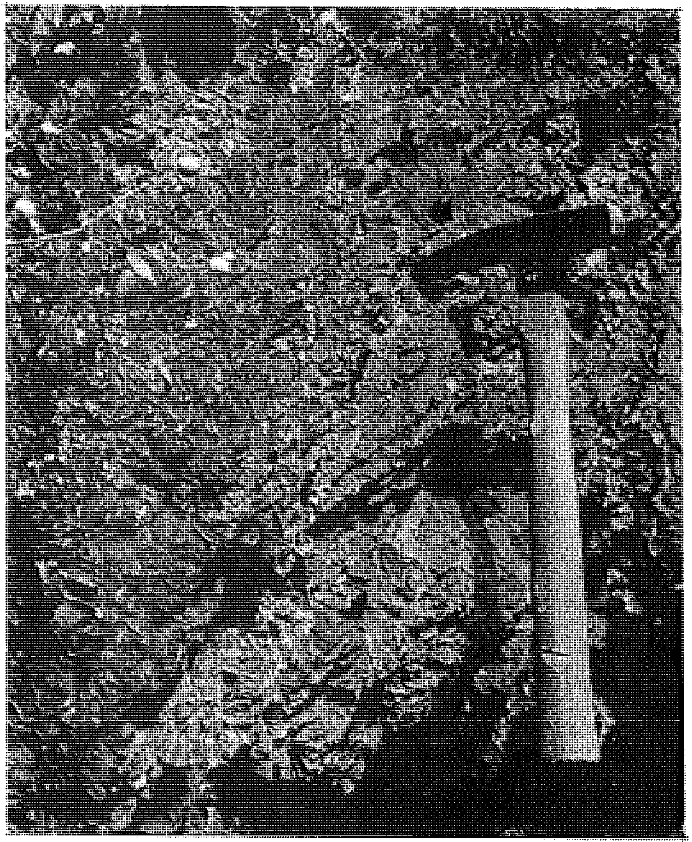
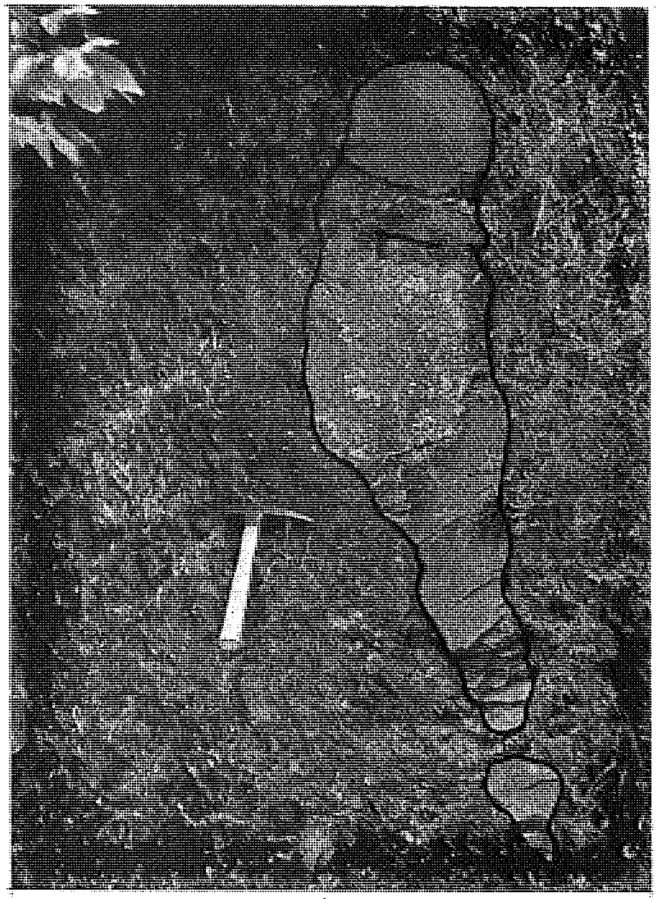


Figure 39. Scaly cleavage.

Petrographic examination of deformed and non-deformed sandstone beds interlayered with the scaly cleaved mudstone were carried out to consider how mesoscopically ductile deformation was accommodated on a microscopic scale. Evidence of widespread cataclasis such as microgouge or shear zones (Dunn and others, 1973; Cowan, 1982a) are rare in the deformed sandstones of the Greenwich slice. Sandstone shows little or no effect of granulation due to shearing regardless of where on the boudin the sample was taken and detrital textures are well preserved (i.e., a minimum of cataclasis; fig. 45A). Mesoscopically undeformed sandstones, however, contain a much greater percentage of shear zones (fig. 45B). The shear zones are typically 0.5 to 1.0 mm thick and are characterized by dark, opaque material and granulated quartz and feldspar grains (fig. 45B) some of which have clearly been fractured. Argillaceous lithic fragments have been severely deformed and are often difficult to differentiate from the original matrix. One cannot rule out the likelihood that pressure solution produced corroded and sutured quartz grain contacts. It is likely that pressure



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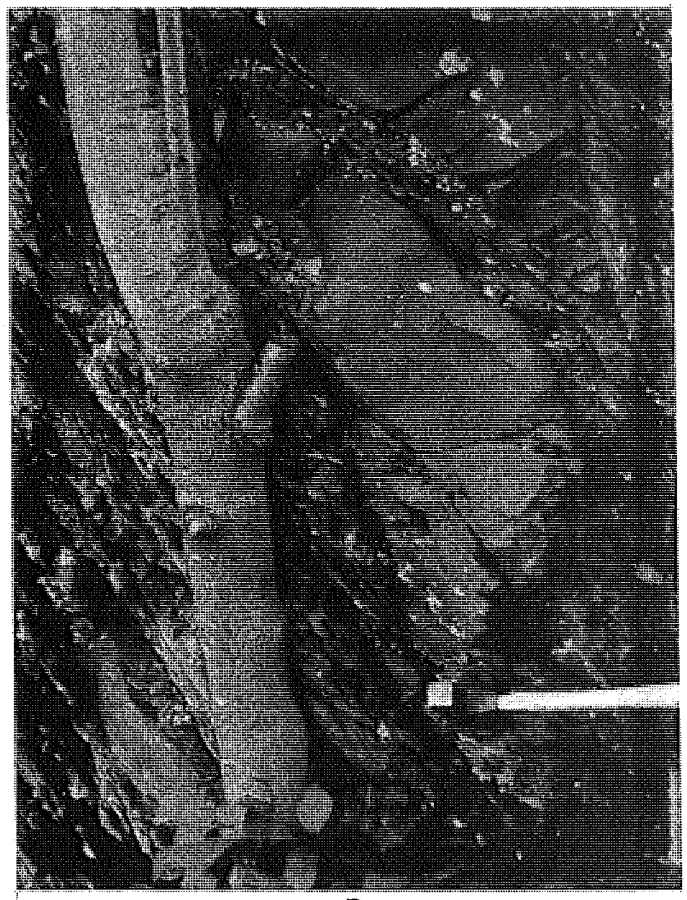


B

Figure 40. Surface irregularities on boudins: A) bulbous protrusions on all sides of a boudin; B) bulbous protrusions on a boudin surface (just above hammer) associated with small "flames" of mud that penetrate into sand. Primary sedimentary structures are lacking in both examples.

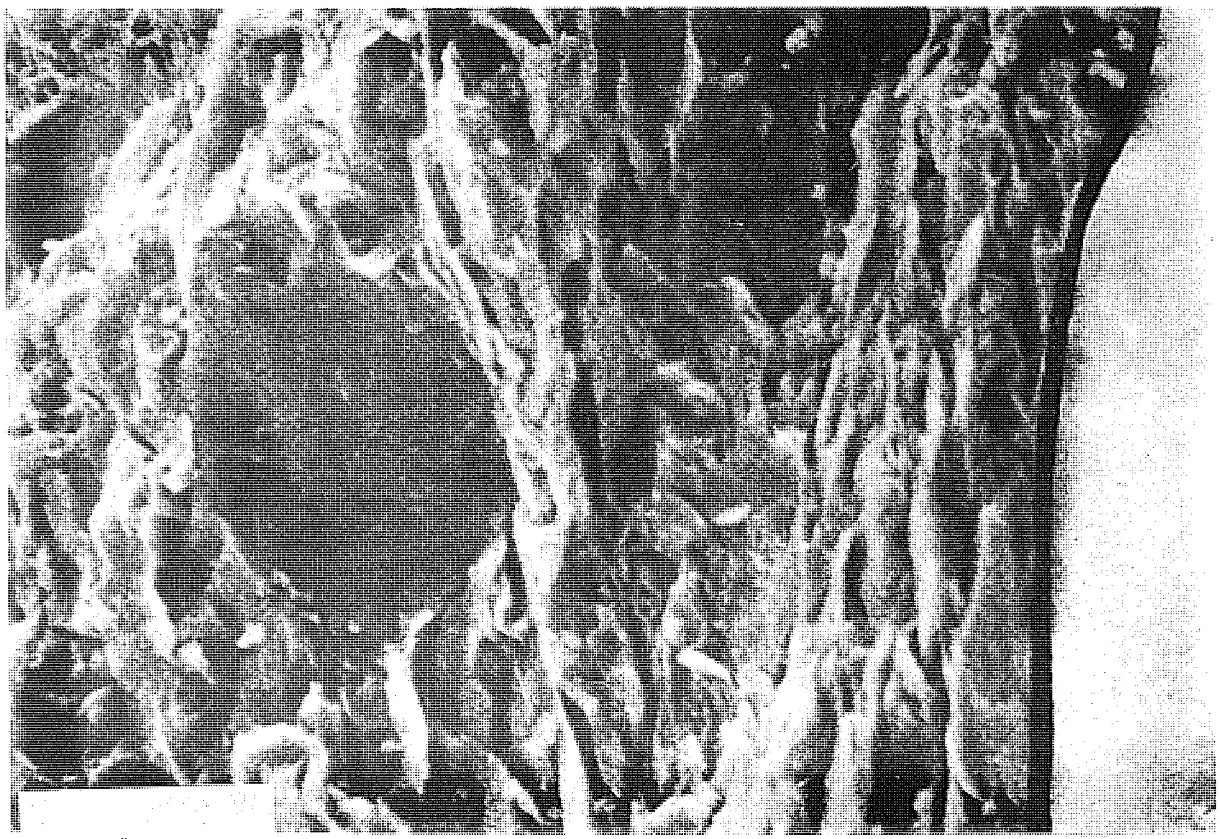


Figure 41. A) Phyllosilicates reoriented into parallelism with a scaly cleavage plane outlined in black at right (the out-of-focus area to the right of the line is the scaly cleavage plane itself, which is oriented approximately 90^{o} to the plane of the photomicrograph (scale=30 microns).

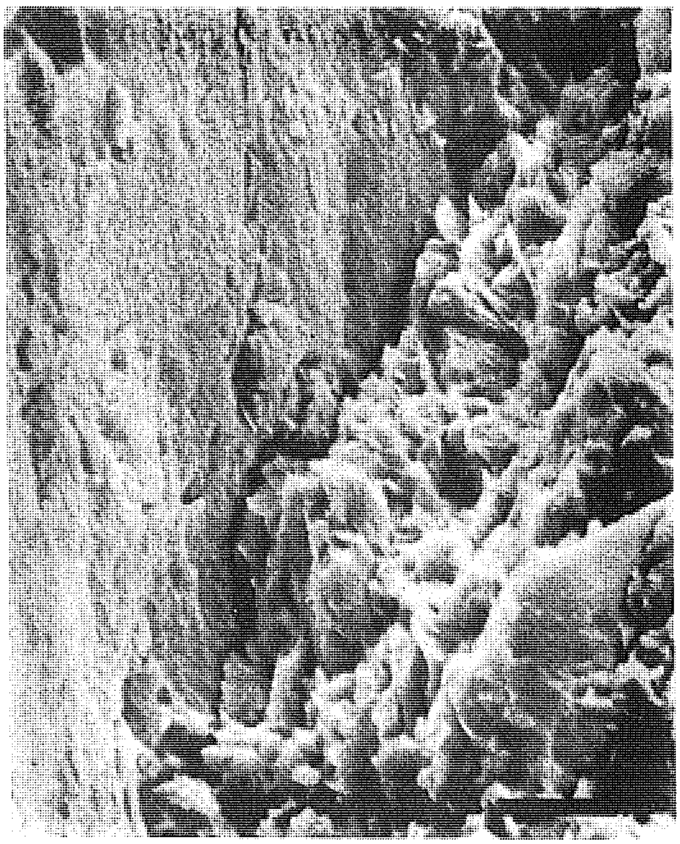


Figure 41. B) Partially reoriented phyllosilicates adjacent to a scaly cleavage surface (scale=30 microns).

Figure 42. Photomicrograph of the surface texture of a scaly cleavage plane. Note the microstriations and microsteps (scale=100 microns).

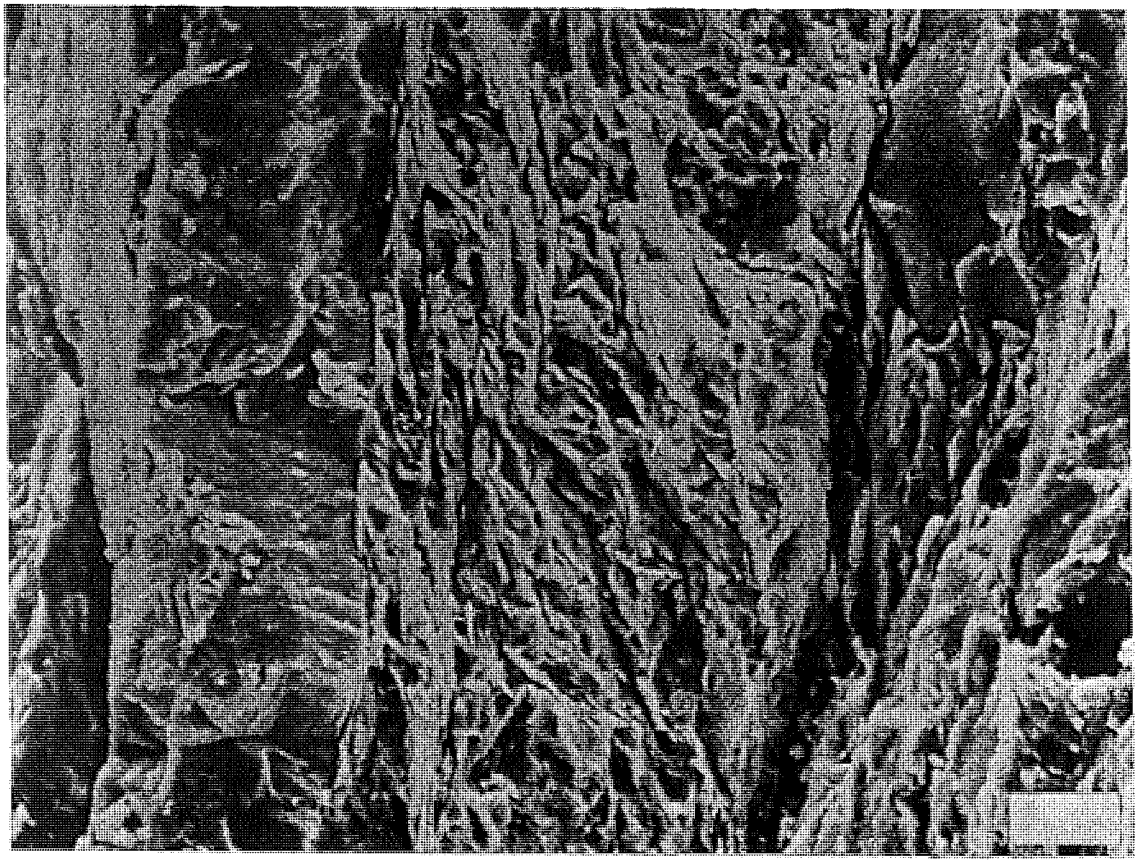


Figure 43. Photomicrograph of a microfracture filled with sparry calcite (at left). Note that phyllosilicates adjacent to the fractures are subparallel to the fracture (scale=100 microns).

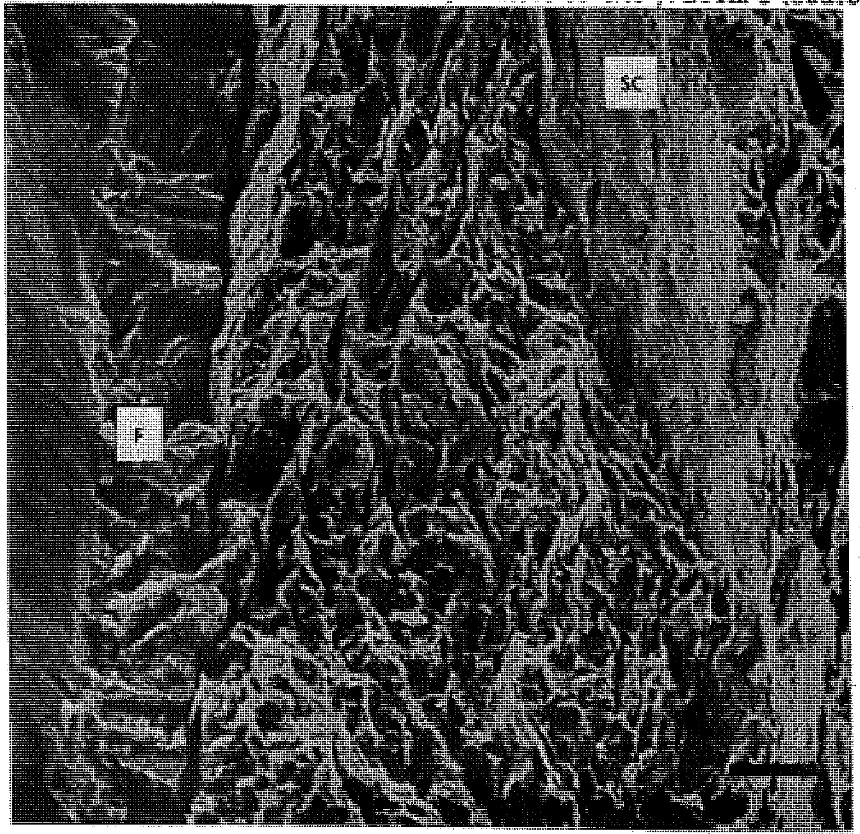
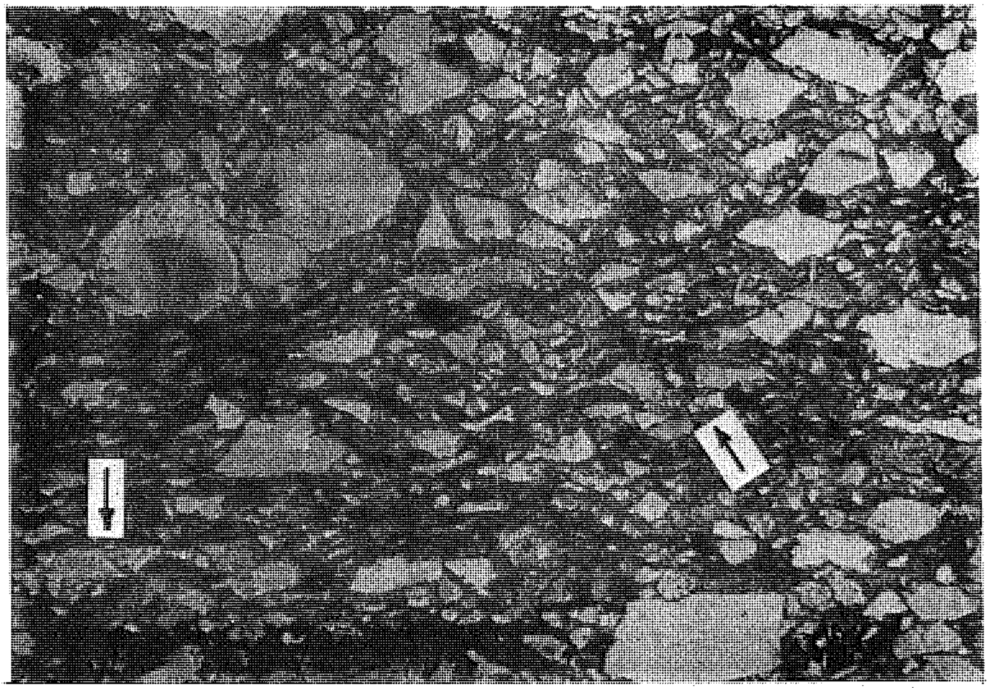
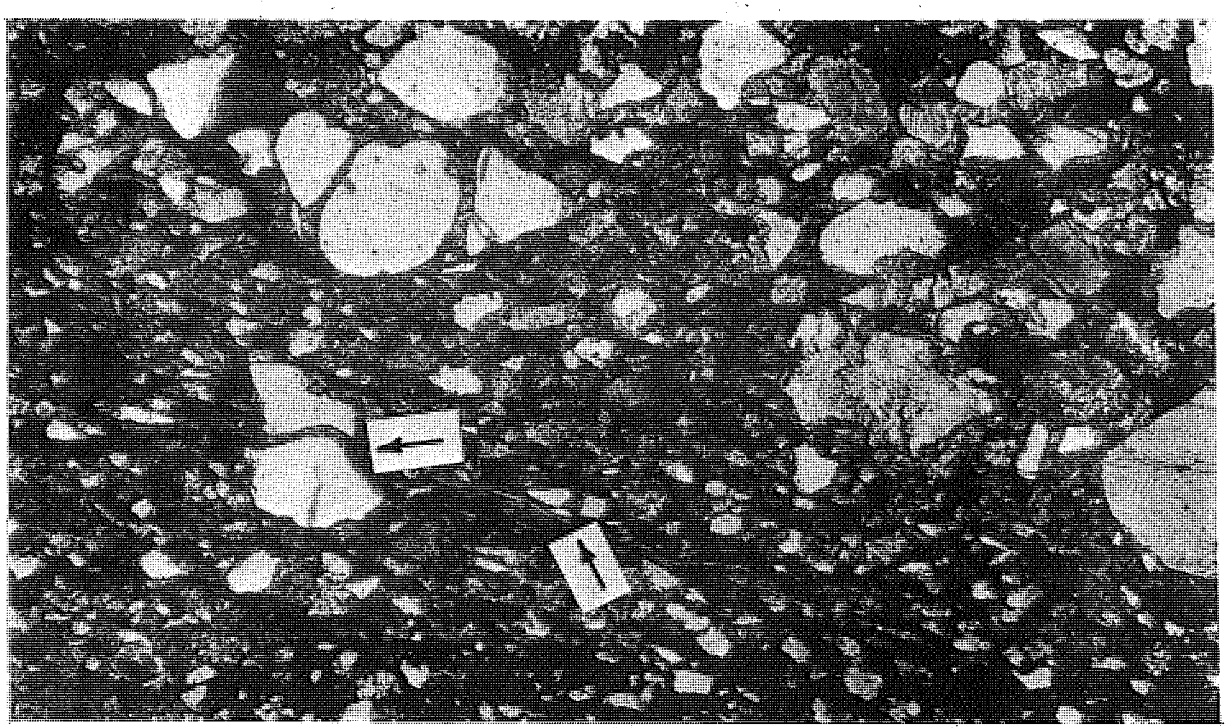


Figure 44. Photomicrographs of subparallel fracture (F) and scaly cleavage (SC) plane. The fracture truncates the scaly cleavage plane toward the top and out of the field of the photomicrograph. Note the presence of reoriented phyllosilicates adjacent to the scaly cleavage plane and the random orientation of phyllosilicates adjacent to the fracture (scale=100 microns).



A



B

Figure 45. Photomicrographs of internally disrupted Greenwich slice sandstone (scale=0.5 mm): A) shear zone, arrow points to grain breakage; B) shear zone, note the effects of grain breakage (arrow on right) and solution (arrow on left) in the grain size reduction zone. In addition, note the abrupt contact of the shear zone and adjacent non-deformed sandstone.

solution acted in conjunction with granulation in these rocks. In any event, the lower percentage of zones of grain size reduction within the mesoscopically deformed sandstones suggests that granulation and/or pressure solution was not as important a process in these rocks as it was in the non-deformed sandstone. The microscopic studies, then, show a conspicuous difference in deformation style between deformed and non-deformed sandstone. That is, the mesoscopically deformed sandstone beds are actually less deformed on a microscopic scale than the non-deformed sandstone beds.

State of sediment during deformation.

As stated above, a major stumbling block in the understanding of chaotic sedimentary rocks is the differentiation of fabrics formed by (1) gravity-driven slumping and sliding of soft sediment, and (2) those resulting from tectonically induced deformation of either unconsolidated or lithified sediment. Comparison of deformed sediment (boulder conglomerate) and unequivocal olistostromal deposits of the Greenwich slice indicates that the disrupted horizons are not submarine slide deposits (compare figs. 19, 20, and 37). In addition, the lack of folds that are common to submarine slide deposits (Jacobi, in press) argues against an olistostromal origin for the disrupted horizons of the Greenwich slice. Differentiation of tectonically induced deformation of unconsolidated sediment from similar deformation of lithified sediment, however, is not very easy.

Handin and others (1963) maintained that the deformational behavior of sand is controlled by the relative intensities of confining and fluid pressures. They noted that at low fluid pressure sandstone deforms by cataclastic flow whereas during conditions of lithostatic fluid pressure (i.e., confining pressure equals lithostatic fluid pressure) deformation occurs by intergranular movement of grains cushioned by pore fluid. The lack of cataclastic deformation in disrupted sandstone beds of the Greenwich slice is suggestive of the latter and indicates that these beds were poorly consolidated and water-rich (i.e., highly pressured to overpressured pore fluid) during deformation. This is supported on a mesoscopic scale by the presence of the load features found on surfaces of some of the boudins described earlier (see fig. 40).

Abrupt variations in the style of deformation of sandstone (i.e., zones of ductily deformed sandstone interlayered with sequences of non-deformed sandstone beds) is a particularly intriguing characteristic of the Greenwich slice. Cowan (1982a) described a similar behavior of sandstone sequences of the California Coast Ranges and related it to local variations in pore fluid content and state of dewatering at the initiation of deformation rather than variations in confining pressure and strain rate during deformation. Results of the present study substantiate Cowan's observations and suggest that the zones of ductilely deformed sandstone in the Greenwich slice were water-rich horizons during deformation whereas the less deformed and non-deformed beds were characterized by lower fluid contents and pressures.

Deformation behavior of mud is more difficult to assess. Petrographic and SEM investigations of deformed mudstone of the Greenwich slice indicate that phyllosilicates were preferentially reoriented along scaly cleavage planes. Reorientation of phyllosilicates could have occurred as a result of preferential dewatering of water-rich mud (Lundberg and Moore, 1981) or brittle shear of semi- to fully-consolidated mud along scaly cleavage planes (Cowan and others, in press). The presence of fractures filled with sparry calcite in deformed mudstone is a feature common to deformed semi-lithified or indurated mud (Arthur and others, 1980) and is indicative of restricted pore fluid migration and at least locally overpressured pore fluid conditions. In addition, the presence of microslickensides and microsteps on scaly cleavage surfaces suggests that the mudstone was at least semi-lithified when it deformed. Moore, Biju-Duval, and others (1982) noted the presence of slickensided clays at Site 541 of IPOD Leg 78A in an area where water pressure is roughly equal to lithostatic pressure (Westbrook and Smith, 1983). The presence of healed fractures and slickensided scaly cleavage surfaces of disrupted mudstone of the Greenwich slice records a deformation history similar to that of the disrupted clays at Site 541. It is likely, then, that the Greenwich slice mudstone was at least semi-lithified at the time of deformation and that scaly cleavage formation occurred as a result of brittle failure along shear planes in the compacted mud probably concomitant with and related to stratal disruption of sand.

An important consideration at this point is the influence that compaction and dewatering of mud had on interlayered sand. Cowan (1982a) proposed that mud compacts at a rate greater than that of interlayered sand. It is likely that the deformed Greenwich slice sand served as a reservoir for water expelled from tectonically (Carson, 1977) compacted water-rich mud horizons and as such compacted later than the mud (Cowan, 1982b). This is supported by the common association of deformed sandstone beds and shale-rich horizons noted earlier. The consolidated mud would then act as an impermeable barrier to fluid migration from the sand thereby promoting increased pore fluid pressures within the sand bodies during continued deformation. Internal disruption of primary structures within the sand occurred during deformation. Overpressured pore fluid conditions within the sand ultimately led to hydrofracturing of the surrounding lithified mud. The fact that the fractures can be seen cutting scaly cleavage planes indicates that hydrofracturing occurred after scaly cleavage formation. The presence of domains of oriented platy grains adjacent to a small number of fractures (see fig. 43) suggests that some scaly cleavage planes evolved into fractures and may have to some extent controlled the orientation of fractures. The fact that all of the fractures observed are subparallel to bedding is consistent with propagation of fractures in a direction roughly perpendicular to the major compressional axis (perpendicular to bedding) and is consistent with fluid movement along horizontal hydrofractures toward the front of an accretionary complex, as suggested by Westbrook and Smith (1983, p. 282).

The key feature of the process described above is the presence of water-rich mud horizons. These zones are likely sites of fluidization and liquefaction of interbedded sand during tectonically induced compaction of the mud. The increased fluid content of the sand resulting from compaction of thick, water-rich mud horizons allowed for intergranular flow during deformation. Horizons characterized by high sand/shale ratios, on the other hand, may not have the additional water required to attain lithostatic fluid pressure conditions and as such are characterized by a more brittle type of internal failure. Faas (1982) noted a water-rich, undercompacted hemipelagic mud horizon such as those postulated in the present study at DSDP Site 499 in the Middle America Trench off the coast of Guatemala. He maintained that variations in pore fluid content, sediment strength, and state of compaction of Middle America Trench sediments at Site 499, a possible analogue of the site of deposition of the Greenwich slice turbidite deposits, are likely the result of variations in lithology and sedimentation rate rather than tectonic stresses related to subduction of the Cocos Plate. Whereas low rates of sedimentation tend to promote overcompaction of sediments, greater depositional rates may trap pore fluid in impermeable sediment such as hemipelagic mud, resulting in a buildup of pore fluid in underconsolidated horizons (Faas, 1982; Carson and Bruns, 1980). Upon tectonically induced compaction, this water may migrate into nearby sand, resulting in the variations in deformation style illustrated by rocks of the Greenwich slice.

In summation, analysis of the microscopic fabric of the Greenwich slice suggests that vertical variations in pre-deformation pore fluid content of near-trench and trench axis sediments on the subducting plate played a great role in determining the style of deformation of the sediment during subsequent accretion-related compaction and deformation. The lack of cataclastic fabric in disrupted sandstone beds suggests that they were unconsolidated and water-rich at the time of deformation. Mesoscopically ductile deformation was accommodated on a microscopic scale by intergranular movement cushioned by highly pressured pore fluid. Non-deformed sandstones, on the other hand, show a greater degree of cataclasis and pressure solution suggestive of lower pore fluid contents during deformation. Local variations in the state of consolidation and original pore fluid content related to such variables as rate of sedimentation and lithology are apparently responsible for abrupt variations in style of accretion-related deformation of the Greenwich slice sediments. Field relations, in particular the association of thick mudstone intervals and severely deformed sandstone beds, microscopic analysis, and consideration of previous investigations concerning the behavior of mud suggest that it compacted sooner than the interlayered sand and probably contributed to the high water content of deformed sand during the early stages of tectonically induced compaction (Carson, 1977) of the water-rich hemipelagic mud horizons. Water escape from the compacting mud may have been facilitated by formation of veins (Cowan, 1982b; Lundberg and Moore, 1981; Arthur and others, 1980).

Any evidence of the prior existence of veins has apparently been obliterated by subsequent (scaly cleavage formation) deformation. Mesoscopically ductile behavior of the semi- to fully-consolidated mud was accommodated on a microscopic scale by brittle failure along anastomosing scaly cleavage planes, concomitant with deformation of the unconsolidated sand. Overpressured fluid conditions within the mesoscopically deformed sand during deformation probably resulted in hydrofracturing of the lithified mud subsequent to scaly cleavage formation.

Review of offscraping and accretion processes.

As shown earlier, important influences on the rate and amount of sediment accreted at a convergent margin include (1) the rate and amount of sediment accumulating in the trench and (2) plate convergence rate perpendicular to the trench axis. Of these rate of sedimentation appears to be the most important. Sedimentation rate in a particular trench is related to nearby drainage patterns, proximity to sediment source areas, and variations in sea level, all of which may be independent of convergence rate. Plate convergence rate, however, does determine the residence time or the amount of time sediment deposited in a trench remains in the trench prior to offscraping and accretion. Significantly, the rate of sedimentation and the residence time of the sediment in a trench can exert a major influence on the amount of pore fluid in the sediment prior to offscraping and accretion.

Carson (1977) noted that the greatest amount of tectonically induced compaction, dewatering, and associated deformation occurs in the lower inner trench slope immediately after sediment is offscraped. Thus the physical nature of the sediment of the descending plate, that is the thickness and internal strength of the sediment, is very important in determining the amount of sediment accreted and the manner in which the sediment deforms in the lower inner trench slope after it is accreted. Moore (1975) proposed that the predominance of turbidite deposits in most accretionary complexes is the result of preferential accretion of buoyant, relatively weak, clastic trench sediments along a decollement whereas the older and stronger diagenetically mature biogenic ooze deposits would be subducted. Many workers have suggested that abnormally high fluid pressure along the decollement may facilitate the uncoupling and underthrusting associated with accretion of sediment. Recent investigations of Karig and Ingle (1975) and Nasu and others (1982) have illustrated a decollement between trench wedge and pelagic sediment at DSDP Site 198 in the Nankai Trough. Moore, Biju-Duval, and others (1982) noted selective subduction of diagenetically mature pelagic sediment and accretion of overlying weak hemipelagic sediment in IPOD Leg 78A in the Barbados Ridge Complex. Von Huene and others (in press), however, illustrated that the decollement at the eastern end of the Aleutian Trench is not found at the boundary between trench fill and underlying hemipelagic and pelagic sediment although these authors did note that the decollement does occur at a significant change in shear strength of the sediment. McCarthy and others (1982) suggested that the presence of a thick sequence of subducting pelagic sediment may act as a buffer zone between offscraped sediment and the topographically irregular oceanic basement and may prevent incorporation of oceanic crust into the accretionary complex.

Lundberg and Moore (1982) classified forearc terranes on the basis of variations in deformation style into brittle (folding, open deformation) and ductile (stratal disruption, melanges, broken formations). They maintained that variations in internal structure reflect variations in sediment input and not necessarily plate convergence rate. Along these same lines Silver and Moore (1982) suggested that major differences in the internal structure of forearc terranes can be related to variations in the internal strength (i.e., pore fluid content) of the sediment on the subducting plate prior to its incorporation into an accretionary prism. Therefore, deformation of weak accreting sediment will result in stratal disruption.

Offscraping and accretion in the Greenwich slice.

Consideration of the internal strength of the Greenwich slice turbidite and hemipelagic deposits in the context of the overall lithology of the Greenwich slice allows for speculation concerning offscraping and accretion processes during Middle Ordovician subduction in the central Appalachians. Phillips and others (1976), as noted in the previous section, suggested an early Paleozoic convergence rate of less than 2 cm/yr for the Proto-Atlantic Ocean. A slow convergence rate and therefore greater residence time would tend to favor gradual dewatering and strengthening of sediment prior to accretion and deformation. Deformation of a sequence such as this would result in a more brittle or open style of deformation such as that of the Scottish Uplands (Leggett and others, 1979). If, however, sedimentation rates are high and/or the convergence rate was greater than that postulated, gradual dewatering may not have occurred and pore fluid may have become trapped in non-permeable horizons such as thick hemipelagic mud layers (e.g., DSDP Site 499, Faas, 1982). The fact that the majority of highly deformed sandstone beds of the Greenwich slice are associated with thick mudstone intervals lends support to this idea. The abrupt variations in style of deformation of the Greenwich slice sediments record early pre-deformation variations in state of compaction (Faas, 1982) as noted above. Subsequent tectonically induced dewatering and compaction of the undercompacted zones immediately after offscraping (Carson, 1977) resulted in liquifaction of the interlayered sand. Much of this deformation was accomplished under highly pressured to overpressured pore fluid conditions, a feature common to deformation in active accretionary complexes (Moore, Biju-Duval, and others, 1982; Westbrook and Smith, 1983; Arthur and others, 1980).

The paucity of pelagic sediments and the lack of slivers of oceanic crust in the Greenwich slice is best explained by the concept of "selective subduction" (Moore, 1975). The internally weak, water-rich, near-trench deposits were selectively offscraped and accreted while older, more lithified pelagic deposits were carried deeper into the subduction zone. In this respect the Greenwich slice is most like deformed sequences of "clastic-dominated" margins (Shepard and others, 1981). The presence of relatively thick (60 m (200 ft)) sequences of pelagic sediments in the Greenwich slice suggests that the decollement separating the offscraped sediment from the subducted plate did not always root at the pelagic-turbidite (hemipelagic) boundary but often within the pelagic muds. Any pelagic ooze that might have underlain the pelagic clays was probably subducted.

The decollement which resulted in uncoupling of the Greenwich slice deposits from the descending plate was probably in incompetent and overpressured mud (Westbrook and Smith, 1983; von Huene and Lee, 1983). The lack of along-strike continuity of the slivers of pelagic sediment suggests that the decollement did not parallel specific stratigraphic horizons.

A number of studies of active accretionary complexes have cited the internal strength of accreted sediment as an important control in determining the internal structure of the forearc region. The internal structure of the Greenwich slice cannot be classified as either ductile or brittle according to the classification of Lundberg and Moore (1982) as it contains attributes of both styles. However, the importance of sediment strength on the manner in which the sediment deforms can be seen, although on a smaller scale, in the Greenwich. As suggested in a preceeding section, zones of ductile deformation are related to stratigraphic horizons that were water-rich at the time of deformation whereas a more coherent style of deformation is found in horizons of lower pore fluid content and therefore greater internal strength. This illustrates the importance of pore fluid and sediment strength in determining subduction processes and forearc evolution (i.e., internal structure).

Conclusion.

The style of deformation of the Greenwich slice of the Hamburg klippe in eastern Pennsylvania is quite similar to deformation styles in other ancient, as well as active, accretionary complexes around the world. The important roles that local (i.e., vertical) variations in pore fluid content and lithology play in determining the manner in which sediment may deform in the early stages of accretion can not be overemphasized. The meso- and microscopic fabric of the Greenwich slice formed as a result of tectonically induced deformation of offscraped and accreted water-rich sediment at the base of the inner trench slope during Middle Ordovician subduction in the central Appalachian orogen. The present study points out a number of interesting aspects concerning deformation of the Greenwich slice sediments that may have some application in the study of sediment deformation in other ancient as well as active accretionary complexes:

(1) Mud compacted and lithified earlier than interlayered sand. It is likely that at least in some areas sand served as a reservoir for water being expelling out of thick mud horizons during the initial stages of accretion-related deformation.

(2) The association of intensely disrupted sandstone beds and thick intervals of scaly cleaved mudstone can be accounted for by dewatering in the early stages of accretion-related deformation. Much of the expelled pore fluid would probably migrate into adjacent sand beds, liquifying and weakening them, and thereby obliterating any previous sedimentary structures.

(3) Mesoscopic stratal disruption of sandstone beds was accomodated on a microscopic scale by intergranular movement of grains cushioned by high fluid pressure. Adjacent non-deformed beds, however, show a greater degree of cataclasis and/or pressure solution in the form of microgouge zones which is suggestive of lower pore fluid conditions.

(4) Development of scaly cleavage in deformed mudstone occurred by brittle failure along anastomosing shear surfaces in semilithified mud concomitant with deformation of associated sand.

(5) The presence of microfractures filled with sparry calcite attest to locally overpressured pore fluid conditions and restricted pore fluid migration during deformation of the mud. Hydrofracturing of the semilithified mud, which occurred after scaly cleavage development, probably resulted from highly overpressured pore fluid conditions within the deforming sand.

(6) Abrupt local variations in structural style on an outcrop scale are probably related to variations in lithology (i.e., sand versus mud), sedimentation rate, and pre-deformation pore fluid content rather than variations in confining pressure or strain rate. These variations illustrate on a small scale the influence that pore fluid and, therefore, internal strength of offscraped sediment, has on the manner in which sediment deforms the early stages of accretion. Ultimately, this greatly influences the internal structure of the accretionary complex.

Consideration of the physical state of sediment of the Greenwich slice at the time of deformation allows speculation concerning accretion-related processes that are known to be occurring at active convergent margins. The internally weak sandstone and hemipelagic mudstone deposited in a near-trench environment were unable to transmit stress and were subsequently offscraped and accreted along with minor amounts of pelagic deposits while more lithified and stronger pelagic ooze deposits were subducted. The lack of slivers of oceanic crust in the Greenwich slice suggests that the subducted pelagic layer was thick enough to act as a buffer zone between the lower inner trench slope and surface irregularities on the oceanic crust (McCarthy and others, 1982). In this sense accretion of the Greenwich slice sediments can be referred to as less "efficient" than coeval accretion of the Scottish Uplands (Leggett and Casey, 1983) and, therefore, somewhat similar to accretion in the Middle American Trench off Mexico (Moore and others, 1982).

Richmond slice

The Richmond slice consists of rocks of the Virginville Formation (Lash and Drake, 1984), a 565 m (1850 ft) thick sequence of siliciclastic rocks, micritic limestone, calcarenite, quartzose limestone, peloidal limestone, carbonate-clast conglomerate, and black shale and mudstone, that structurally overlies the Greenwich slice. The Virginville Formation was named for exposures in the southwest corner of the Kutztown quadrangle and the southern part of the Hamburg quadrangle. Work in progress, however, has traced the Richmond slice (Virginville Formation) as far west as the Palmyra quadrangle (approx. 70 km (40 mi)).

Virginville Formation

Rocks of the Virginville Formation range from Late Cambrian to Middle Ordovician in age and can be divided into three members; the Sacony (Late Cambrian), the Onyx Cave (Late Cambrian), and the Moselem (Middle Ordovician) (fig. 46). The Sacony Member is gradational into the overlying Onyx Cave Member and both units tectonically overlie the younger Moselem Member. The fault separating these units crops out at a few locations and slip-line analysis (Hansen, 1971) of folds formed in response to movement on the fault indicates that the Sacony-Onyx Cave sequence was thrust to the northwest over the younger Moselem Member.

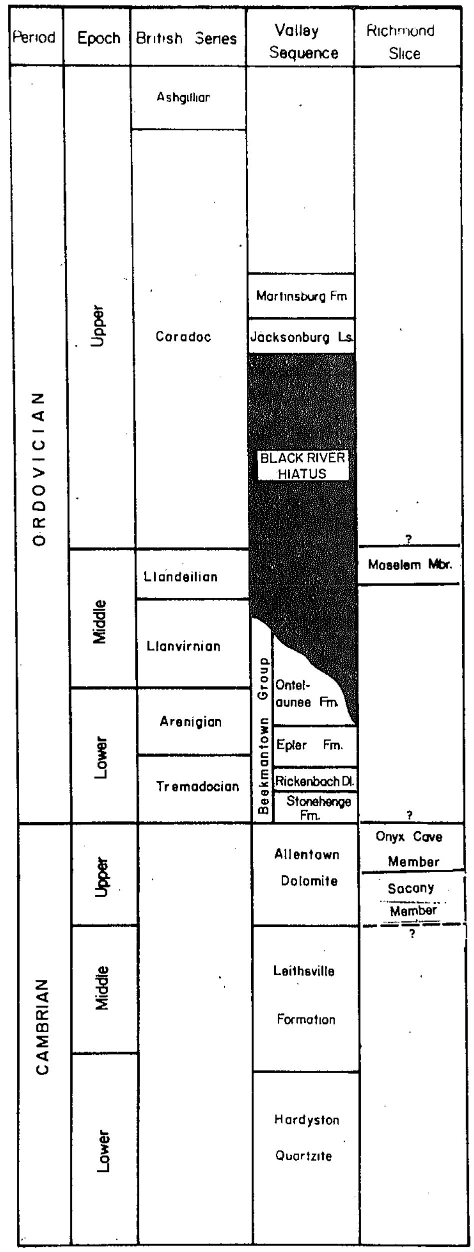


Figure 46. Age relations of the Virginville Formation/Richmond slice and Lehigh Valley sequence. Lehigh Valley sequence from Ross and others (1982).

Sacony Member.

The Sacony Member is named for clastic rocks exposed along Sacony Creek in the Kutztown quadrangle. It was originally described as "Martinsburg Shale" by Miller (1937) but recent investigations by Alterman (1972) assigned these rocks in the allochthon. The Sacony Member has a minimum thickness of about 245 m, an estimate based upon construction of geologic cross-sections.

Rocks of the Sacony are typically massive, grayish-olive-green (5GY 3/2) to pale-blue-green (5BG 7/2) and locally grayish-red (10R 4/2) micaceous quartzofeldspathic siltstone and sandstone. These rocks are interbedded with grayish-green (10GY 7/2) to pale-blue-green (5BG 7/2) micaceous shale and mudstone that weather to a distinctive rust color that is quite evident in soil and aids in mapping areas of poor exposure. Minor amounts of highly cleaved grayish-black (N2) to light-gray (N6) and light-green (5G 7/4) thinly-laminated, silicified shale are locally interbedded with the siltstone and sandstone.

Bedding is marked by lithologic and color contrasts, by planar surfaces having a micaceous sheen, and by discontinuous layers of quartz and dolomite. In some places "roll" structures may have resulted from bioturbation or viscoplastic deformation of bedding. Lenticular or ovoid fractures are also distinctive of the Sacony Member, and are believed by Alterman (1972) to be sedimentary in origin. Well defined primary structures, however, cannot be recognized.

Sandstone petrography of the Sacony Member.—Petrographic analysis of the Sacony Member was limited to samples with 25 percent or less matrix and with no obvious effects of post-depositional alteration. The average grain size of the majority of samples falls within the fine-sand size range (Table 5), thus reducing the influence of variations in grain size on detrital composition (Potter, 1978). Stained thin sections were examined employing techniques outlined by Dickinson (1970).

Table 5. Framework Modes of the Sacony Member Siltstone and Sandstone

	Q	F	L	Q_m	F	L_t	P/F
S1	54	46	0	53	46	1	.7
S2	60	40	0	59	40	1	.9
S3	49	51	0	47	51	2	.9
S4	53	47	0	52	47	1	.8
S5	60	39	1	59	39	2	.7
S6	50	50	0	48	50	2	.8
S7	51	49	0	49	49	2	.7
S8	48	52	0	46	52	2	.6
S9	50	49	1	48	49	3	.7
S10	55	45	0	52	45	3	.8
S11	58	42	0	56	42	2	.9
S12	53	47	0	53	47	0	.9
S13	49	49	2	45	49	6	.8
S14	57	42	1	56	42	2	.6
S15	56	41	3	52	41	7	.7
S16	59	40	1	56	40	4	.8
S17	52	43	5	50	43	7	.7
S18	48	51	1	47	51	2	.8
S19	56	41	3	56	41	3	.7
S20	50	46	4	48	46	6	.7

Average Grain Size: 2.75 phi

Average Grain Sorting: 1.35

Modal compositions presented as volumetric proportions of the following categories of grains: (l) stable quartzose grains, Q, including monocrystalline (Qm) and polycrystalline lithic (Qp) grains, (2) monocrystalline feldspar grains, F, and (3) unstable polycrystalline lithic fragments, L. Total lithics, Lt, include L plus Qp). P/F=plagioclase to total feldspar ratio. 400 to 500 grains counted per section.

Table 5 summarizes the modal composition of 20 samples of the Sacony Member. In general, the samples studied are poorly to moderately sorted and grain size within a particular sample varies from fine-silt to medium-sand. Monocrystalline quartz is the dominant framework component in roughly half the samples. Individual grains are very angular to subrounded without overgrowths. Most quartz grains have undulatory extinction. Feldspar grains are unaltered and are generally equant and angular. Albite and oligoclase, both twinned and untwinned, are roughly equal in abundance. Microcline grains are conspicuous by their gridiron twinning. Plagioclase feldspar greatly exceeds K-feldspar, but much of the original K-feldspar may have been either replaced by albite or selectively leached by intrastratal solution (Dickinson and others, 1982). Most plagioclase grains are angular and have well defined boundaries suggesting a detrital origin (Lajoie and others, 1974). The small amount of lithic fragments (Table 5) probably reflects the fine-grain size of the Sacony Member. Most lithic fragments are chert and quartzite, with lesser shale and mudstone fragments in the coarser-grained samples. Heavy minerals (averaging less than 3.5 percent) include biotite, euhedral zircon, apatite, tourmaline, hornblende, and opaques.

The high feldspar content of the Sacony implies a source area of moderate relief. The predominance of albite and oligoclase feldspar and the small but conspicuous amounts of microcline and orthoclase is consistent with erosion of granitic-type igneous and low to medium grade metamorphic rocks (Potter, 1978). Drake (1969) noted the presence of oligoclase-quartz gneiss, albite-oligoclase granite, and sodic granitic rocks in the Proterozoic rocks of the Reading Prong of Pennsylvania and New Jersey. It is likely that similar rocks exposed during Cambrian time served as a source for the Sacony sediment.

QFL (quartz-total feldspar-lithic fragments) and QmFLt (monocrystalline quartz-total feldspar-total polycrystalline lithic fragments) plots of Sacony detrital modes (fig. 58) illustrate that they fall within the continental block (uplifted basement complex) provenance of Dickinson and Suczek (1979) suggesting derivation of detritus from a continental block crossed by rift belts (Q65F26L5, Dickinson and Valloni, 1980). Ancient sands from similar tectonic settings in the Appalachian orogen include the Middle Proterozoic Mechum River Foramtion of central Virginia (Q76F26L3) and the Late Proterozoic Grandfather Mountain Formation of North Carolina (Q62F24L14, Schwab, 1974, 1977).

Onyx Cave Member.

The Onyx Cave Member, named for exposures at Onyx Cave in the Hamburg quadrangle, conformably overlies the Sacony Member and is dominated by four major facies: (1) thick-bedded calcarenite and quartzose limestone facies, (2) ribbon limestone facies, (3) conglomerate facies, and (4) laminated facies. The minimum thickness of the Onyx Cave Member, based on the construction of cross-sections, is 90 m (295 ft). The Onyx Cave Member crops out as isolated patches that are erosional remnants of a more extensive unit. Most of the major caves in the area have been formed along faults within the Onyx Cave Member.

Thick-bedded calcarenite and quartzose limestone facies.--Thick- to very thick-bedded, massive and structureless to parallel-laminated and long-wavelength ripple-laminated beds (fig. 47) of medium-light-gray (N6), poorly- to non-graded calcarenite and quartzose limestone are a major lithology of the Onyx Cave Member. Beds of this facies have a maximum thickness of 4 m (13 ft) and are locally amalgamated. Locally small lens-shaped beds cut underlying shale and argillaceous limestone. These beds (channels?) have a maximum depth of 18 cm (7 in) and widths of 1.5 m (5 ft).The bases of thicker beds are generally contain abundant rip-up clasts of underlying black shale and micritic limestone (fig. 48). These rocks are composed of recrystallized carbonate and well-rounded detrital quartz and minor K-feldspar grains (fig. 49).

The thick-bedded calcarenite and quartzose limestones are similar to Facies B and C of Mutti and Ricci-Lucchi (1975). The parallel horizontal and ripple-laminations are probably related to progressive "freezing" of traction grain carpets at the base of highly concentrated

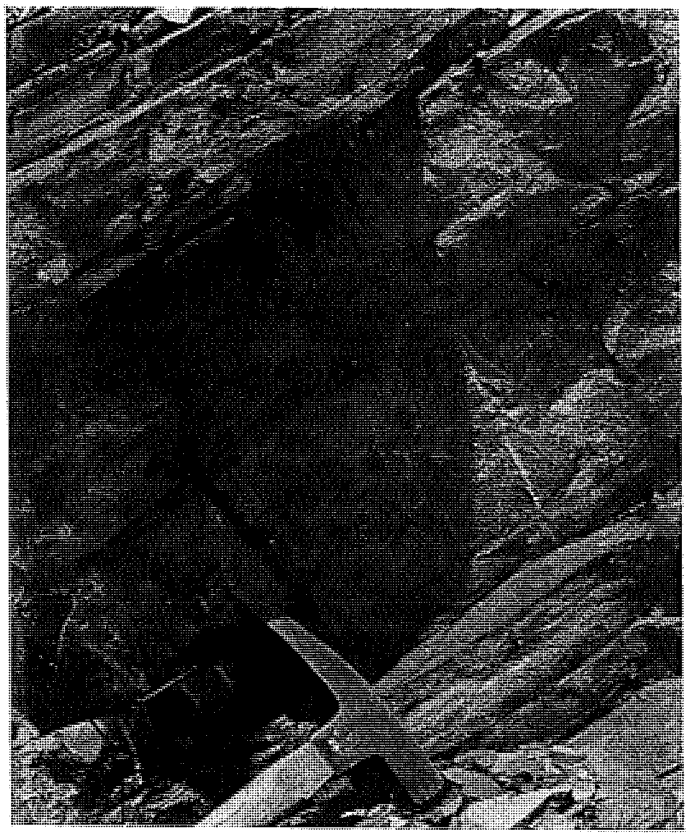


Figure 47. Laminated quartzose limestone bed. This bed may be a T_{ab} Bouma sequence as the lower part of the bed is graded.

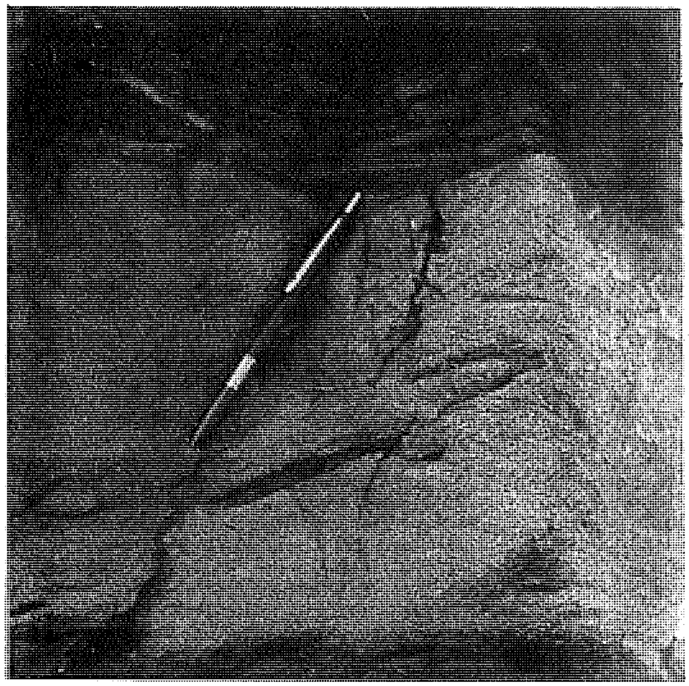
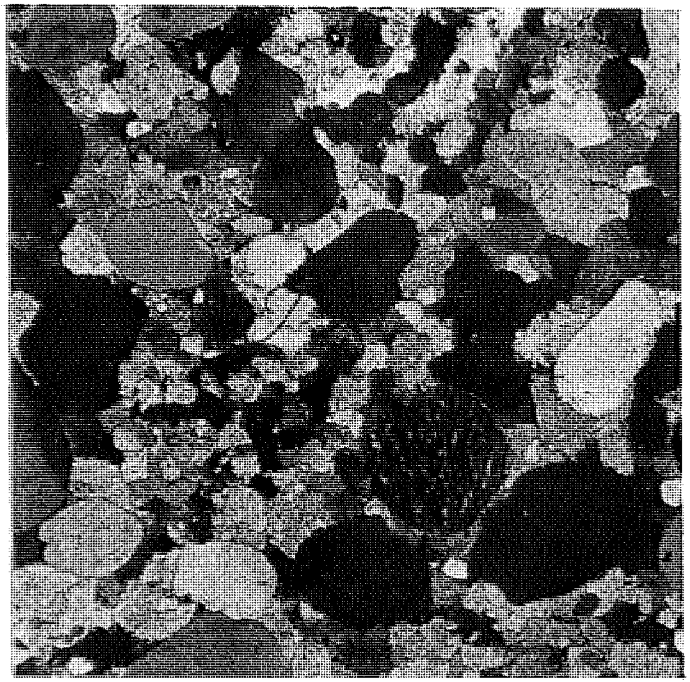
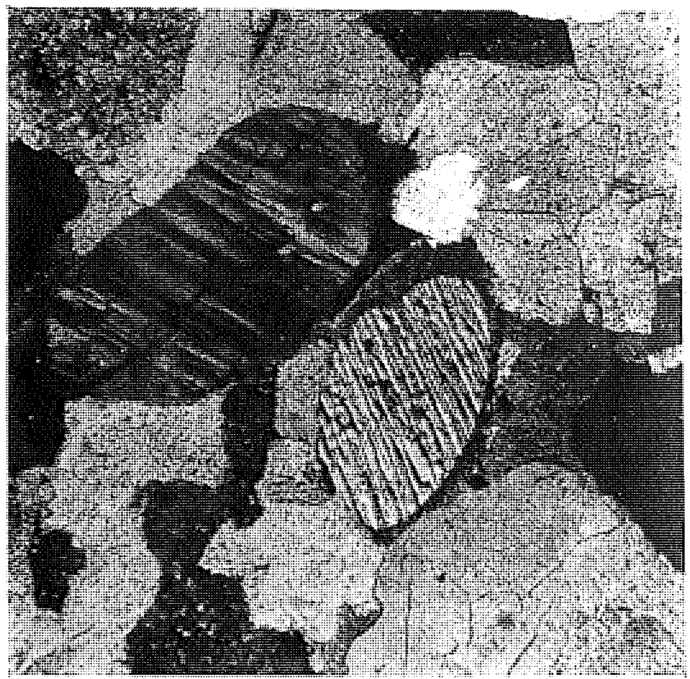


Figure 48.1 Black shale rip-up clast in a quartzose limestone bed.



A



B

Figure 49. Photomicrographs of quartzose limestone of the Onyx Cave Member (56x): A) well-rounded quartz and minor K-feldspar (K) in a recrystallized carbonate matrix that may include lithic fragments (L); B) well-rounded K-feldspar grains which may have overgrowths.

turbidity currents (Hiscott and Middleton, 1979; Lowe, 1982). The channels at the base of many of the beds and the occurrence some amalgamated beds supports this mechanism of sedimentation.

Figure 50. Ribbon limestone of the Onyx Cave Member.

Ribbon limestone facies.--The ribbon limestone facies of the Onyx Cave Member consists of thin- to medium-bedded, black (N1) to dark-gray (N3) lime mudstone to pelmicrite interbedded with thinner grayish-black- (N2) to medium-dark-gray- (N4) argillaceous lime mudstone that weathers to light-gray (N7) (fig. 50). The limestone is typically parallel- to cross-laminated and resembles Tcde and Tde partial Bouma sequences. In some cases beds that appear massive in the field are found to be laminated upon polishing and staining. Graded bedding, although present, is difficult to recognize in the field. Boudinage is common and is apparently due to differential compaction of interbedded mud and carbonate. Penecontemporaneous folds, although not common, are associated with some horizons of sedimentary boudinage (fig. 51). Beds of the ribbon limestone facies are similar to Facies D of Mutti and Ricci-Lucchi (1975) and are probably the result of deposition from dilute turbidity currents.

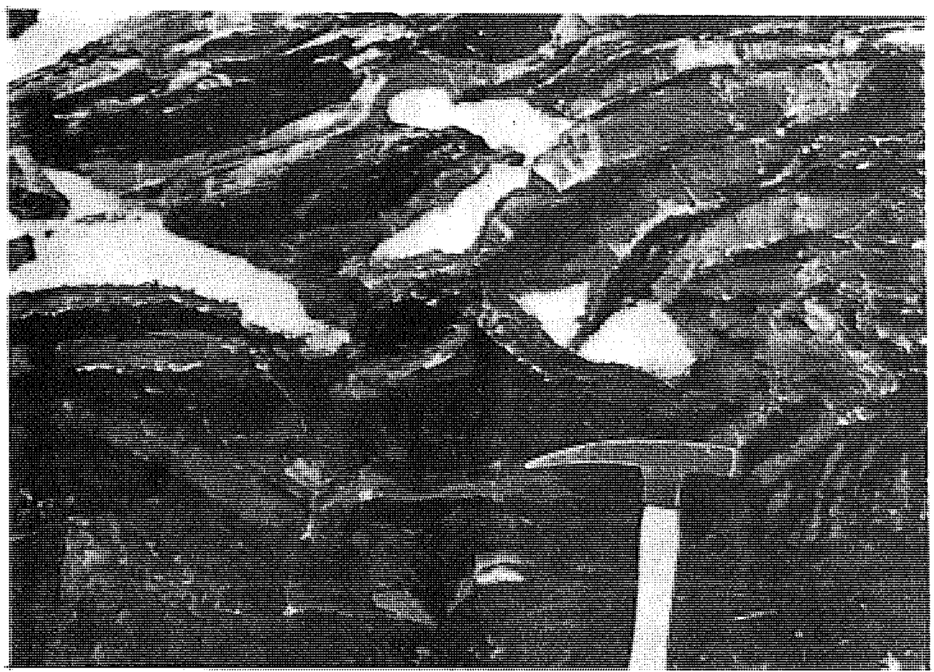


Figure 51. Slump folded ribbon limestone.

Conglomerate Facies.--Polymictic carbonate-clast conglomerate is a major lithology of the Onyx Cave Member. Individual beds range from 0.3 to 2 m (1 to 7 ft) in thickness and are laterally continuous across an outcrop width of 12 m (40 ft). Two major types of conglomerates include: (1) coarse-grained, sandy to gravelly matrix conglomerates, and (2) mud-matrix conglomerates. The latter constitutes only 15 percent of the conglomerate beds studied. They are significantly thinner than the coarse-matrix conglomerates (average thickness 0.3 m) and are characterized by a significantly higher clast:matrix ratio (average of 8:1). These beds are associated with rocks of the ribbon limestone facies and consists of laminated ribbon limestone clasts in a black mud matrix (fig. 52). Clasts are generally imbricated and are oriented sub-parallel to bedding, although in a few exposures they are perpendicular to bedding. The proximity of these beds to obvious source rocks suggests that they resulted from debris flow resedimentation of slumped and disrupted turbiditic ribbon limestone beds and hemipelagic mud (Naylor, 1981).

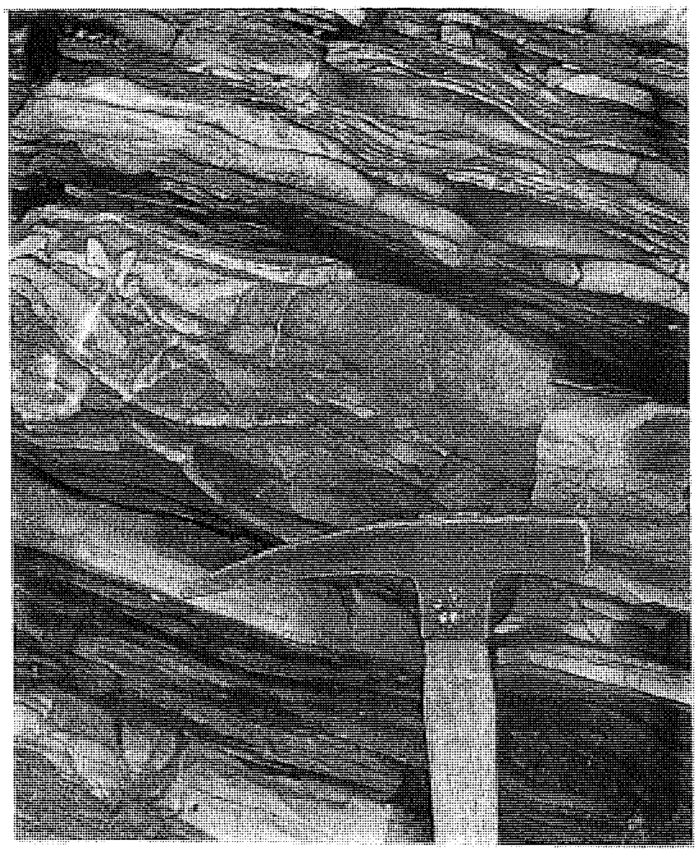


Figure 52. Mud-matrix-rich conglomerate.

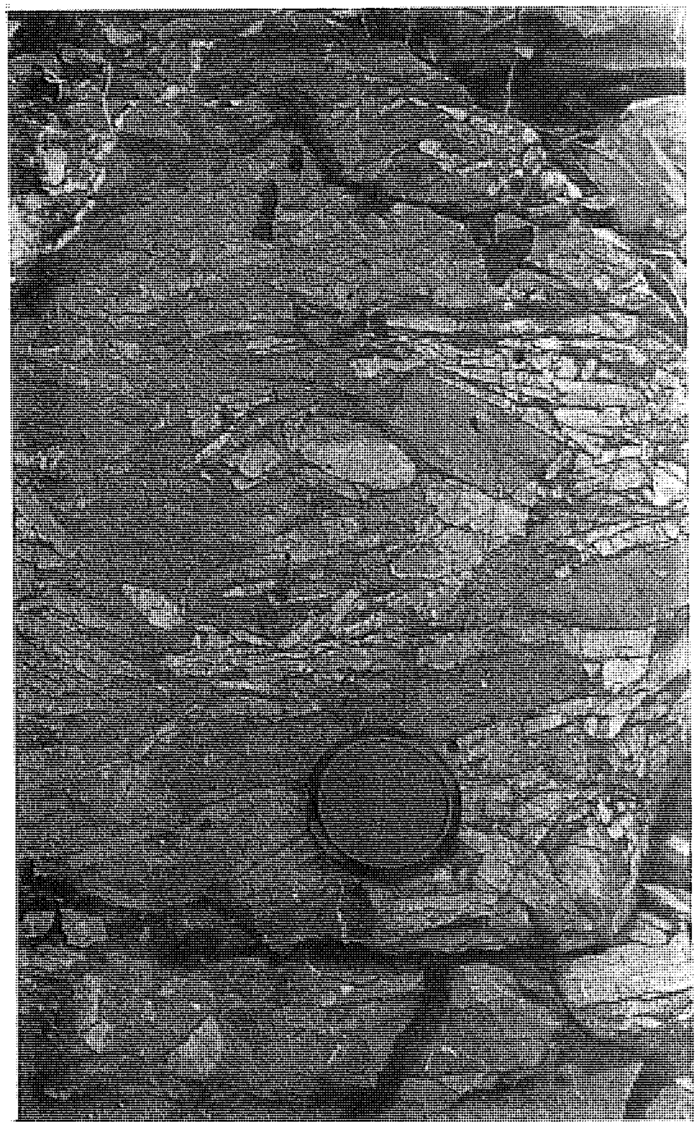


Figure 53. Poorly-sorted, coarse-matrix conglomerate.

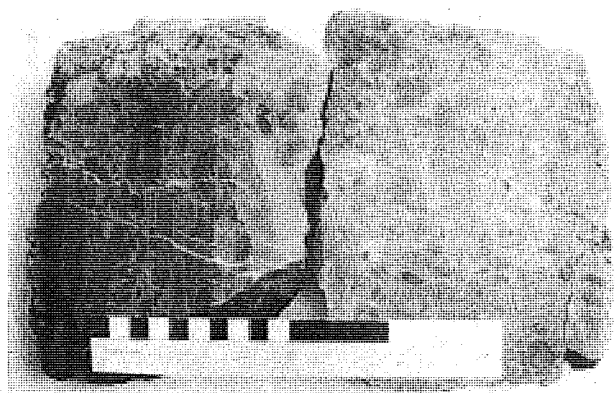


Figure 54. Polished slab of lower 30 cm of a coarse-matrix conglomerate bed. Note planar clast fabric at the bottom of the bed.

The coarse-grain matrix conglomerates are significantly thicker than the mud matrix conglomerates (average thickness 1.5 m) and are characterized by a lower clast:matrix ratio (3-1:1). The clasts are composed of micrite, calcarenite, grainstone, peloidal wackestone, conglomerate, and mudstone, all of which are common in the Onyx Cave Member. The sandy to gravelly matrix is poorly sorted and includes 20-40 percent well rounded and sorted quartz grains, 10-20 percent peloids, and 10-15 percent mud in addition to coarse-grained, locally recrystallized carbonate and sand- to gravel-size clasts.

Clasts are typically tabular, poorly sorted, and range up to 35 cm (13.8 in) in length (fig. 53). They generally lie parallel to bedding, particularly near the coarser tops and bottoms of beds, whereas the areas in between are generally massive (fig. 54). Elsewhere the clasts are at high angles to bedding.

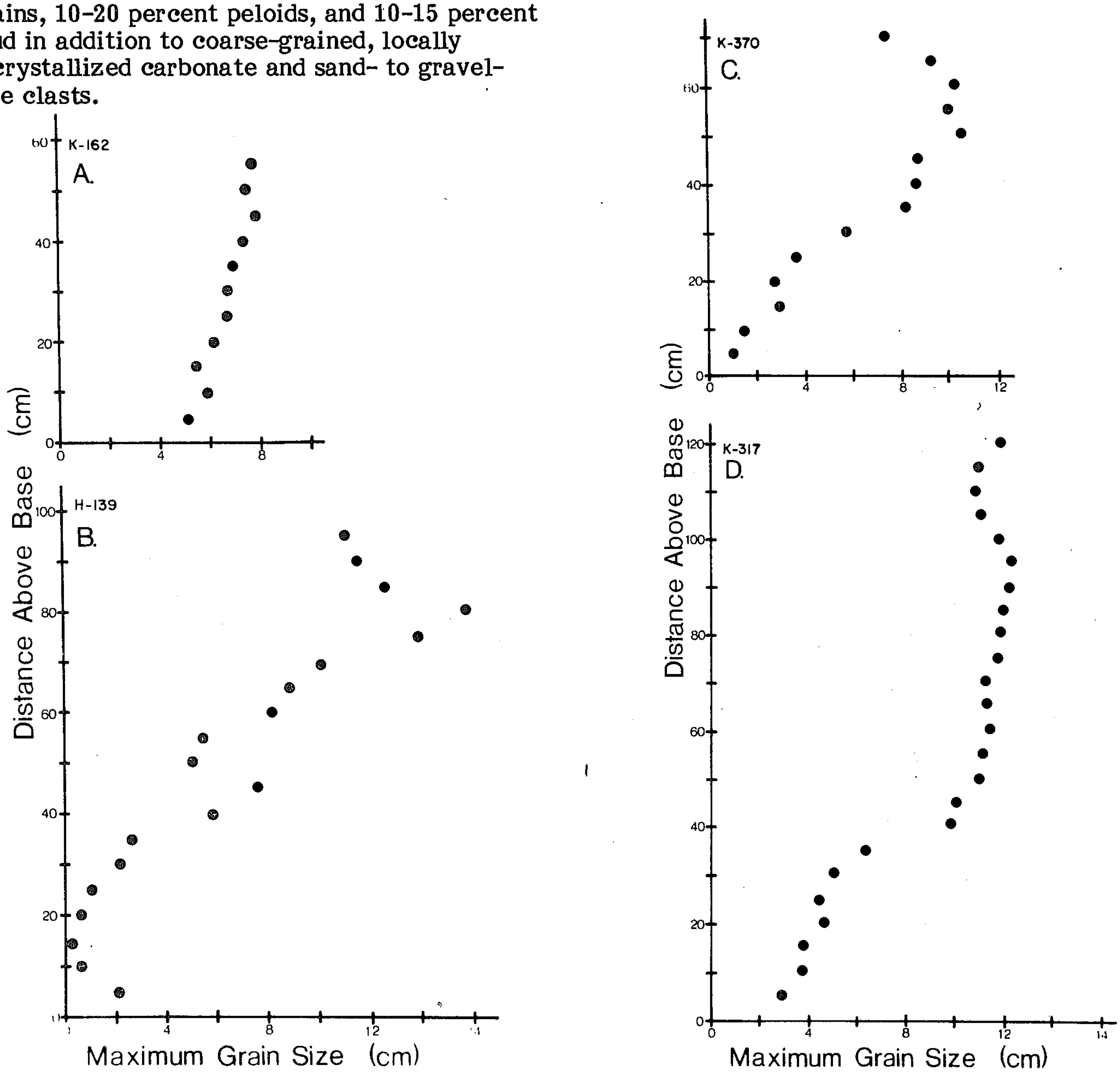


Figure 55. Plots showing vertical size variation of maximum grain size in four coarse-matrix conglomerate beds:

A. inverse grading

B. normal grading at the top and bottom, inverse grading in middle

C. normal grading at top

D. inverse grading

The conglomerates are not laminated except for thin parallel to cross-laminated grainstone units on top of about 20 percent of the beds. Most conglomerates are normally graded on top and less commonly on the bottom (fig. 55B, C). The tops are generally overlain by thinly laminated grainstone caps. Figure 56 is a composite diagram of the major features of the sandy- and gravelly-matrix conglomerates.

The coarse-matrix conglomerates are the result of density-modified grain flow (Lash, 1984b), a little described mechanism intermediate between true grain flow (Lowe, 1976) and debris flow (Middleton and Hampton, 1976), the main variable being the amount of mud in the flow. Mechanisms of grain support in density-modified grain flow include buoyancy forces provided by dense interstitial fluid, dispersive pressure, and matrix strength supplied by small amounts of clay. Evidence for a density-modified grain flow origin of the coarse-grain matrix conglomerates includes inverse grading (fig. 55A, D), tabular clasts oriented at high angles to bedding, the presence of a poorly-sorted, gravelly matrix, and the small amount of matrix mud. Turbulence on top of the flow is indicated by normal grading of clasts and parallel- and cross-laminated grainstone at the top of some beds which may be related to backward shearing of displaced sediment and water in front of the flow (Krause

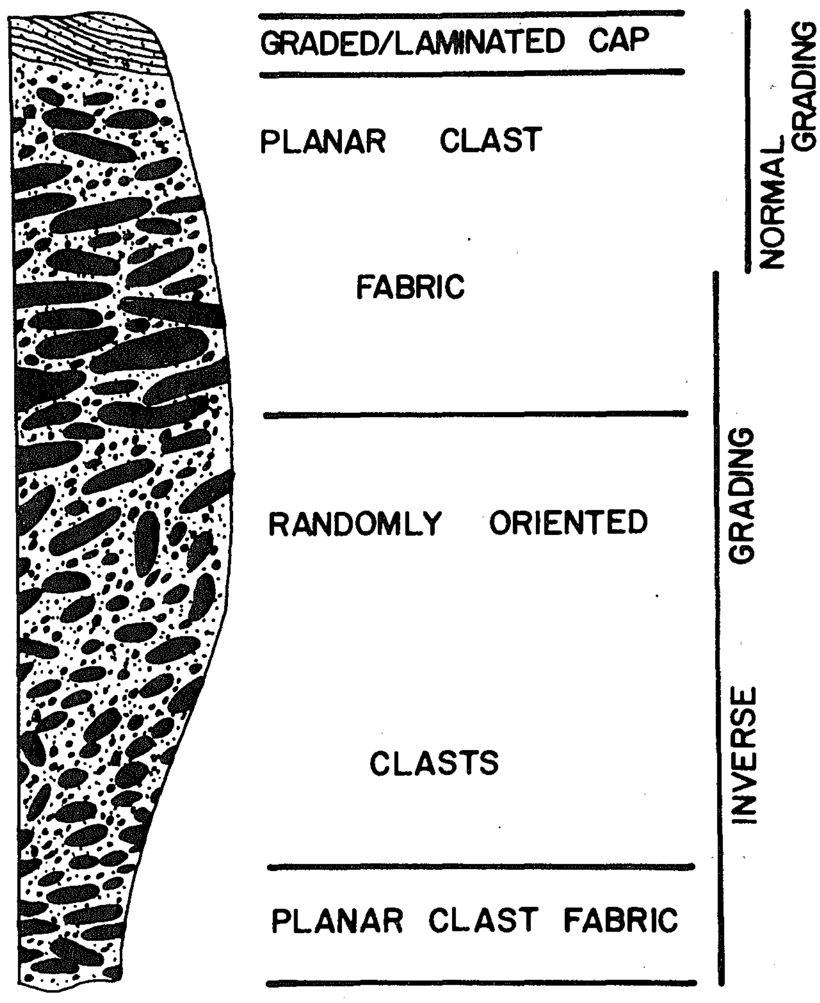


Figure 56. Composite diagram illustrating the major features of the thick-bedded, sand to gravel matrix, limestone clast conglomerates of the Onyx Cave Member. Bed thickness ranges from 0.6 to 2 m.

and Oldershaw, 1979) or to pore fluid expulsion during movement of the sediment mass (M. Hampton, written commun., 1983). The coarse-grain conglomerates of the Onyx Cave Member are similar to modern density-modified grain flow deposits from the base of the Bahama Escarpment (Mullins and Van Buren, 1979).

The mud-matrix and coarse-grain conglomerates of the Onyx Cave Member are similar to Facies F (chaotic deposits transported by slumping or sliding) and Facies A of the classification of Mutti and Ricci-Lucchi (1975).

Laminated Facies.--Thinly-laminated to laminated black (N1) to dark-gray (N3) and dark-yellowish-orange (10YR 6/6) to pale-yellowish-orange (10YR 8/6) dolosiltite and calcisiltite occur throughout the Onyx Cave Member. These sequences commonly show evidence of syndepositional deformation in the form of small folds and pinch-and-swell structures. Bedding is continuous to discontinuous and commonly ripple-laminated. The mechanism of deposition of these rocks is not fully understood but may be, in part, the result of bottom-following or contour-current reworking of hemipelagic and turbiditic sediment (Stow, 1979). This is supported by a absence of graded beds or load-induced features suggestive of turbiditic sedimentation (Piper, 1978).

The conformable contact between the Sacony and overlying Onyx Cave Member can be seen in an exposure approximately 1.5 km (1 mi) southeast of Onyx Cave in the Hamburg quadrangle. Structures in the Onyx Cave at this locality indicate that the sequence is right-side-up. Green siltstone of the Sacony is overlain by thinly-laminated black shale and orange dolostone and calcisiltite. This in turn grades up into varying proportions of ribbon limestone, carbonate-clast conglomerate, calcisiltite, calcarenite, and thick-bedded quartzose limestone (fig. 59).

Rocks of the Sacony Member have yielded no fossils. However, the Onyx Cave Member has yielded Late Cambrian conodonts (J. E. Repetski, personal communication, 1980, 1981). These are the first Cambrian fossils collected from the klippe and indicate that the Sacony Member is at least as old as Late Cambrian.

Moselem Member.

The Moselem Member is a 230 m (755 ft) thick sequence of cleaved black and green mudstone and shale and lesser carbonate rock which are best exposed along Maiden Creek in the southern half of the Hamburg quadrangle. The Moselem is tectonically overlain by the Sacony and Onyx Cave Members and differs from the Onyx Cave in that it contains much more shale and mudstone and significantly less limestone.

The predominant carbonate lithology of the Moselem Member is very thin- to thin-bedded lime mudstone and micrite interbedded with black (N1) shale. Limestone beds generally have parallel and ripple laminations. Moreover, polished and stained slabs of beds that appear massive in the field generally reveal the presence of laminations. Ribbon limestone beds are typically found in packages up to 3 m (9.8 ft) thick that are intercalated with black shale. The ribbon limestone beds are similar to Tcde and Tde partial Bouma sequences and apparently were deposited by dilute turbidity currents.

Penecontemporaneous deformation is common to the ribbon limestone. Soft-sediment folding is indicated by draping of undeformed sediment over folds and by the association of boudinage and folds (fig. 57). The limestone boudins show a wide range of cross-sectional shapes that reflect competence contrasts of the limestone and shale at the time of deformation (Ramsay, 1967). In general, the transition from gently tapering boudins to more blocky, blunt-ended boudins reflects a transition from limestone deformed soon after deposition to limestone deformed after partial lithification. Boudin axes are commonly parallel to slump fold axes. However, there are areas where there is no detectable preferred orientation of boudin axes, indicating that extension occurred easily in all directions in the plane of layering. This style of deformation, referred to as axially symmetric extension by Cowan (1982a), requires a laterally unconfined environment of deformation. Cowan (1982a) suggested that

Figure 57. Slump folds and sedimentary boudinage in ribbon limestone beds of the Moselem Member. Outcrop is found at STOP 2.

axially symmetric deformation may be typical of vertical shortening and related horizontal elongation due to gravitational collapse and spreading at shallow levels on a slope. This is certainly consistent with other soft sediment deformation characteristics of these rocks.

Sequences of thinly-laminated black (N1) to dark-gray (N3) graphitic shale and dark-yellowish-orange (10YR 6/6) to pale-yellowish-orange (10Y 8/6) well sorted dolosiltite and calcisiltite are commonly associated with the ribbon limestone. These sequences are, in some places, highly contorted. Beds are typically discontinuous and cross-laminated. The irregular bedding, good sorting, and the lack of evidence of deposition from turbidity currents point to reworking of calcisiltite and dolosiltite by contour or bottom currents (Stow, 1979).

Carbonate-clast conglomerate beds are rare in the Moselem Member. Those that are present are significantly thinner than those in the Onyx Cave Member. Clast lithologies include laminated micrite, pelmicrite, calcisiltite, and minor but conspicuous black chert (flint). These beds probably formed by submarine sliding and disruption of ribbon limestone and associated hemipelagic mud, followed by resedimentation of this material in mud-matrix debris flows (Naylor, 1981) and coarse-grained density-modified grain flows.

Mudstone and shale, locally well cleaved, are the dominant lithologies of the Moselem Member. The argillaceous rocks are quite variable and include thin- to thick-bedded graphitic grayish-black (N2) mudstone and shale interbedded with dusky-brown (5YR 2/2) mudstone and very thin-bedded ribbon limestone. Pyrite lenses up to 18 cm (7.1 in) long are found locally. Minor thin- to thick-bedded grayish-black (N2) dolostone interbedded with medium dark-gray (N4) medium-gray (N5) mudstone and light-gray (N7) siliceous shale are locally associated with these rocks..

In some areas the dominant shale is dark-gray (N3) to dark-greenish-gray (5G 4/1) and siliceous and contains lenses and discontinuous beds of silicified mudstone. Thin-bedded, very light-gray (N8) quartzite is interbedded locally with the mudstone and shale.

Originally Lash (1980) suggested that the Moselem Member was a distal facies correlative of the Onyx Cave Member. This is now untenable because conodonts from the Moselem Member are Middle Ordovician in age (J. E. Repetski, personal communication, 1982), significantly younger than the tectonically overlying Sacony-Onyx Cave sequence.

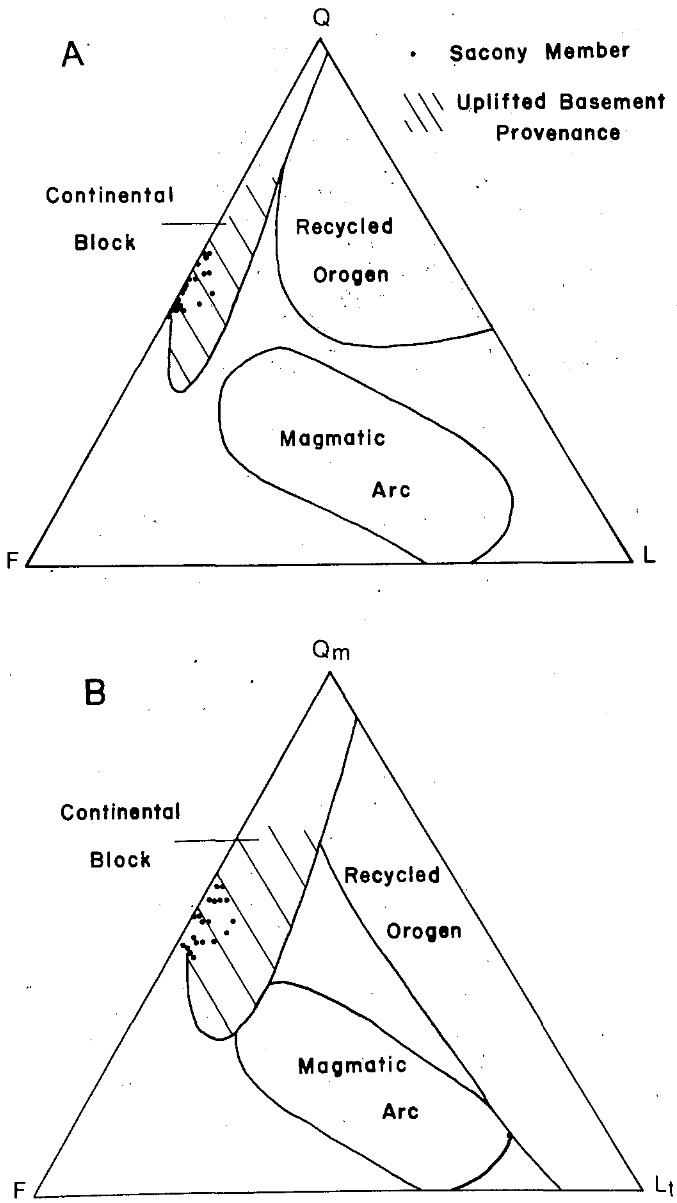


Figure 58. Triangular plots showing framework modes of Sacony siltstone and sandstone samples: (a) QFL, (b) QmFLt.

Depositional model for the Virginville Formation.

The transition between the Sacony Member and the Onyx Cave Member is relatively abrupt and is characterized by a change from quartzofeldspathic sandstone (fig. 58) to thinly-laminted black shale and calcisiltite. These rocks are overlain by the thick conglomerate and calcarenite beds of the Onyx Cave (fig. 59). The lack of wave-produced structures and the moderate to poor sorting of the Sacony

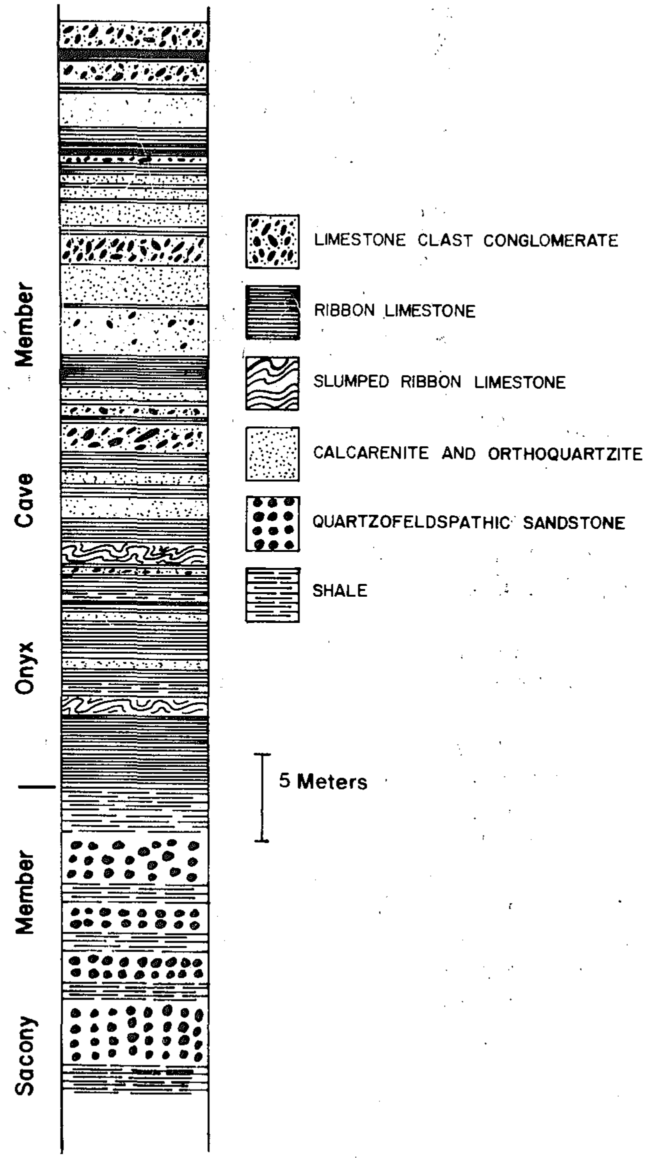


Figure 59. Stratigraphic column, based on composite sections, illustrating lithologic characteristics of the Sacony-Onyx Cave contact. (Only the lower and upper parts of the Onyx Cave and Sacony Members, respectively.)

quartzofeldspathic detritus is suggestive of deposition below wave base. The lithologic characteristics of the Onyx Cave Member discussed earlier indicate that the thick-bedded calcarenite, limestone-clast conglomerates, and ribbon limestones were deposited by sediment gravity flow mechanisms which include turbidity currents, density-modified grain flow, and debris flow. Detailed vertical facies analysis of the few continuous sections of these rocks illustrates the applicability of clastic deep sea fan models to the Onyx Cave Member. A particularly interesting section of part of the Onyx Cave Member is exposed along the Shartlesville-Bernville Road. Analysis of this section indicates the presence of a thinning- and fining-upward megasequence (see Stop 1 description). It contains Facies A and B conglomerates and quartzose limestones in its lower part, some of which are amalgamated. These have erosive bottom contacts and rip-up clasts of black shale. They grade into Facies B and C quartzose limestones with such turbidite features as graded bedding and the basal divisions (a, b) of the Bouma sequence. The c, d, and e parts of the Bouma sequence, where present, are composed of calcisiltite and black shale. These beds are overlain abruptly by parallel horizontal and ripple laminated ribbon limestone (Facies D) and black shale beds. This cyclic arrangement is typical of channelized sediments (Ricci-Lucchi, 1975; Ruiz-Ortiz, 1983). The lower part of the cycle consisting of Facies A and B conglomerate and quartzose limestone beds is the channel facies. In particular, the very thick, massive conglomerates probably are the channel "plug". The Facies B and C beds overlying the "plug" are probably interchannel sediments deposited by channel spillover. Finally, the ribbon limestones with climbing ripples and ripple laminations at the top of the section are probably levee and overbank deposits related to lateral migration of an adjacent channel (fig. 60). These beds are locally slumped and folded and contain thin, locally derived conglomerates that are probably related to slumping of the channel margin and levee (Pickering, 1982). These deposits are similar to levee deposits of clastic channels (Normark, 1970, 1974).

The erosive and coarse-grained nature of the plug deposits (Facies A and B beds) is difficult to reconcile with the abundance of levee deposits. For example, a great dominance of channelized deposits over levee deposits in a particular submarine fan is typical of low efficiency suprafan systems in which the sediment supplied to the fan is characterized by a very high sand:mud ratio. The presence of channel as well as levee deposits in the Onyx Cave Member suggests deposition from thick turbidity currents with approximately equal proportions of sand and mud. This mix of sediment would contribute to the erosion of the channel whereas the fine-grained material would provide abundant overbank material. This is consistent with recent concepts that channelized turbidity currents may rise hundreds of meters above levee crests (Piper and Normark, 1983). Piper and Normark (1983) have suggested that flow stripping of the upper parts of sandy turbidity currents leads to increased sedimentation on levee and interchannel areas and rapid deceleration of the lower part of the flow and possible plugging of the channel. They maintain that both grain-size and sand:mud ratio of turbidity currents are the dominant controlling factors of this process.

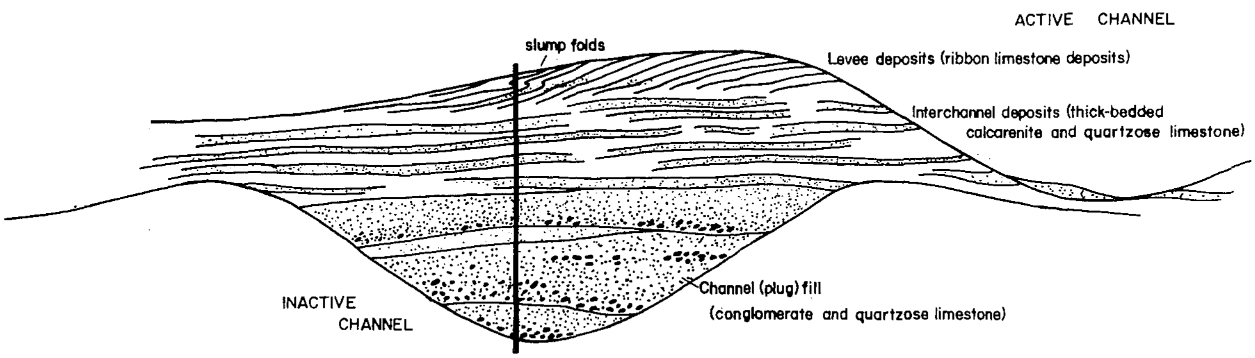


Figure 60. Diagram illustrating plugging and avulsion of Onyx Cave submarine channels. Refer to text for discussion.

The sedimentology of the Onyx Cave Member differs from modern carbonate-turbidite systems such as in Bahamian interplatform-basins (Bornhold and Pilkey, 1971; Schlager and Chermak, 1979; Pilkey and others, 1980; Crevello and Schlager, 1980) which are characterized by lobe-shaped, non-channelized turbidite deposits derived from an active linear source. In contrast, the sedimentology of the Onyx Cave Member is similar to that of recent and ancient clastic submarine fans. The Onyx Cave was probably part of a small fan fed by a submarine canyon, a point source, and deposited as a system of laterally migrating channels and associated interchannel levee deposits. The well-developed levee and channel deposits suggests that the sediment consisted of a mixture of sand and mud resulting in the development of a submarine fan intermediate between the "efficient" and "inefficient" end member types (Mutti, 1979; Ghibaudo, 1980; Reynolds and others, 1983). It is likely that sediment of the Onyx Cave Member was deposited in a small fan similar to those of the California continental borderland. The dispersal system of the Onyx Cave Member is not unique among ancient carbonate shelf edge sequences. Similar systems have been described from the Meteritzia Formation of the Othris Zone (Price, 1977) and the Loma del Toril Formation of the Betic Cordillera, southern Spain (Ruiz-Ortiz, 1983). Sedimentation of the channelized deposits of the Onyx Cave Member records progradation of the North American carbonate shelf edge in Late Cambrian time.

Age	Eastern Pennsylvania - Western New Jersey	Richmond Slice, Hamburg Klippe	
Lower Ordovician	Beekmantown Group	?	
Upper Cambrian	Allentown Dolomite	Onyx Cave Member	Virginville Formation
		?	
Middle Cambrian	Leithsville Formation	Sacony Member	
		?	
Lower Cambrian	Hardyston Quartzite		

Figure 61. Stratigraphic relations of the eastern Pennsylvania-western New Jersey Cambrian sequence (from Drake, 1969) and the Onyx Cave and Sacony Members of the Virginville Formation (from Lash, 1980).

The predominance of hemipelagic black shale in the Moselem Member, in combination with the regional tectonic scenario to be discussed later is suggestive of sedimentation in a subsiding basin under the influence of an expanding oxygen minimum zone (Leggett, 1980b). Abundant soft sediment deformation in the rare but conspicuous limestone-clast conglomerate beds suggests that the subsiding basin may have been subjected to tectonic movement related to subsidence. Equal area plots and slip-line analyses (Hansen, 1971) of slump fold axes indicate northwest overturning of folds, which is consistent with movement on a slope inclined to the northwest.

Sedimentologic evolution of the Richmond slice

The Sacony Member is overlain by carbonate deposits of the Onyx Cave Member suggesting rapid progradation of a passive carbonate shelf edge in Late Cambrian time (Lash, 1980,; Lash and Drake, 1984). The transition from quartzofeldspathic to carbonate shelf edge sedimentation is abrupt and appears to record an increase in water depth. The significance of the relation of the Sacony Member to the Onyx Cave is best understood when the sedimentation of coeval shelf deposits of the authochthonous (parautochthonous) Pennsylvania Great Valley sequence is considered.

The Onyx Cave Member is at least partly correlative with the Allentown Dolomite and perhaps the Leithsville Formation (fig. 61). The Leithsville Formation, an arenaceous dolomite, and the Allentown Dolomite were both deposited as part of the early Paleozoic eastern carbonate platform (Drake, 1969). These rocks overlie the Hardyston Quartzite, a basal transgressive quartz arenite. The Hardyston consists of feldspathic sandstone and conglomerate (apparently derived from underlying Proterozoic crystalline rocks) that grade up into well- sorted, orthoquartzite and silty shale (Aaron, 1969). The transition from the Hardyston to the Leithsville and Allentown indicates lowering of the crystalline source area in response to tectonic stabilization and subsidence of this part of the rifted craton in Early Cambrian time. This was significantly earlier than source area lowering during the transition from the Sacony to the Onyx Cave.

Palinspastic reconstruction of the Richmond slice relative to rocks of the North American carbonate platform (Lash, 1980; Lash and Drake, 1984) suggests that the Sacony Member was deposited east of the stable platform. A critical consideration of this restoration is whether compositionally immature quartzofeldspathic sediment could have been transported across the stable, low-relief, tropical (Friedman, 1982) carbonate platform and deposited to the southeast. Unfortunately, the lack of current indicators in the Sacony deposits precludes detailed sediment dispersal analysis. The fact that the Sacony is poorly to moderately sorted (Table 5) and composed of angular quartz and feldspar grains even in the coarser samples is not consistent with eolian transport (Freeman and Visher, 1975) across the platform from the northwest. On the other hand, terrigenous detritus in calcarenite and orthoquartzite beds of the overlying Onyx Cave Member is compositionally mature (quartz arenite), well to very well sorted, and composed of well-rounded and sometimes frosted grains (Filock and Lash, 1984). These deposits, unlike those of the Sacony Member, were clearly derived from a shoreface or dune area and swept out to the shelf edge as the carbonate platform prograded to the southeast. It is difficult to imagine how compositionally immature clastic sediment of the Sacony could survive the prolonged period of weathering during transportation across the low-relief, tropical Middle Cambrian platform. Basu (1976) noted that plagioclase is not stable in a humid climate which prevailed during shelf sedimentation. The abundance of angular plagioclase grains in the Sacony is clearly not

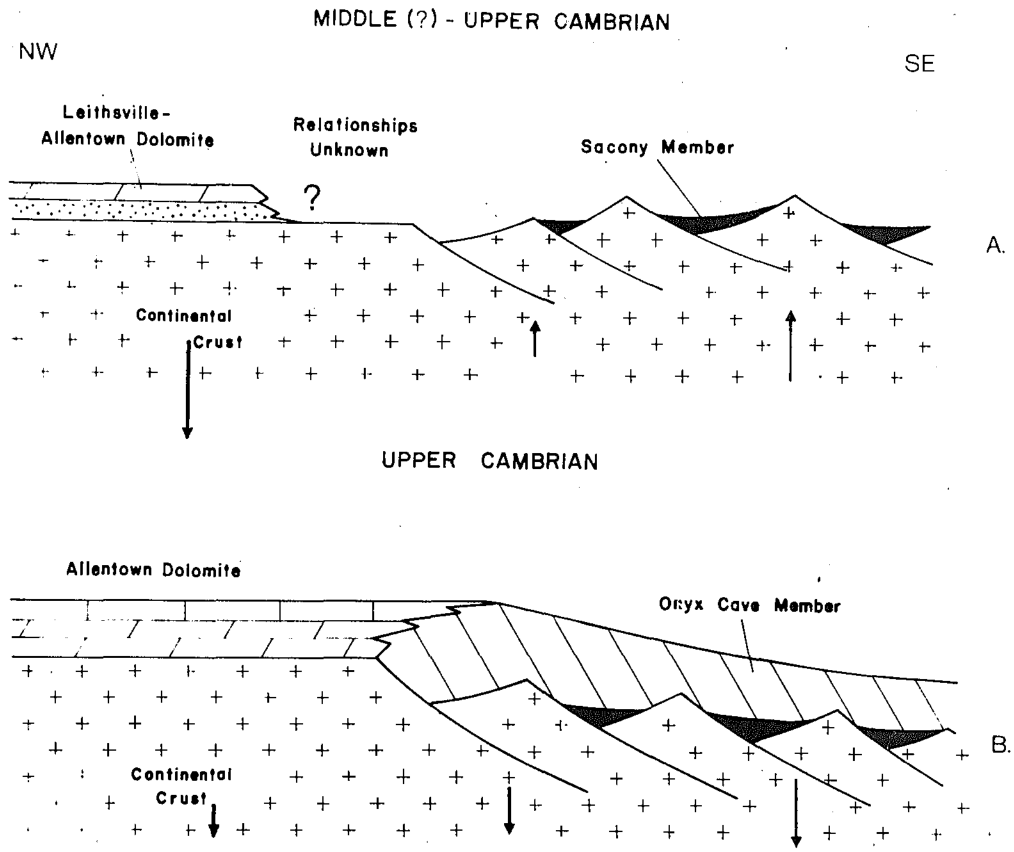


Figure 62. Conceptual evolution of the early Paleozoic continental margin of the central Appalachian orogen. Length of arrows indicates relative rates of uplift or subsidence, (a) Middle(?) to Late Cambrian, (b) Late Cambrian.

consistent with tranport across the Cambrian shelf.

It can be argued that the Sacony source area was located to the southeast of the carbonate platform and was characterized by horst blocks of crystalline rocks (fig. 62A). In this sense the Sacony source area may have been an outer high similar to those described by Scheupbach and Vail (1980) from seismic profiles of passive margins. Increased relief to the southeast of the subsiding platform in Middle Cambrian time may have been because this area was nearer the Late Proterozoic-Early Cambrian rift axis and its associated thermal bulge. Vierbuchen and others (1983) described a model that addresses the formation of outer highs of passive margins such as that proposed as the Sacony source area. In the early stages of rifting, subsidence dominates over uplift as a result of crustal thinning. Approximately 35 m.y. after initiation of rifting, however, a rapid acceleration in asthenospheric upwelling results in uplift toward the rift axis (fig. 62A; Sacony source area) whereas areas located farther from the outer high (fig. 62A; Hardyston-Leithsville-Allentown depocenter) continue to subside (Vierbuchen and others, 1983, fig. 4B). Vierbuchen and others (1980) suggested that uplift of outer highs generally ranges from 0.5 to 1.0 km (0.3 to 0.6 mi) and occurs over a 10 to 30 km (6 to 19 mi) width. The proposed origin of the Sacony sediments being derived from an elevated crystalline source area outboard of the Cambrian carbonate platform is consistent with the thermal-mechanical model presented here.

It is impossible to determine when sedimentation of the Sacony was initiated because the lower contact is not exposed due to faulting. The fact that the Sacony underlies the Onyx Cave suggests only that uplift of the Sacony source area was probably during the Middle Cambrian or perhaps earlier. The fine grain size of the Sacony and the lack of conglomerates typical of rifting is suggestive of deposition far from the source and indicates that some of the youngest deposits related to uplift are preserved. The initial age of rifting in the central Appalachians is difficult to determine. Published radiometric ages of rift-related volcanic rocks in the Appalachians range from 820 to 565 m.y. (Williams and Stevens, 1974). If rifting in the central Appalachians was initiated during the latter age, the 35 m.y. post-rift development time of the outer high would place the time of rifting sometime in the Early Cambrian.

The change from quartzofeldspathic to black shale sedimentation in Late Cambrian time (fig. 62B) indicates rapid subsidence in which (1) the quartzofeldspathic source was no longer contributing sediment and (2) basin subsidence exceeded rate of sedimentation. This may record sediment-load-enhanced thermal subsidence that McKenzie (1978) and Dewey (1982, fig. 17) maintain occurs approximately 180 m.y. after rifting. The fact that the subsidence rate exceeded that of sedimentation suggests that thermal contraction played a greater role than sediment loading. This was probably related to contraction of the upwelled asthenosphere that lead to initial development of the outer high.

Finally, the presence of long-standing horsts of Precambrian basement at the edge of the continental shelf may have been responsible for the Late Cambrian intrashelf basins in the Appalachian orogen described by Markello and Read (1982). Inasmuch as intrashelf basins on rimmed shelves of passive margins have been described from sequences of various ages (Ginsburg and James, 1974; Eliuk, 1978; Markello and Read, 1982), the nature of the elevated outer rim is not always understood. Eliuk (1978) noted that the Mesozoic shelf edge of eastern Canada is localized above basement horst blocks. Markello and Read (1982) suggeted that rapid subsidence of the platform area relative to the rim may control formation of intrashelf basins. Indeed, the thermal-mechanical model of Vierbuchen and others (1983) predicts localized sedimentation in continually subsiding depocenters on the craton-side of the rising outer high (Vierbuchen and others, 1983, fig. 4B). Derivation of Sacony Member sediment from an uplifted outer high lends support to the structural control of rimmed intrashelf basins at least in the central Appalachians.

The tectonic significance of the Moselem Member must be considered independent of the older Sacony-Onyx Cave sequence. Four important points should be considered:

1. The Middle Ordovician age of the Moselem Member and the fact that it was deposited during the Blackriveran hiatus (period of non-deposition/erosion on the North American carbonate shelf sequence; fig. 46);
2. The presence of conspicuous black chert or flint clasts within conglomerate beds of the Moselem Member;
3. Slump folds that show predominant movement directions to the northwest;
4. Crustal convergence in the central Appalachians concomitant with Middle Ordovician sedimentation of the Moselem Member as indicated by the Greenwich slice.

The Moselem Member was deposited at a time when the North American platform Beekmantown Group was subaerially exposed (fig. 63). Profound erosion of the shelf resulted in the Blackriveran hiatus and has been related to passage of the North American craton across the outer swell or peripheral bulge (Lash, 1980; Lash and Drake, 1984; Jacobi, 1981; Shanmugam and Lash, 1982). The presence of black flint

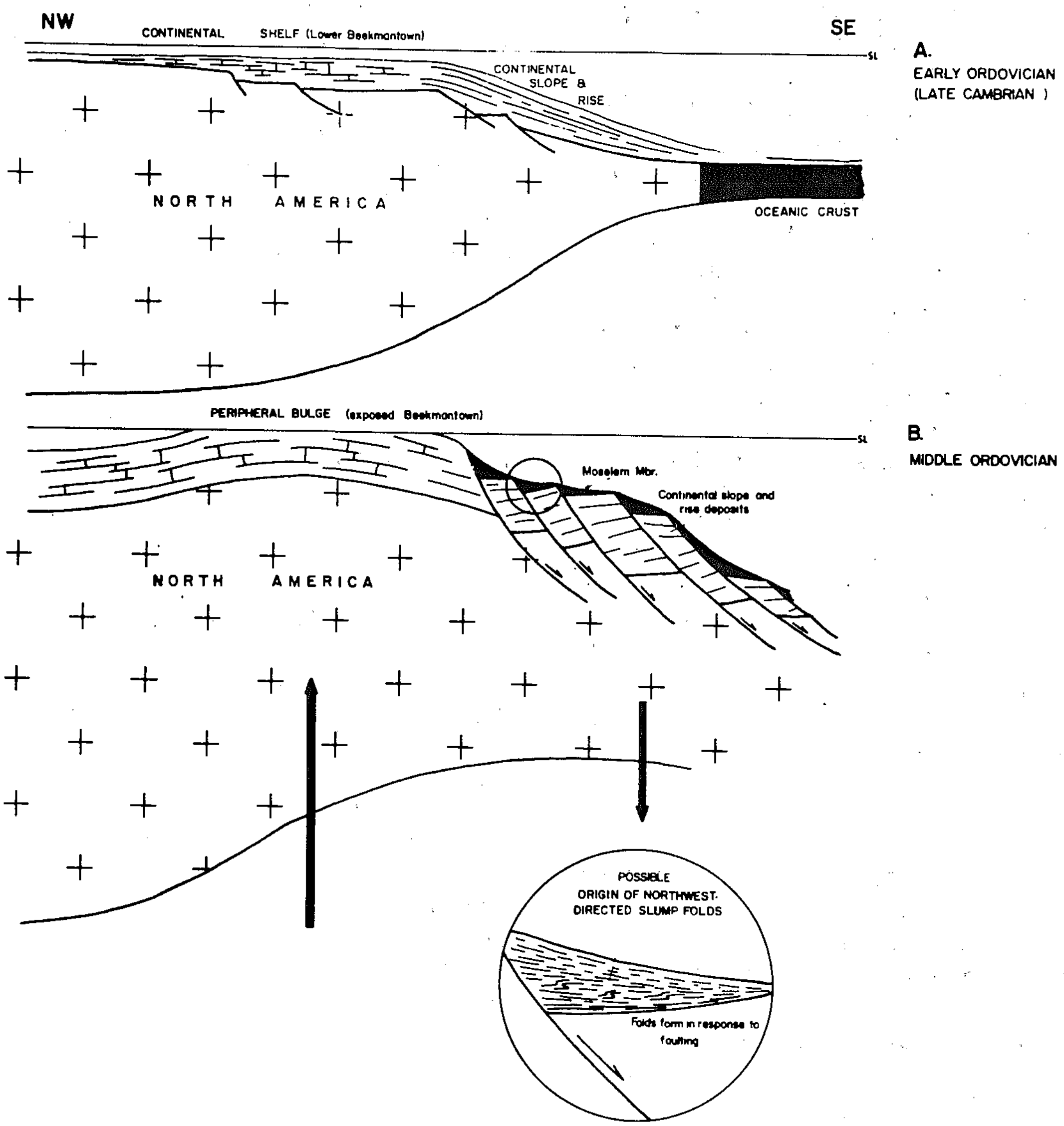


Figure 63. Proposed behavior of the eastern margin of North America in response to southeasterly drift and attempt subduction of North America in early Paleozoic time, (a) Early Ordovician-stable platform configuration, (b) Middle Ordovician-uplift of platform (peripheral bulge) concomitant with subsidence of the platform edge and deposition of the Moselem. Inset shows development of northwest-directed slump folds within this tectonic scenario.

clasts within conglomerate beds of the Moselem Member suggests derivation from lithified and diagentically mature deposits. The most logical source for these clastics would be eroded Beekmantown carbonate shelf rocks, in particular the Ontelaunee Formation (see fig. 46), which contains black chert. The eroding shelf supplied the fine-grained calcisiltite of the ribbon limestone turbidite deposits of the Moselem as well as the flint clasts of the conglomerate beds. This scenario would place the Moselem Member to the southeast of the peripheral bulge (fig. 63).

The lithology of the Moselem Member is consistent with sedimentation in a subsiding basin in which the rate of subsidence is dominant over that of sedimentation. The mechanism of subsidence cannot be determined for certain rocks that were stratigraphically beneath the Moselem Member cannot be seen and the amount of subsidence cannot be ascertained. Uplift of the North American platform due to passage across the outer swell followed by sudden collapse of the shelf as it foundered toward the trench is recorded by the Beekmantown disconformity and Jacksonburg Limestone of the eastern Pennsylvania Great Valley (Lash, 1980, 1983; Shanmugam and Lash, 1982). The presence of the flint clasts in Moselem Member conglomerates, the fact that the Moselem was deposited during the Blackriveran hiatus, and palinspastic restoration that places the Moselem to the southeast of the Beekmantown shelf suggests that the Moselem depocenter had already crossed the outerswell. The Moselem Member, then, probably records foundering of the North American continental rise as it approached the early Paleozoic subduction zone.

In general, analysis of slump folds of the Moselem Member indicates dominant movement to the northwest. This is easily explained as a consequence of growth or rollover faults that accompanied subsidence of the basin (fig. 63). Similarly, down-to-basin faults inclined to the southeast resulted in slumping in the Middle Ordovician Dolgeville facies of the Mohawk Valley of New York, lithologically identical and roughly correlative to the Moselem Member (Ritter, 1983; Fisher, 1979).

A modern analog that takes into account the important characteristics of the Moselem Member is the Sahul Shelf of northern Australia, which is presently foundering into the Timor Foredeep along numerous normal faults (Von der Borch, 1979). The tectonic history of the Sahul Shelf has also been chosen as an anaolgy to explain facies variations and folding of the Dolgeville facies of New York (Cisne and others, 1982). The Moselem Member represents rocks deposited in the foundering continental slope and rise area of North America as it approached the central Appalachian subduction zone (fig. 63). The thicker part of the craton, apparently that area beneath the Beekmantown depocenter, was uplifted and exposed at this time (Lash, 1983) whereas the continental slope and rise area, an area underlain by thinner continental crust and possibly weakened by early rift-related normal faults, was able to move toward the trench. This is analogous to the movement of the Sahul Shelf toward the Timor foredeep. Subsidence was probably accomplished along normal faults, some of which may have been reactivated rift-related faults at the edge of the craton (Cohen, 1982).

SPITZENBERG CONGLOMERATE

Spitzenberg, in the Kutztown quadrangle, the local name for "pointed hill" (see cover and fig. 64), is underlain by about 45 m (148 ft) of carbonate-clast conglomerate and red and green sandstone (Spitzenberg Conglomerate of Willard and Cleaves (1939). Behre (1933) first suggested a Triassic age for these rocks. This view was reiterated by Whitcomb and Engel (1934) and Whitcomb (1941) who noted lithologic similarities with Triassic border fault fanglomerates to the south. Stephens (1969) and Platt and others (1972) recognized the presence of Spitzenberg-like rocks beneath the Tuscarora Sandstone in Sharps Mountain west of Spitzenberg and suggested a Late Ordovician age for these rocks. However, Kaplan and Faas (1971) maintained that the Spitzenberg Conglomerate is Triassic in age. Most geologists now agree that these rocks are Late Ordovician to Early Silurian in age (Berg and others, 1980). It is likely that the Tuscarora Sandstone once rested upon the Spitzenberg at Spitzenberg but was subsequently removed by erosion.

Figure 64. Limestone-pebble conglomerates and sandstones hold up the sharp-crested hill of Spitzenberg in the center of the photograph. Sharps Mountain, the high ridge to the left underlain by northward dipping quartzites of the Tuscarora Sandstone, is marked by the white band of talus near the top of the mountain. The shoulder beneath the Tuscarora is underlain by the late Taconic sequence made up of sandstone, graywacke, and shale. These are in turn underlain by more complexly deformed red shale and graywacke of the Hamburg klippe.

The Spitzenberg Conglomerate consists of medium- to thick-bedded, coarse- to medium-grained sandstone. Primary structures include horizontal laminations and lenticular and wedge-shaped cross-bedding (fig. 65). The sandstone is moderately to poorly sorted and composed predominantly of rounded quartz grains and significantly lesser amounts of microcline, plagioclase feldspar, and lithic fragments. Accessory minerals include apatite, zircon, and opaque minerals. Clay matrix is sparse. Stephens (1969) presents a more detailed petrographic description of these deposits.

Polymictic, thin- to thick-bedded (up to 1 m (3.3 ft)), carbonate-clast conglomerate is a major rock type at Spitzenberg, but is not present at Sharps Mountain. Clasts include green chert, milky white calcilutite and horizontal- to cross-laminated calcisiltite, red and maroon mudstone, brown and gray sandstone and siltstone, and penecontemporaneous red and green sandstone. Clasts are typically rounded and ellipsoidal and range up to 25 cm (10 in) long. Clast:matrix ratios vary from bed to bed but range up to 9:1.Difference in fabric such as those described by Walker (1975) can be seen in the Spitzenberg Conglomerate. In general, sandstone beds are cross-bedded, rarely graded, and many have a long axis of pebbles transverse to flow fabric as deduced from associated foreset beds. In addition, some beds have imbricated clasts (fig. 66). Fabrics such as these are consistent with bed load rolling. Conglomerate beds, on the other hand, are rarely cross-bedded nor graded, and have a poorly developed fabric that is parallel to flow direction. This fabric is typical of sediment transported by a sediment gravity flow mechanism rather than by rolling (Walker, 1978).

The sedimentologic characteristics of the Spitzenberg Conglomerate are consistent with deposition on an alluvial fan (Bull, 1972, and Bluck, 1967). In particular, Bluck (1967) recognized four conglomerate facies in the Old Red Sandstone; Facies A, B, C, and D which reflect a change from alluvial fan sedimentation (Facies A and B) to fluvial stream deposits (Facies C and D). The rocks on Spitzenberg are most similar to Bluck's Facies B in that they are made up of continuous conglomerate beds that are overlain by cross-bedded gravelly sandstone. Beds of this facies probably result from sheetfloods, relatively low viscosity flood flows of sand and gravel that do not extend far from the downstream end of alluvial fan channels and deteriorate into patterns of braided channels and bars (fig. 67; Bull, 1972; Bluck, 1967). Thus, they are part of the alluvial fan and not the fluvial system. Individual conglomerate beds and overlying sandstone beds probably resulted from a single sheetflood in which a lag gravel was deposited followed by deposition of sandstone as the flood waned. Abundant paleocurrent data confirm a southeasterly source (fig. 68). However, the variation in current directions indicated in Figure 68 suggests that the basin was open to variable current movement.

One of the most interesting features of the conglomerate beds at Spitzenberg is the vertical variation in the ages of the clasts (see

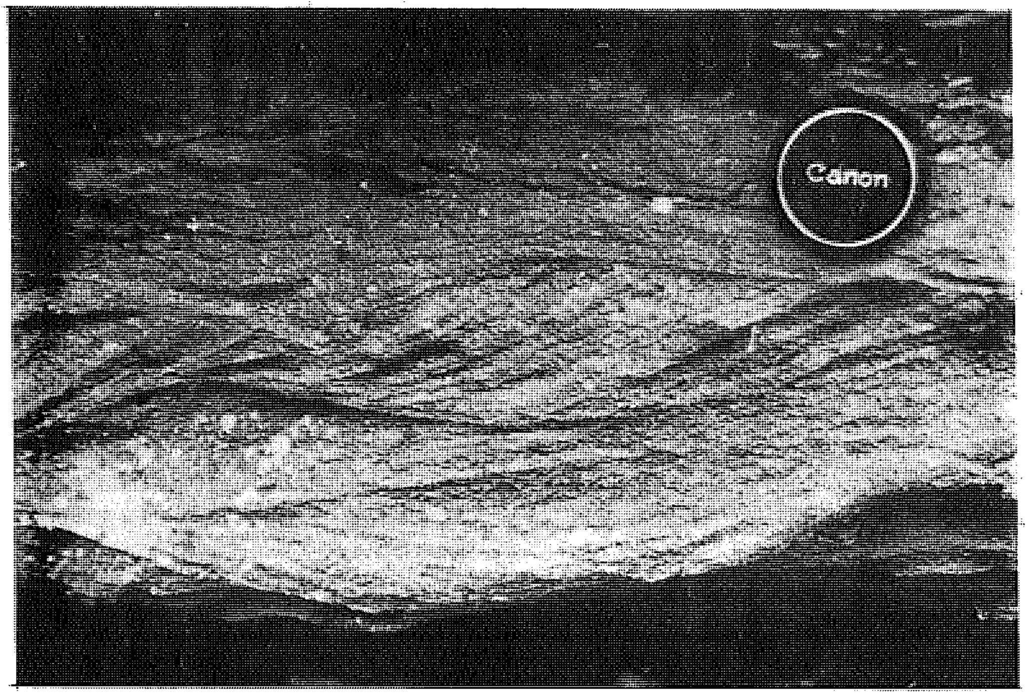


Figure 65A. Sandstone of the Spitzenberg Conglomerate: trough cross-stratification.

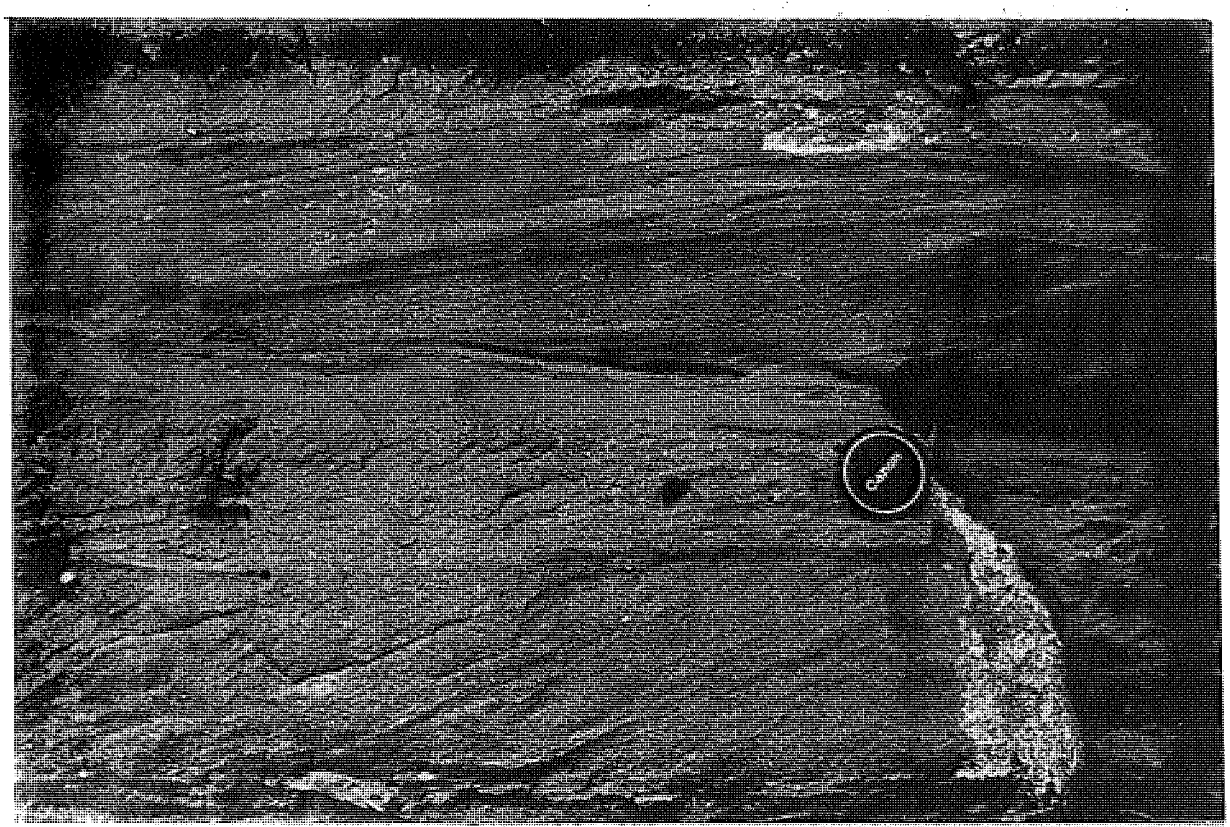


Figure 65B. Sandstone of the Spitzenberg Conglomerate: tabular cross-stratification.

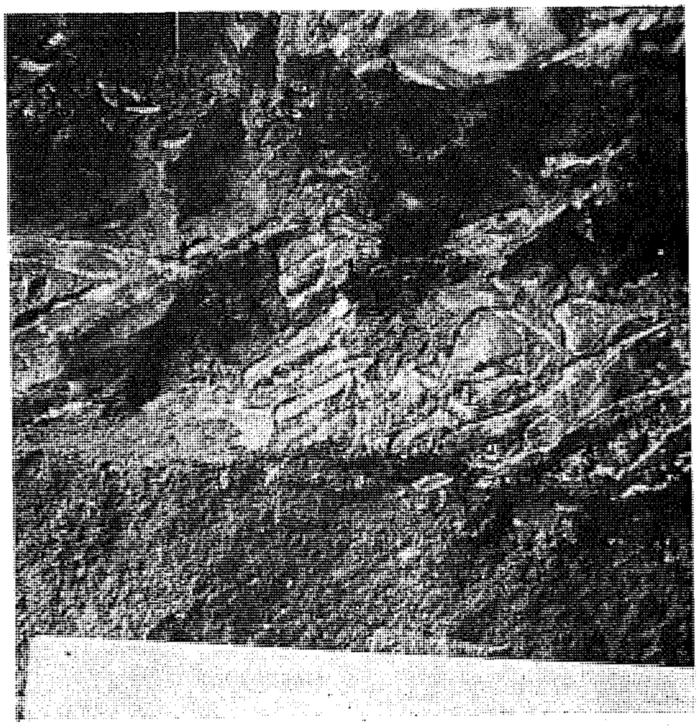


Figure 66. Imbricated limestone clasts within sandstone of the Spitzenberg Conglomerate (scale at bottom of photograph=12 cm).

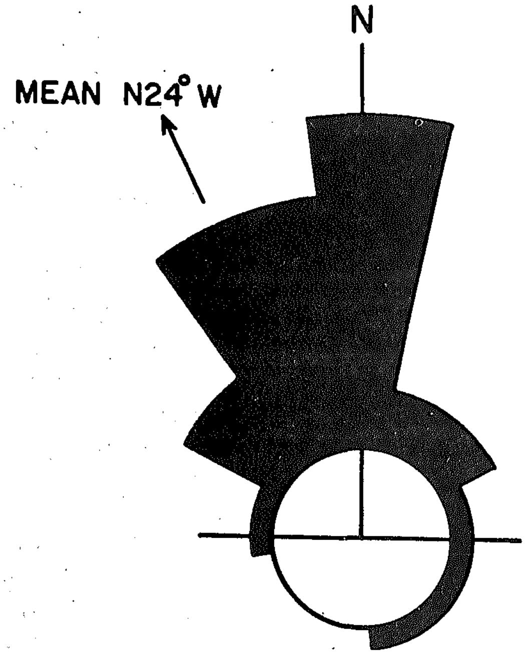


Figure 68. Paleocurrent data from the Spitzenberg Conglomerate.

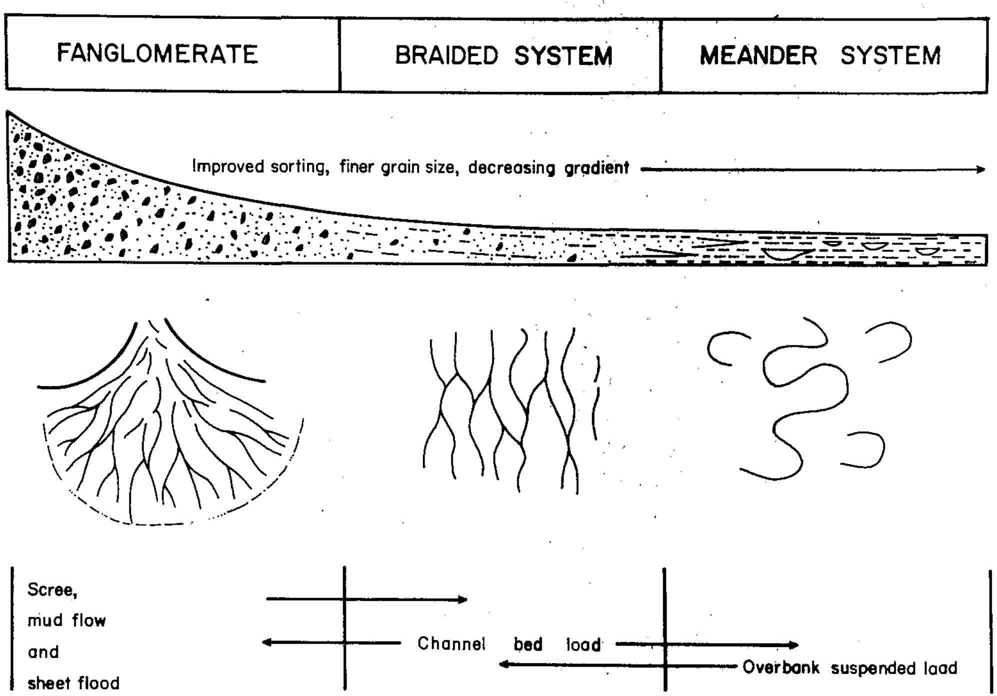


Figure 67. Changes in sedimentary processes and deposits from alluvial fans to braided and meandering channel systems. After Selley (1978).

section "Conodonts from Spitzenberg" by John E. Repetski). Brachiopod-bearing clasts collected from the lower part of the section are of Jacksonburg Limestone (Rocklandian-Kirkfieldian) age. Conodonts collected from limestone cobbles upsection are of late Early Ordovician age and comprise a Balto-Scandic fauna suggesting that they were derived from the underlying Hamburg klippe. Conodonts collected from limestone cobbles at the top of the section are Early Ordovician in age and comprise a North American Midcontinent fauna. These inverted age relations do not date the Spitzenberg Conglomerate but they do document uplift and erosion of a regionally upright sequence. The clasts in the conglomerate beds and sedimentary characteristics are consistent with unroofing and erosion of underlying allochthonous rocks and rocks of the Great Valley followed by sheetflood sedimentation of these deposits in the lower part of an alluvial fan. Triggering events of the sheetfloods may have been earthquakes and/or oversteepening of the fan slope. Although the Spitzenberg Conglomerate cannot be directly dated, we agree with the suggestion of Platt and others (1972) that it is Late Ordovician in age.

The sandstones underlying Sharps Mountain are interbedded with graywackes and, based on the nature of float, dark-gray shales. The graywackes and shales are similar to many rocks found in the Hamburg klippe. Sandstones similar to those underlying Sharps Mountain occur on Little Mountain about 50 km (31 mi) to the west. The Tuscarora and Shawangunk sit unconformably on a variety of older rocks in Berks and Lehigh Counties, including the upper (Pen Argyl) and middle (Ramseyburg) members of the Martinsburg Formation, the Shochary Sandstone and New Tripoli Formations, on red shales and graywackes of the Hamburg klippe, on the sandstones underlying Sharps Mountain, and probably on the Spitzenberg Conglomerate as well as the sandstones of Little Mountain farther to the west. The sandstones in Sharps Mountain and Little Mountain are similar to those of the Quassaic Quartzite underlying Marlboro Mountain, 190 km (118 mi) to the northeast, near Newburgh, New York. As the Sharps Mountain and Marlboro Mountain sandstones contain rock fragments from the underlying Hamburg and Taconic allochthons, respectively, we believe that they are fairly localized submarine fan deposits derived from unroofing and erosion of the Taconic rocks, which occurred prior to the profound period of uplift and erosion when the Tuscarora and Shawangunk clastic wedge was deposited. At Marlboro Mountain, it is possible that the Quassaic Quartzite was folded into a large syncline during a post-Taconic, pre-Silurian deformation.

THE MARTINSBURG FORMATION AND THE ROCKS OF SHOCHARY RIDGE

Stratigraphic and sedimentologic relations

The Martinsburg Formation of the parautochthonous Lehigh Valley sequence contains three members, the lower Bushkill, the middle Ramseyburg, and the upper Pen Argyl, and has a total thickness of as much as 4475 meters (14,320 ft) (Drake and Epstein, 1967). The thin-bedded ribbon slates of the Bushkill (fig. 69) contrast sharply with the thin- to thick-bedded carbonaceous slates of the Pen Argyl, allowing very easy recognition of both of these units. Stops 7 and 9 on this trip will illustrate the distinctive differences. The intervening Ramseyburg contains about 20 percent graywacke and graywacke siltstone that exhibit typical flysch sedimentary structures such as graded beds, planar and cross-laminated

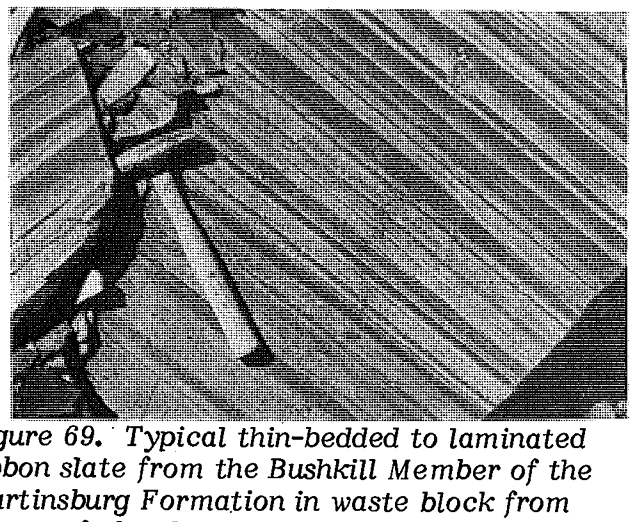


Figure 69. Typical thin-bedded to laminated ribbon slate from the Bushkill Member of the Martinsburg Formation in waste block from dumps of abandoned Theodore Whitesell quarry, in the southeast corner of the Wind Gap 7 1/2-minute quadrangle, 2.4 km (1.5 mi) ENE of Stockertown, Pa. Thickest bed seen on cleavage surface is less than 5 cm (2 in) thick.

beds, convolute beds and load casts (see Plate 1 for a detailed description of these rocks). Most of these fairly deep-water turbidites are basal cut-out (T_{c-e}) sequences (Bouma, 1962). On this field trip we will not have a chance to visit the Ramseyburg Member.

The rocks of the Shochary Ridge include the Shochary Sandstone of Middle and Upper Ordovician age and the New Tripoli Formation of Middle Ordovician age. The name Shochary Sandstone as used in this report differs somewhat from the original definition of Willard and Cleaves (1939). As redefined by Lyttle, Lash, and Epstein (1985), it includes only the upper sandstone member of the Shochary Ridge sequence found on Shochary Ridge. It is in gradational contact with the underlying New Tripoli Formation (Lyttle, Lash, and Epstein, 1985) and is not meant to include any of the graywacke-rich rocks of the Ramseyburg Member of the Martinsburg. The Shochary Sandstone (this report) is equivalent with parts of Units 3 and 4 of Wright and Stephens (1978, Figure 7). The New Tripoli Formation is partly equivalent to Units 2 and 3 of Wright and Stephens (1978, Figure 7).

The Shochary Sandstone is a medium-dark-gray, thin- to thick-bedded (5-75 cm, 2-30 in) calcareous, pyritiferous graywacke turbidite interbedded with medium-dark-gray, light-olive-brown weathering slate, calcisiltite, and minor thin beds of conglomerate. The amount of graywacke averages between 10 and 20 percent with rare instances up to 50 percent. In general, the percentage of graywacke increases going up in section. These beds commonly contain abundant faunal debris locally concentrated in rusty-weathering channels that also contain significant pyrite. The graywacke beds and conglomerates contain rare rounded chert fragments. The unit is approximately 1520 m (5000 ft) thick. However, the upper part of the unit is not present due to faulting and erosion. It is structurally and unconformably overlain by the Tuscarora Sandstone (see Plate 1). The lower contact of the Shochary Sandstone is transitional over 15 m (50 ft) with the underlying New Tripoli Formation.

The New Tripoli is characterized by medium-dark-gray, light-olive-brown weathering, thin, very evenly-bedded calcareous graywacke interbedded with fairly thick slate and calcisiltite beds (up to 50 cm, 20 in). In some places the fine parallel laminations in these sandstones are disturbed by bioturbation. Shelly fossil debris is common, especially near the upper contact, but not as abundant as in the Shochary Sandstone. The minimum thickness of the New Tripoli is 1500 m (4900 ft). The total thickness of the Shochary sequence, therefore, is approximately 3020 m (9900 ft), which is considerably greater than earlier estimates of Wright and Stephens (1978, see Figure 3). The contact between the New Tripoli and the Shochary is placed where sandstones commonly make up over 10 percent and become thicker than 15 cm (6 in). The lower contact is not exposed because of faulting of both Taconic and Alleghanian age.

Wright and Stephens (1978) have determined current directions in the Shochary Ridge rocks and show that in the lower part of the section the direction is predominantly due north. Climbing up the stratigraphic section, the current directions begin to fan out over a wider range from northwest through north to northeast. As suggested by Wright and Stephens, this is consistent with a northward prograding delta. The Shochary delta probably prograded into the same depositional trough that was accepting Martinsburg sediments, and was quite limited in extent. This is in considerable contrast to the Martinsburg Formation which was deposited along the entire length of a very large northeast-southwest trending trough. McBride (1962) showed that the currents carrying the turbidites of the Martinsburg generally flowed parallel to the length of the basin, or toward the northeast or southwest.

It is possible that the rocks of Shochary Ridge may have been deposited in slightly shallower water than the Martinsburg. A collection of graptolites from the New Tripoli Formation appear to be limited to an assemblage that is found in shallower water than the assemblages that are common in the Bushkill Member of the Martinsburg (Berry, 1977, written communication; see section "Graptolite zones in the Ordovician clastic rocks" below). The sandstones of the Shochary and New Tripoli are somewhat more mature in composition than those of the Martinsburg, and much more mature than those of the Windsor Township. Therefore, it is also possible that the rocks of the Shochary sequence are reworked Windsor Township sediments that were shed from the toe of the stack of thrust sheets carrying the Greenwich slice at its base. As this stack of thrusts moved from the southeast toward the north or northwest, approached and entered the Martinsburg depositional basin, the basal thrust (Weisenburg Church fault) locked and a new, structurally lower thrust (Game Preserve fault), cut through the Shochary Ridge rocks and carried them as part of an even larger stack of thrusts onto the Martinsburg Formation deeper in the basin and farther to the north (see section entitled "Structural geology of the Great Valley and Valley and Ridge"). The fact that the eastern end of the synformally folded thrust sheet containing the Shochary Ridge section coincides with the eastern end of the Hamburg klippe thrust sheets lends support to this idea.

Graptolite zones in the Ordovician clastic rocks

Based on a large number of new graptolite localities, Wright, Stephens and Wright (1979) presented a revised stratigraphy of Ordovician clastics in the field trip area. Relying on that revised stratigraphy they also proposed a tectonic history that is significantly different than the one we propose in this guidebook (see section entitled "Structural geology of the Great Valley and Valley and Ridge"). Because we disagree with many of their conclusions based on this faunal data, we would like to address some of the problems of graptolite zonation that have direct bearing on all work in Middle and Upper Ordovician clastic rocks.

Using the zonation and faunal lists of Riva (1969, 1974), Wright, Stephens and Wright (1979) show that the rocks of the Martinsburg Formation extend from the Diplograptus multidens Zone near the base through the Corynoides americanus and Orthograptus ruedemanni Zones to the Climacograptus spiniferus Zone at the top. In addition, they claim that the rocks of the Shochary Ridge area span the same zones, have identical lithologies, and are therefore part of, and continuous with, the Martinsburg Formation. They dispute the division of the Martinsburg into three members, arguing that the upper Pen Argyl Member is older than the middle Ramseyburg Member and equivalent with the upper part of the lower Bushkill Member. Before discussing some details of the faunal evidence presented by Wright, Stephens and Wright (1979), it is helpful to look at recent reviews of modern graptolite zonation schemes.

Based on recent work in Canada (Walters, Lesperance, and Hubert, 1982) and a careful study of Riva's own faunal lists by Finney (1982), Ross and others (1982) suggests that the range of the Climacograptus spiniferus Zone of Riva may overlap the ranges of the Corynoides americanus and Orthograptus ruedemanni Zones. That is, the Climacograptus spiniferus Zone may be the lateral equivalent of several other zones, and the graptolites that characterize and define these zones may be influenced by environmental factors. The possibility of facies control of graptolites has been gaining popularity in recent years (Skevington, 1974; Cisne and Chandlee, 1982; Finney (1984)). We feel that a new look at the graptolite data in the field trip area in the light of this work is extremely important.

Most of the graptolite data pertaining to our field trip area is presented in Figure 2 and Table 1 of Wright, Stephens and Wright (1979). A patient and careful comparison of our geologic map (Plate 1) with their Figure 2 suggests that what we map as the Pen Argyl and Ramseyburg Members of the Martinsburg have totally overlapping graptolite zones as presented by Wright, Stephens and Wright (1979). That is, the Pen Argyl and Ramseyburg on Plate 1 both range through the Corynoides americanus, Orthograptus ruedemanni, and Climacograptus spiniferus Zones. What we map as Bushkill Member ranges from the Diplograptus multidens into the Corynoides americanus Zone. Because we feel that we have excellent sedimentological evidence establishing the relative stratigraphic position of the members of the Martinsburg, we have difficulty reconciling the contradictory faunal evidence with our detailed mapping. One possible escape from this geological cleftstick, as mentioned above, is the possibility of ecological or environmental controls on the distribution of graptolites.

Wright, Stephens and Wright (1979, see Figure 2) draw lines denoting boundaries of graptolite zones on a map displaying two lithic types, sandstone and shale. These two lithic types correspond on our Plate 1 with the 3 members of the Martinsburg, 2 formations of the Shochary Ridge sequence, and 5 members of the Windsor Township Formation. In several locations the graptolite zone boundary lines are at high angles to structures and contacts that we have mapped. Although it is possible to imagine diachroneity of units on a regional scale, it seems unlikely that one or more mapped units change age abruptly along strike. In addition, we feel that the faunal evidence is not sufficient in several areas to properly constrain these boundary lines.

Wright, Stephens and Wright (1979) contend that the rocks of the Shochary Ridge rest conformably upon the rocks of the Hamburg klippe. Arguing that the age of the oldest Shochary Ridge rocks falls within the *Diplograptus multidens* Zone, they suggest that the klippe was emplaced during or prior to *Diplograptus multidens* time. Like others before us (Stose, 1930; Alterman, 1972) we feel that the klippe rests structurally, in fault contact, on top of the Shochary Ridge rocks. Plate 1 shows the rocks of Shochary Ridge completely bounded by thrust faults. One of these faults, the Kistler Valley fault, will be seen at Stop 8 on this field trip. Because the Game Preserve fault marks the base of the Shochary Ridge stratigraphic section, it is impossible to know how much of the lower New Tripoli Formation is missing. In addition, we feel that the Kistler Valley fault, which brings the klippe structurally on top of the Shochary Ridge rocks, is Alleghanian in age. For all of these reasons the *Diplograptus multidens* age of the New Tripoli Formation should only be used as a constraint on the earliest possible time of emplacement of the klippe on the Shochary Ridge rocks. The emplacement of the allochthon and the Shochary Ridge rocks as a single entity onto the Martinsburg also must post-date the *Diplograptus multidens* Zone age and perhaps the *Corynoides americanus* Zone age of the Bushkill.

STRATIGRAPHY AND SEDIMENTOLOGICAL HISTORY OF SILURIAN AND DEVONIAN ROCKS IN EASTERN PENNSYLVANIA

Generalized descriptions of the Silurian and Devonian stratigraphic units to be seen on this field trip are given in Plate 1. Exposures are generally dreadful, mainly because of deep weathering west of the Wisconsinan terminal moraine, which crosses the outcrop belt of these rocks near Saylorsburg, Pennsylvania. In places, some of the shaly limestones and limy shales are a sedimentary rock saprolite more than 150 feet deep (Epstein and Hosterman, 1969). For those of you familiar with these rocks to the northeast in easternmost Pennsylvania, and to the west in central Pennsylvania, there are a few facies and nomenclatural changes that are worth noting here. The quarry at Andreas, Pennsylvania (Stop 10) affords a rare opportunity to see decent exposures of the section from the Poxono Island Formation to the Selinsgrove Limestone.

Sandstones and conglomerates that overlie the Ordovician rocks in the field trip area are called the Shawangunk Formation in the east and Tuscarora Sandstone and Clinton Formation in the west. The Shawangunk in eastern Pennsylvania contains several well-defined members. However, five km (3 mi) east of Andreas on Blue Mountain, only two members can be recognized. Thus, to the west of that point the names Tuscarora and Clinton are applied. In this regard, Plate 1 differs slightly from the distribution of these units shown on the state geologic map of Pennsylvania (Berg and others, 1980).

In easternmost Pennsylvania, the Upper Silurian through lower Middle Devonian interval consists of more than 450 m (1500 ft) of sandstone, shale, limestone, and dolomite (Epstein and others, 1967). These decrease in thickness to about 150 m (500 feet) at Andreas, Pennsylvania (fig. 72). The thickness continues to decrease and about 72 km (45 mi) southwest of Andreas the entire interval is absent. Concomitant with the southwestward thinning of individual formations, several units become more clastic as an ancient low-lying positive area is approached. The area was called the "Harrisburg axis" by Ulrich (1911) and Willard (1941) and termed the "Auburn Promontory" by Swartz (in Willard and others, 1939). Wood and others (1969) delineated this area of stratigraphic thinning in several isopach maps.

The Poxono Island Formation, Bossardville Limestone, and Decker Formation thin to the southwest (fig. 72), especially the Poxono Island. Their lithologic characteristics, however, are similar to those in easternmost Pennsylvania. The interval from the Rondout Formation through the Port Ewen Shale of the Helderburg Group, on the other hand, changes dramatically and most of these units pinch out. The Rondout Formation, Minisink Limestone, and Port Ewen Shale are not recognized southwest of Bossardsville. The New Scotland Formation does not extend west of the Pennsylvania Turnpike. The Andreas Red Beds of Swartz and Swartz (1941) are only found near Palmerton and to the southwest. It is a unique red bed interval whose age is uncertain, but it immediately overlies limestones of the Decker Formation, whose age has now been determined to be Pridolian (uppermost Silurian), according to conodont collections made by Denkler (see section entitled "Upper Silurian biostratigraphy of the Andreas Quarry, Pennsylvania" in this volume). The Andreas may be correlative to the Rondout Formation or upper part of the Decker Formation. The age of sandstones lying above the Andreas Red Beds in the environs of Andreas are problematic. They may be a westward clastic facies of the Coeymans Formation, which in the Delaware Valley of eastern Pennsylvania consists of four members. These are, from base upwards, the Shawnee Island (limestone), Peters Valley (calcareous sandstone), Depue Limestone, and Stormville (Calcareous sandstone and conglomerate) Members (Epstein and others, 1967). The Peters Valley and especially the Stormville increase in thickness from a feather edge in north central New Jersey to the Bossardville area at the expense of the interbedded limestone members. At Kunkletown these sandstones contain the typical Coeymans fossil Gypidula coeymanensis. At Andreas, these sandstones lack this definitive fossil, but they appear to occupy the Coeymans position, based on regional mapping. Thus, the unit is tentatively called the "Stormville Formation", because it is the Stormville Member that increases most in thickness to the southwest when last seen in areas of good exposure northeast of Bossardville, Pa.

The "Stormville Formation" is overlain by deeply weathered cherty shale (also leached limestone?) and sandstone of the Oriskany Group. These are poorly exposed, as are rocks of the overlying Schoharie and Esopus Formations. The Palmerton Sandstone, a westward facies of the uppermost part of the Schoharie Formation and lower Buttermilk Falls Limestone, is the fourth unit in the Andreas area that contains sandstone (the others are the Andreas Red Beds, "Stormville Formation", and Oriskany Group). Near Schuylkill Haven, 34 km (24 mi) southwest of Andreas, a five-foot thick bed of conglomeratic sandstone overlies the Decker Formation and underlies the Selinsgrove Limestone. That sandstone bed could be any of the four units mentioned above. However, reconnaissance mapping southwest of Andreas suggest that the Palmerton Sandstone quickly disappears. The sandstone bed at Schuylkill Haven does not contain the molds of Spirifer arenosus that are typical of the Oriskany, and red beds do not extend that far southwestward. Therefore, it is believed that the bed at Schuylkill Haven is most probably the "Stormville Formation", but this cannot be proved at this time.

The Buttermilk Falls Limestone of eastern Pennsylvania is the lateral equivalent of the Onondaga Limestone of New York State (Epstein, 1985). The Buttermilk Falls characteristically contains chert, but between the Pennsylvania Turnpike and Andreas, chert drops out of the section and the unit is referred to the Selinsgrove Limestone. Dark-gray shales at the bottom of the Selinsgrove may be the correlative of the Needmore Shale of central Pennsylvania.

Rapid shallowing of the deep Martinsburg basin was accomplished by deposition of thick Martinsburg detritus and by tectonic uplift due to intense Taconic mountain building, which peaked during the Late Ordovician. This period of orogenic activity and regional uplift was followed by deposition of coarse terrestrial deposits of the Tuscarora Sandstone and Shawangunk Formation. The contact between the Tuscarora/Shawangunk and underlying Ordovician rocks is a regional angular unconformity, and at Schuylkill Gap (Stop 6) the two units are nearly at right angles to each other.

The conglomeratic sandstone members of the Shawangunk Formation (Weiders, Minsi, and Tammany Members), and the Tuscarora Sandstone, are believed to be fluvial in origin (Epstein and Epstein, 1972) and are interposed by a transitional marine-continental facies (the Lizard Creek Member of the Shawangunk and the Clinton Formation). The fluvial sediments are characterized by rapid alternations of polymictic conglomerate with quartz pebbles more than 15 cm (6 in) long, conglomeratic sandstone, and sandstone (cemented with silica to form quartzite), and subordinate siltstone and shale. The bedforms indicate rapid flow conditions. Crossbed trends are generally unidirectional to the northwest. Minor shales and siltstones are thin, and are partly mudcracked, indicating subaerial exposure. These features indicate that deposition was by steep braided streams with high competency and erratic fluctuations in current flow and channel depth. Rapid runoff was undoubtedly aided by lack of vegetation cover during the Silurian. The finer sediments present are mere relicts of any that may have been deposited in overbank and backwater areas.

The Lizard Creek Member and Clinton Formation contain many rock types and a variety of sedimentary structures that suggest a complex transitional (continental-marine) environment, including tidal flats, tidal channels, barrier bars and beaches, estuaries, and shallow neritic condition. These are generally highly agitated environments, and many structures, including flaser bedding (ripple lensing), uneven bedding, rapid alternations of grain size, and deformed and reworked rock fragments and fossils support this interpretation. The occurrence of collophane, siderite, chlorite nodules, and Lingula fragments indicate nearshore marine deposition. Many of the sandstones in the Lizard Creek are supermature, laminated, rippled, and contain heavy minerals concentrated in laminae. These are believed to be beach or bar deposits associated with the tidal flats.

The outcrop pattern of the Shawangunk Formation and Tuscarora Sandstone and the coarseness of some of the sediments suggest that they were deposited on a coastal plain of alluviation with a linear source to the southeast and a marine basin to the northwest. Farther west in central Pennsylvania the Tuscarora is largely marine in origin (Cotter, 1983). Erosion of the source area was intense and the climate, based on study of the mineralogy of the rocks, was warm and at least semi-arid. The source was composed predominantly of sedimentary and low-grade metamorphic rocks with exceptionally abundant quartz veins and small local areas of gneiss and granite.

As the source highlands were eroded, the braided streams of the Shawangunk gave way to gentler streams of the Bloomsburg Red Beds. The rocks in the Bloomsburg are in well- to poorly defined upward fining cycles that are characteristic of meandering streams. The cycles are as much as 4m (13 ft) thick and ideally consist of a basal crossbedded to planar-bedded sandstone that truncates finer rocks below. These sandstones were deposited in stream channels and point bars through lateral accretion as the stream meandered. Red shale clasts were derived from caving of surrounding mud banks. The basal crossbedded sandstone grade up into laminated finer sandstone and siltstone with small-scale ripples indicating decreasing flow conditions. These are interpreted as levee and crevasse-splay deposits. Next are finer overbank and floodplain deposits containing irregular carbonate concretions. Burrowing suggests a low-energy tranquil environment; mudcracks indicate periods of desiccation. The concretions are probably caliche precipitated by evaporation at the surface. Fish scales in a few beds suggest a fluvial or a lagoonal environment. A bothersome point to the fluvial interpretation of the Bloomsburg is that epsilon crossbeds are not prevalent.

The source for the Bloomsburg differed from that of the Shawangunk because the red beds required the presence of iron-rich minerals, suggesting an igneous or metamorphic source. Evidently, the source area was eroded down into deeper Precambrian rocks.

From Poxono Island time through Oriskany time, the fluvial deposits of the Bloosmburg gave way to transgression of a shallow marine shelf. The area was maintained near sea level and a complex of alternating supratidal and intertidal flats, barrier bars, and subtidal zones was maintained.

Sediments indicative of supratidal flats contain edgewise conglomerates, laminations of possible algal origin, fine-grained laminated to very thin-bedded massive dolomite and limestone, replacement dolomite, very restricted fauna (mainly leperditiid ostracodes) or no fossils at all, and mudcracks. Supratidal sediments are found in the Poxono Island Formation, Bossardville Limestone, Decker Formation, and Rondout Formation in eastern Pennsylvania.

Intertidal flat sediments are characterized by graded laminated and thin-bedded partly quartzose limestone, cut-and-fill structures, small-scale crossbedding, and intraclasts (edgewise conglomerates), abundant leperditiid ostracodes, storm-tossed shell debris, and some mudcracks. Intertidal flat sediments are found in the Poxono Island Formation, Bossardville Limestone, Decker Formation, and Rondout Formation.

Barrier-bar and beach deposits are distinguished by calcareous sandstone, conglomerate, and quartzose limestone with foreshore laminations and crossbedding, cut-and-fill structures, intraclasts, skeletal debris of a variety of marine organisms, and scattered burrows. These deposits are common in the Decker Formation, Coeymans Formation, and Ridgeley Sandstone of the Oriskany Group.

Neritic deposits consist predominantly of calcareous shale and limestone that may contain abundant chert. Faunas are diverse and abundant, and burrowing may be extensive. Reefs developed locally in the Shawnee Island Member of the Coeymans Formation. Neritic units include the Decker Formation, Coeymans Formation, New Scotland Formation, Minisink Limestone, Port Ewen Shale, and Shriver Chert of the Oriskany Group.

The wandering shoreline migrated northwestward during deposition of the Poxono Island through the Oriskany, and the area became emergent following Oriskany time. This was followed by a rapid change to moderate to deep neritic conditions during deposition of the Esopus and lower Schoharie Formations. These rocks are characterized by persistence over a wide geographic area (eastern Pennsylvania to east-central New York), lack of abundant skeletal debris, abundant hexactinellid sponge spicules, and abundant Taonurus. A regressive phase followed from Schoharie into Buttermilk Falls time as indicated by an upwards transition from horizontal to vertical burrows, an increase in marine fauna (including corals), and an increase in limestone. These features indicate water depths within the photic zone and warm, well-oxygenated, and gently circulating water. The Palmerton Sandstone, found in the southwestern part of the area between the Schoharie Formation and Buttermilk Falls Limestone, was most likely a marginal marine (bar or beach) linear sand body. It is massive and generally lacks distinctive internal structures, making a precise interpretation of its depositional environment a bit uncertain.

The black pyritic shales of the Marcellus Shale, with its depauperate fauna, reflects development of an anoxic basin below wave base. The Tioga Bentonite Bed, an altered volcanic ash at the base of the Marcellus in the Lehigh Valley and in the upper part of the Buttermilk Falls in the Delaware Valley, marks the top of the Onesquethaw Stage. It marks a period of igneous activity with a volcanic center lying somewhere in the Piedmont of central Virginia (Dennison, 1969).

The overlying Mahantango Formation, which contains coarser siltstones than the Marcellus, and very diverse fauna (brachiopods, corals, pelecypods, bryozoans, trilobites, etc.), indicates a shallower marine environment with better circulation than the Marcellus. Local biostromes containing abundant corals attest to the return of more "normal" marine conditions.

The Trimmers Rock Formation contains many features suggesting deposition from turbidity currents, including graded sequences, scoured bases, transported fossil hash, and sole marks. These rocks are transitional up into deltaic deposits of the Catskill Formation, and were probably deposited in the prodeltaic apron in front of the advancing Catskill delta.

STRUCTURAL GEOLOGY OF THE GREAT VALLEY AND VALLEY AND RIDGE

Thrust faults of several ages are the dominant structures in the field trip area. All of the rocks in the area have been folded at least twice and some of this folding may be directly related to the thrust faulting. If this is true, and the thrusting has developed sequentially throughout a range of time in the Paleozoic, the resultant folds and associated cleavages may well have developed sequentially over a range of time. Much of the effort to determine the age of the regional slaty cleavage in the central Appalachian Great Valley assumes that it is one age throughout the region. This may be a faulty premise. Instead of the main regional slaty cleavage being the result of a single deformation affecting all of the rocks in the central Appalachian Great Valley, the cleavage development, as envisioned by Mitra and Elliott (1980) for Proterozoic gneisses in the Blue Ridge, may be progressive through time and space.

The work of Mitra and Elliott (1980), as well as work by Perry (1978) and Nickelsen (1979), suggests that thrust faults are progressively younger approaching the foreland. Even though a sequential development of thrust faults can be documented in the field trip area, reactivation of older faults by younger ones, and the cutting of old faults by younger ones, complicates any simple "younging-to-the-foreland" pattern. Looking at the problem from a larger perspective, for example by examining the cross section on Plate 1, or cross section A-A' of Lyttle and Epstein (1985), it is possible to argue that the major thrust faults in the central Appalachians as a whole become younger and younger going from the Piedmont toward the Valley and Ridge province. This field trip is located in an important transition zone where it appears that thrust faults of both Taconic and Alleghanian age occur, whereas to the south of the area, in the Piedmont, the age of thrusting appears to be almost entirely Taconic, and to the north of the area, in the Valley and Ridge, entirely Alleghanian.

Examples of the older faults in the area, interpreted to be Taconic in age, are the Weisenburg Church fault and the Game Preserve fault (fig. 70). The Weisenburg Church fault brings the rocks of the Windsor Township Formation of the Hamburg klippe on top of the Shochary Ridge rocks and the Martinsburg, and

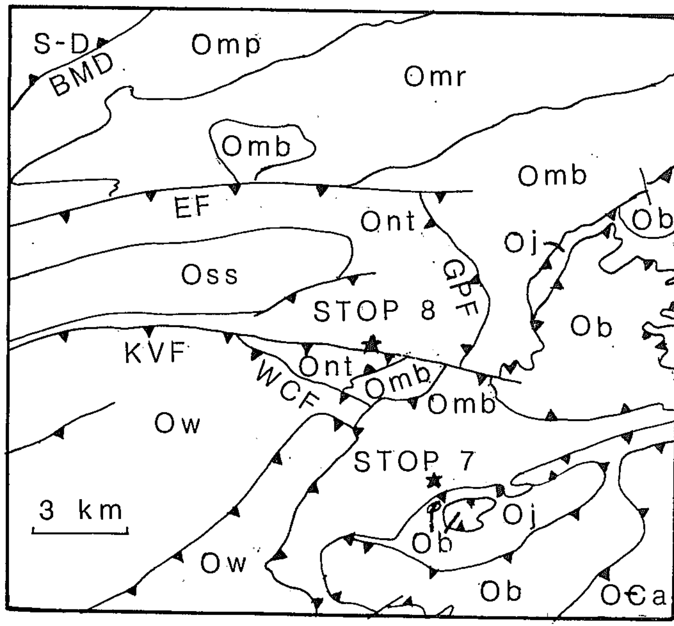


Figure 70. Simplified geologic map in the vicinity of STOPS 7 and 8 showing the following thrust faults: Blue Mountain decollement (BMD), Eckville fault (EF), Kistler Valley fault (KVF), Weisenberg Church fault (WCF), Game Preserve fault (GPF). The unnamed fault immediately south of Stop 7 is shown to be cut by the Kistler Valley fault 3.3 km (2 mi) east of Stop 8. The unnamed fault south of STOP 7 is also highly folded while the Kistler Valley fault is quite straight, also a reflection of their relative ages. This map also shows that the Game Preserve fault (GPF) is folded along with the rocks of Shochary Ridge into a large synform. Formations abbreviated as follows: S-D, all Silurian and Devonian rocks; Omp, Pen Argyl Member; Omr, Ramseyburg Member; Omb, Bushkill Member; Oss, Shochary Sandstone; Ont, New Tripoli, Oj, Jacksonburg Limestone; Ob, Beekmantown Group; O€a, Allentown Dolomite; Ow, Windsor Township.

represents the only place where one may be seeing the true structural base of the klippe at its eastern end.

Like the Weisenburg Church fault, the Game Preserve fault is a subtle structure difficult to recognize in the field. It is placed at the contact of the New Tripoli Formation and the Bushkill Member of the Martinsburg, and is mostly exposed along the western edge of the Cementon quadrangle (Drake, unpublished maps). It is likely that that this fault carried partially consolidated Shochary Ridge rocks over Martinsburg prior to the formation of the regional slaty cleavage. Subsequent deformation and low grade metamorphism have partly healed this fault and other faults of Taconic age and make recognition difficult. Although this fault, where exposed at the surface, only juxtaposes New Tripoli against the Bushkill, it is interpreted to have cut up-section through all three members of the Martinsburg. As shown on the cross section A-A' of Plate 1, part of the Game Preserve fault has been reactivated on the north side of the Shochary syncline. That is, the later, presumably Alleghanian, Eckville fault (see fig. 70), first noted by Behre (1933), cuts the Game Preserve fault at a fairly high angle at its eastern end and is asymptotic with the Game Preserve fault farther west. This relation is possible since the Game Preserve fault has been folded into a large synform.

If sediments now constituting the Shochary Sandstone and New Tripoli Formation are partly reworked detritus of the Greenwich slice of the klippe as suggested above, they may have been shed from the advancing thrust slices as they entered the Martinsburg depositional basin. If so, they were probably transported just ahead of, and structurally beneath, the overriding Greenwich thrust sheet or tectonic slice. The Weisenburg Church fault probably represents the rarely exposed base of the Greenwich slice. Therefore, the klippe and Shochary Ridge rocks may have travelled together as a single package during the Taconic, with most of the movement being taken up on the lower Game Preserve fault. The Weisenburg Church fault may have ridden along piggy-back with only minor amounts of renewed movement during the final stages of klippe emplacement.

The Game Preserve fault, as well as the rocks both above and below it, were later folded into the large Shochary synform (Plate 1 cross section A-A'). An excellent slaty cleavage, regional in extent, and well-developed in the Martinsburg, in the Shochary Ridge rocks, and only locally and sporadically in the Windsor Township, is axial planar to this synform and other related east-northeast-trending folds. At the shared corner of the Slatedale, Topton, Cementon, and Kutztown 7 1/2 minute quadrangles, the New Tripoli Formation, as well as the Bushkill and the easternmost Windsor Township, all exhibit a pervasive smearing or extension lineation on southeast-dipping cleavage planes suggesting that all of these rocks may have been overridden by a northwest-travelling thrust sheet. The thrusting that produced this stretch lineation must either be the last movement of the thrusting that formed the Shochary synform or must post-date it.

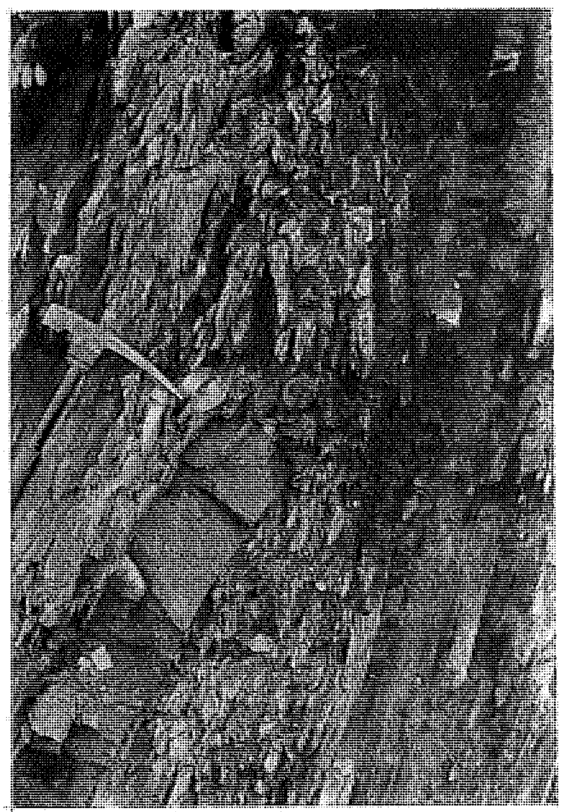
The age of the Shochary synform and associated slaty cleavage remains unknown. Their formation certainly post-dates the formation of the Game Preserve and Weisenburg Church faults. The late-Barnveldian age of the Shochary Sandstone puts a limit on the earliest age of the Game Preserve fault, whereas the Late Ordovician age (upper subzone of graptolite zone 13 of Berry) of the Pen Argyl puts a limit on the earliest age of the Weisenburg Church fault. Epstein and Epstein (1969) and Lash (1978) have suggested that the slaty cleavage in the Slatedale and surrounding quadrangles is Alleghanian in age. They contend that the earliest mapped folds mentioned above have also folded Silurian and younger rocks to the north. Data gathered in much of the field trip area are consistent with this view, but do not demand it. In other areas Drake (1969) and Lash (1980), argue for a Taconic age for the slaty cleavage. No absolute ages for cleavage development have been established, however, leaving the question open.

The Shochary synform and related folds were later warped by more east-west-trending upright open folds (examples of this generation of folding will be seen at Stop 7). These folds, too, may be the result of thrust faulting; in this case, numerous steeply south-dipping upthrusts that splay from a more gently dipping detachment surface at depth. In several localities these upthrusts reach the present erosion surface. Two such faults bound the Shochary Ridge rocks to the north and south, the Eckville and Kistler Valley faults, respectively (figs. 70 and 71). Both dip steeply to the south and show movement up on the south side. The Eckville fault brought Shochary Ridge rocks at least several hundred meters upward to their present configuration where they are in contact with all three members of the Martinsburg (see Plate 1). The fault cuts higher into the Martinsburg section going to the west. This fault was first recognized by Behre (1933) who noted that slaty cleavage and bedding had both been greatly rotated by movement along the fault. Rotation of cleavage over several tens of meters is seen only immediately adjacent to the late, presumably Alleghanian faults. Additional evidence for the Eckville fault consists of the truncation of map units, particularly on the north side, the formation of small kink folds only within a few meters of the fault, abundant vein quartz, which, when found in outcrop, occurs along fault surfaces and numerous small subsidiary faults on both sides of the major fault. The same features can be seen along the Kistler Valley fault. The Kistler Valley fault is preserved in a spectacular road cut within the New Tripoli Formation, 0.6 km northeast of Weidasville, in the southeastern corner of the quadrangle. This exposure, Stop 8, shows a large number of splay faults that totally disrupt bedding and cleavage.

Both the Eckville and Kistler Valley faults truncate the older Game Preserve fault. The Eckville fault coincides with of the northern limb of the synformally folded Game Preserve fault, but separates from it near the east end of the Shochary synform. The Kistler Valley fault not only truncates and offsets the north-dipping part of the Game Preserve fault, but it also truncates the Weisenburg Church fault. From the map pattern it would appear that the Kistler Valley fault may have both a vertical component of movement, up on the south, and a right lateral component.

The Blue Mountain decollement post-dates the deposition of the Silurian Shawangunk and is therefore, presumably, Alleghanian in age. Although it is impossible to prove, it is likely

A



B

Figure 71. A) A splay within the Kistler Valley fault zone, Kestrel Wildlife Refuge, New Ringgold quadrangle. Looking west, this E-W-trending thrust within the New Tripoli Formation dips 57^{o} to the south (left) and is one of many faults found in this outcrop. Here the New Tripoli is not quite as deformed as at STOP 8. B) Same outcrop as in (A), also looking west. Disruption of bedding and kinks in cleavage are both the result of thrusting within the Kistler Valley fault zone.

that both the Eckville and Kistler Valley faults cut the decollement. Faults mapped in the Silurian rocks in the New Ringgold quadrangle (Stephens, 1969) are directly on strike with the two faults, but can not be traced directly into them due to abundant talus on the slopes of Blue Mountain.

The nature of the contact between the Tuscarora Sandstone, and the correlative Shawangunk Formation farther east, with a variety of underlying Ordovician rocks in eastern Pennsylvania has been debated by geologists for many decades. Surely the contact is an angular unconformity of regional extent, as shown by the sedimentology of the rocks involved (coarse fluvial clastics abruptly overlying deep-marine shales and graywackes), by the discordance in strike wherever the contact is exposed, and by the overlapping of a variety of Ordovician rocks by the Tuscarora and Shawangunk (fig. 72). The haunting question is, how are the folds, faults, and related cleavage that are present in Silurian and younger rocks reflected in Ordovician and older rocks? This question has been discussed in the area to the east by Epstein and Epstein (1969), Epstein (1973), and Epstein, Sevon, and Glaeser (1974) and will not be detailed here. However, we will have an opportunity to briefly ponder the relationships at the contact between the Tuscarora and Windsor Township Formation at Stop 6. We will also be able to compare folds in the Martinsburg Formation at Stop 9 with folds in Silurian and Devonian rocks at Stop 10.

The structure in Silurian and Devonian rocks in the field trip area is typical of the Valley and Ridge province in the central Appalachians of Pennsylvania. Northwest-verging upright and overturned folds of all magnitudes are the most obvious structural features seen (see cross section in Plate 1). Many of these are related to imbricate faults and decollements, many of which are blind or buried and inferred from regional relationships. Faulting becomes more prevalent southwest of Lehigh Gap. Folding is disharmonic and four lithotectonic units, with their own deformation characteristics, were recognized in these rocks by Epstein and Epstein (1967, 1969) and Epstein and others (1974). Zones of detachment, with northwest movement of the overriding plates, were interpreted to separate the various units, but amount of displacement is difficult to decipher. The difference in shortening between adjacent lithotectonic units is believed to have been taken up in incompetent units along bounding discontinuities. Examples of similar disharmonic folding related to decollements in the Canadian Cordillera is presented by Fitzgerald and Braun (1965) and Dahlstrom (1969).

On this trip we will examine the structure in the lithotectonic unit made up of Upper Silurian through lower Middle Devonian rocks, best exposed in the quarry at Andreas, Pennsylvania (Stop 10). Here, and in the surrounding region, folds have wavelengths between about 300 and 1,000 feet (90 and 300 meters). They are generally overturned with steep northwest limbs, but in places the dips of axial planes of some of the folds have been rotated from the southeast to the northwest and the overturned limbs have been rotated more than 180°. In comparison, folds in younger rocks to the north, such as in the Mahantango, Trimmers Rock, and Catskill Formations, are more open and have wavelengths of about 5 miles (8 km). Folds in the rocks beneath are likewise larger, with wavelengths that average about 1 mile (1.6 km). Northwest movement at the base of the complexly folded Upper Silurian-Middle Devonian zone is believed to have taken place in shales of the buried Poxono Island Formation or Bloomsburg Red Beds. In the area of this field trip this movement zone is buried, but at Lehigh Gap, 10 miles (16 km) to the northeast, evidence for this movement is suggested by numerous bedding slip surfaces and tectonic wedges, all indicating northwest movement of overriding beds, even on the northwest limbs of folds where one would normally expect southeast movement up towards the crests of the folds (Epstein and others, 1974). Extensive shear in the Marcellus Formation accounts for movement at the top of this lithotectonic unit. Kehle (1970) has described a similar disharmonically folded sequence as a decollement zone wherein the movement between overlying rocks and underlying rocks is taken up by shear within the zone. The shear zone in the Marcellus is recognized in wells north of the area, and, along the Northeast Extension of the Pennsylvania Turnpike, it is at least 300 feet (91 m) wide. Along the Turnpike the shear zone coincidently also marks the

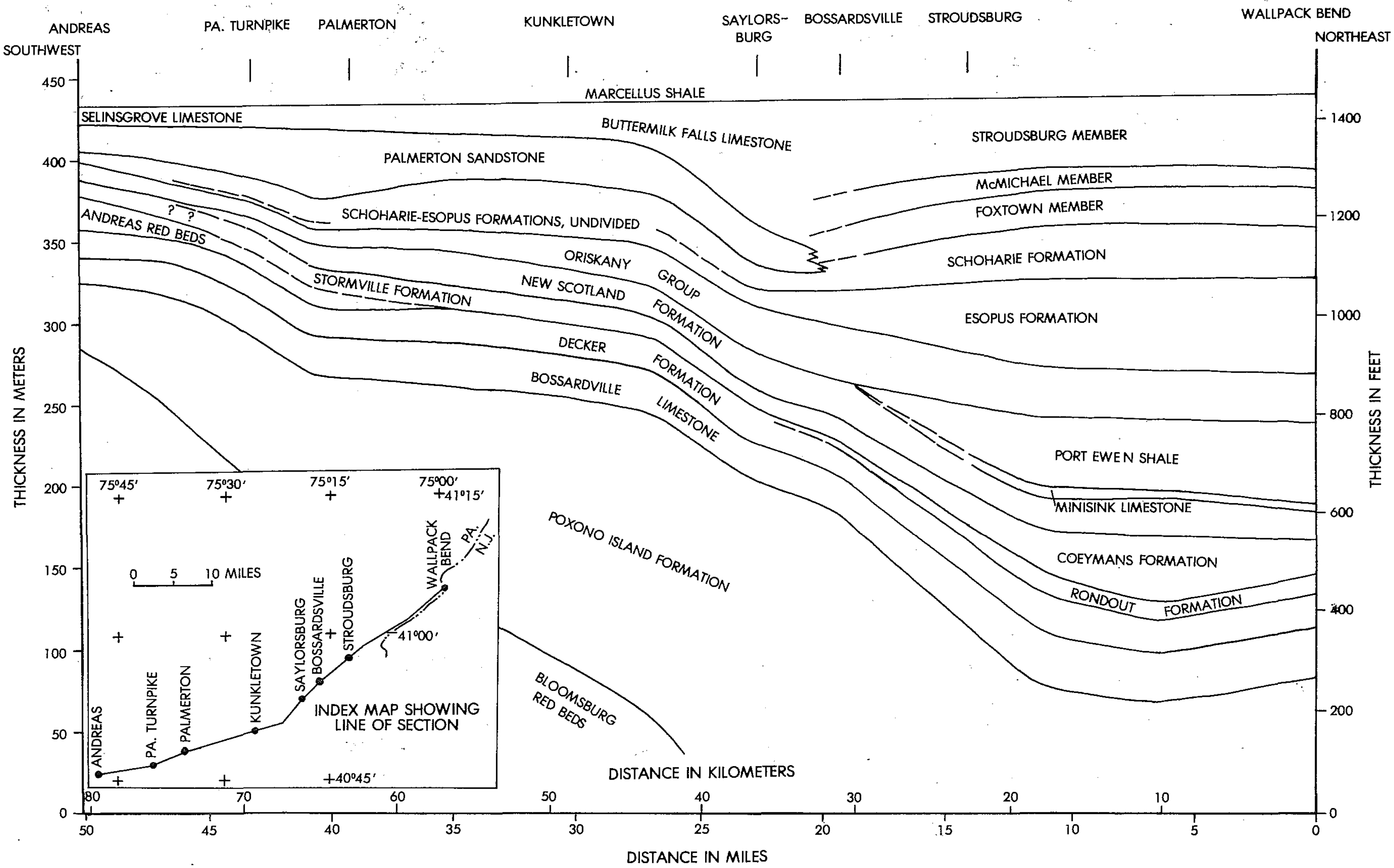


Figure 72. Stratigraphic section of Upper Silurian through lower Middle Devonian rocks comparing the sequence seen at Andreas, Pa. (STOP 10) with rocks 80 km (50 mi) northeastward in Pennsylvania.

trace of the late Sweet Arrow fault, a southeast-dipping thrust which extends for more than 80 miles in this region (Wood and Kehn, 1961; Wood and others, 1969). Mapping, which is still in progress just west of Andreas, shows that stratigraphic and structural anomalies may be partly the effects of the Sweet Arrow fault. This will be discussed at Stop 10.

An alternate explanation for the disharmonic folding is that there has been no relative movement between the lithotectonic units and that the disharmony is the result of buckling of a series of layers of different competencies or viscosities (for example, Ramsey, 1967, p. 380). Thus, no northwest movement between the lithotectonic units need have taken place. The decollement hypothesis is favored because of the evidence of northwest translation of overlying rocks as mentioned above, and because the axial planes of the disharmonic folds do not dip away from the crest of the larger enclosing folds to which they are satellitic (as shown by Ramberg, 1964, p. 322). Rather, the axial planes of the folds in the Upper Silurian-Middle Devonian lithotectonic unit dip to the southeast, except where they are rotated to the northwest.

A new exposure in the Bossardville Limestone and Decker Formation in the quarry of Herbert R. Imbt, Inc., at Bossardsville, Pennsylvania, 30 miles (48 km) northeast of Andreas suggests another way in which these disharmonic folds may have formed. Here, a thin multilayered zone is disharmonically folded in the flat northwest limb of a slightly overturned syncline (fig. 73). It is believed that crowding in the core of the syncline resulted in thickening of the axial region by buckling and thrusting of rocks out of the core into a bedding thrust (or thrusts) on the limb. This resulted in the

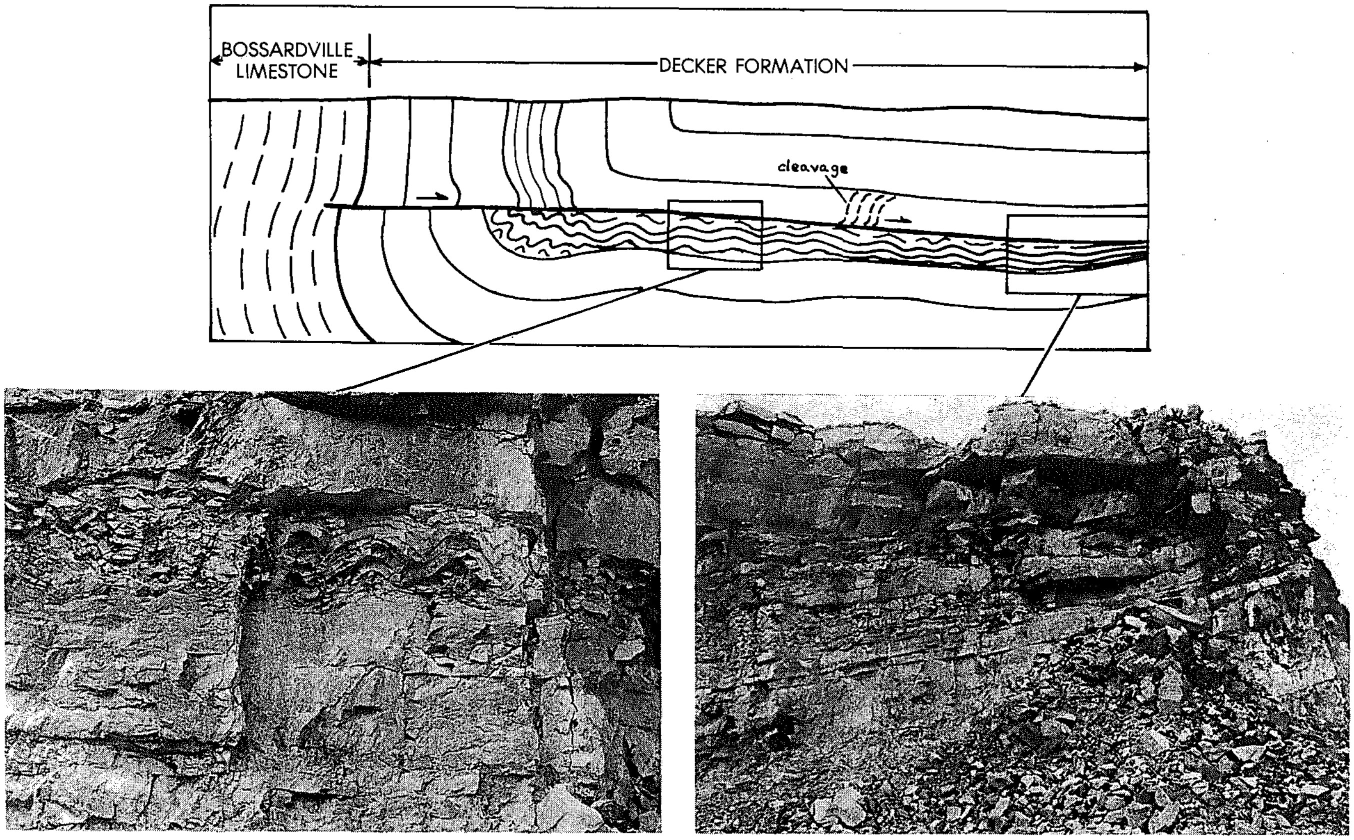


Figure 73. Small-scale disharmonic folding developed in the Decker Formation by movement along a bedding-slip fault out of the tightened core of an oversteepened syncline, quarry of Herbert R. Imbt, Inc., Bossardsville, Pa. Width of the disharmonically folded sequence of rock is 61 m (220 ft).

disharmonic folding within the Decker Formation by frictional drag above a detachment horizon. Observable offset at the core of the fold does not exceed 6 feet (2 m). The displacement decreases southeastward into the Bossardville Limestone and dies out into a series of calcite- and quartz-filled fractures. No disharmonic folding was seen higher up in the overturned limb of the fold. The folding decreases in intensity away from the core of the larger fold and dies out 220 feet (67 m) from the core of this fold. The thickness of the zone diminishes as the internal folding decreases. Evidence of northwest movement above and below the folded zone includes bedding plane slickensides and sigmoidal cleavage at and near the detachment surface.

If this quarry-sized model is applicable to the folding of Upper Silurian through Middle Devonian rocks on a regional scale, then the amount of movement need not have been great to produce the observed folding, although there may have been substantial layer-parallel shortening due to cleavage formation by pressure solution. However, there is one major difference between the disharmonic folding on the two scales. There is no folding in the southeast limb of the larger fold in the Bossardsville quarry, whereas disharmonic folds have developed within the overturned limbs of larger enveloping folds on the regional scale (see the cross sections on Plate 1 of Epstein, Sevon, and Glaeser, 1974, for example).

How do these disharmonic folds die out? About five miles (8 km) southwest of Andreas, the Upper Silurian through Middle Devonian rocks are no longer disharmonically folded, and as far as can be determined, they are sandwiched between and parallel to the beds in overlying and underlying rocks. Exposures in the area where this change takes places are extremely poor and many details are lacking. However, there are a few stratigraphic anomalies that suggest faulting between the two areas. An impure sandstone, similar to the Montebello Sandstone Member of the Mahantango Formation farther west, but which may be in the lower part of the Marcellus, suddenly appears in this area. The sandstone, however, may be in the lower part of the Marcellus Shale. Even so, its appearance is sudden. Also, the Palmerton Sandstone may not extend southwest of this area. Some other stratigraphic units appear to be out of place structurally. These anomalies are interpreted to indicate a fault separating and telescoping the disharmonically folded sequence from the unfolded sequence, as shown in the cross section of Plate 1. There is one problem with this interpretation, however. If the fault is a ramp from the underlying decollement as shown, then a ramp anticline should be seen in overlying rocks. This anticline is not present. Continued mapping in the area may shed some more light on this problem.

CONODONTS FROM THE GREENWICH SLICE OF THE HAMBURG KLIPPE NEAR GREENAWALD, PENNSYLVANIA

by John E. Repetski

Two samples collected by Avery Drake and Peter Lyttle from near Stop 4 yielded diagnostic conodonts. Both limestones were from carbonate flysch beds on the east side of Maiden Creek; one below the Reading Railroad tracks and one about 575 meters north of the tracks' crossing at Greenawald Bridge. Both faunas were relatively sparse, giving 224 identifiable specimens from a total of 11.66 kilograms of rock (Table 6), but these allow both good stratigraphic determination and paleogeographic interpretation of the host rock.

Both samples share their early Arenigian age and North Atlantic faunal province (NATP) affinities with the conodont-bearing limestone slide blocks near Lenhartsville reported by Bergström and others (1972). They differ from those previously reported Hamburg klippe faunas, however, in their species distribution and by being possibly slightly younger.

The two samples from Lenhartsville reported by Bergström and others (1972), as well as subsequent ones collected there (USGS undescribed collections), are dominated numerically or at least have significant proportions of elements belonging to the zonal index species Prioniodus elegans Pander. Those same samples contain very few elements of Paracordylodus gracilis Lindström. P. gracilis elements numerically dominate both of the flysch samples near Greenawald Bridge, and only one of those probably contains P. elegans.

Several reasons could cause this discrepancy. First, if the flysch samples are coeval with the Arenigian-age calcisiltite in the Lenhartsville slide-blocks, then the local distribution of P. elegans and P. gracilis may reflect different environmental preferences of these species, i.e., their distribution was controlled by different water mass, substrate, or food preference/tolerance. Conversely, present distribution could simply reflect hydraulic sorting of the disarticulated skeletal elements during sedimentation/deposition processes. The elements of P. elegans generally are much larger than are those of P. gracilis; samples with large numbers of P. elegans elements may have had the smaller, lighter P. gracilis elements winnowed away prior to lithification.

The sequence near Greenawald might be slightly younger than the Lenhartsville slide block faunas. The Greenawald samples contain rare specimens of Fryxellodontus? corbatoi Serpagli. This species was reported previously only from Argentina, in a sequence also containing NATP conodonts (Serpagli, 1974). In that succession, F.? corbatoi was found only within the Oepikodus smithensis (formerly evae) Zone, which immediately succeeds the P. elegans Zone in the NATP zonal scheme (see Repetski, this volume, Fig. 1). The Pennsylvania occurrences may represent a slight downward extension in the range of F.? corbatoi. Even though F.? corbatoi probably co-occurs with P. elegans at Greenawald, and definitely does so in one of the cobbles from Spitzenberg (see section "Conodonts from Spitzenberg"), these occurrences do not necessarily preclude a lower O. smithensis Zone assignment. Bergström and others (1972, Fig. 3) indicated that P. elegans ranges sporadically into the lower part of the O. smithensis Zone. Given the species present, the age of these flysch samples must be given as P. elegans to early O. smithensis Zones.

The significance of the conodonts from the flysch beds near Greenawald, besides allowing the dating and paleoenvironmental interpretation (cool and/or deep water) of their host-rocks, concerns their relationships with the conodonts from the slide blocks near Lenhartsville (Bergström and others, 1972). Epstein and others (1972) proposed that the source of the slide blocks originally lay to the east or southeast of their present position, perhaps on uplifted blocks of continental rise or on the fringe of oceanic islands within the Iapetus ocean. They argued that the depositional site of the cooler water NATP conodonts probably was quite distant from the eastern edge of the North American carbonate platform, partly because at that time (1972) no NATP conodonts were known from the warm-water facies platform carbonates. Since then, although the conodonts of the Beekmantown Group carbonates are still far from thoroughly

Table 6. Conodonts from the Greenwich slice near Greenawald, Pennsylvania.

Conodont species or element type	Number of specimens	
	Sample number USGS 9138-CO (6.26 Kg.)	Sample number USGS-9139-CO (5.4 Kg.)
cf. Acodus deltatus (Linstrom)		
? acodontiform element		1
distacodontiform el	2	1
Drepanodus arcuatus Pander		
pipaform el.		2
D. spp. s.f.	5	6
Fryxellodontus? corbatoi Serpagli	1 (cf.)	1
Oistodus cf. O. n. sp. 1 (Serpagli, 1974)		
costate el.	2	2
acostate el.	4	5
Paracordylodus gracilis Lindstrom		
paracordylodontiform el.	31	48
oistodontiform el.	8	17
cyrtoniodontiform el.	9	18
Paroistodus parallelus (Pander)		
drepanodontiform el.	1	3
oistodontiform el.	1	4
? P. sp.; drepanodontiform el.		4
Prioniodus cf. P. elegans Pander		
prioniodontiform (P) el.		1
ramiform (S) el.		1
P.? sp.; oistodontiform (M) el.		3
unassigned acodontiform el.		1
unassigned oistodontiform el	3	5
unassigned Sb ramiform el.		1
indeterminate elements	6	27

investigated, there is at least one significant co-occurrence of NATP species with North American Midcontinent Province (NAMP) (warmer water) faunal assemblage. In the southern Champlain Valley, eastern New York, Prioniodus elegans occurs in at least one level of the Fort Cassin Formation, which otherwise contains NAMP species (Repetski, 1977). John Rodgers proposed that that part of the eastern North American carbonate platform was extremely narrow during the Early Ordovician; his map gives it a width of about 35 miles (Rodgers, 1968). If conodonts of the two faunal provinces were separated chiefly by water-mass temperatures, then the NATP species could well have lived over slope or even shelf-edge environments below the level at which the thermocline impinged on the edge of the continent. Where the shelf was unusually narrow, as in the Champlain Valley regions, onshore-blowing storms or other short-term disruptions in the level of the thermocline could most easily explain this anomalous co-occurrence of NATP and NAMP species.

The example from New York does not necessarily invalidate the proposed depositional sites of the Lenhartsville blocks proposed by Epstein and others (1972), but it does remove the faunal argument against an eastward-facing continent-marginal site as the source of those blocks. The flysch near Greenawald may represent a depositional situation very similar to that source of the Lenhartsville blocks, i.e., slope sediments deposited below a thermocline, accumulating skeletons of cooler water conodont species.

As with all conodonts known thus far from the Hamburg klippe, the ones from the samples near Greenawald have a color alteration index (CAI) value of 4.5, indicating that the host-rocks reached temperatures of about 200^{o} to 300^{o} C. This thermal level contrasts sharply with that found in the nearby parautochthonous carbonate rocks, where consistent CAI values of 5 indicate that those rocks were heated to at least about 300^{o} C, or higher.

CONODONTS FROM SPITZENBERG

By John E. Repetski

Conodonts have been recovered from a number of individual limestone and dolostone cobbles from the conglomerate that forms the upper half of Spitzenberg, and they suggest that the conglomerate records the progressive erosion of nearby bedrock during the Late Ordovician or earliest Silurian. The conglomerate contains cobbles at several horizons. Most of these are of a single lithic type, that is, rounded-edged tabular clasts of medium to dark gray micrite to calcisiltite. Some of the clasts of this lithology appear structureless on hand-lens inspection. Others though, exhibit internal ripple lamination and/or grading. Thus most of the clasts resemble most closely the thin-bedded limestone in the limestone and dark gray shale "ribbon-bedded" sequences within the nearby Hamburg klippe. The limestones were described by Epstein, Epstein, and Bergström (1972) and conodonts from them, of early Arenigian (early to medial Early Ordovician) age were identified and discussed by Bergström, Epstein, and Epstein (1972). In the latter paper, the authors not only dated these klippe carbonates near the town of Lenhartsville, but also discussed the provincial differences between these conodonts and those found in the nearby, coeval, parautochthonous Beekmantown Group limestones and dolostones. The conodonts from the Hamburg klippe limestone near Lenhartsville are indicative of the North Atlantic conodont Province (NATP). The distribution of that association of species generally has been attributed to regions of cooler (and/or deeper) marine environments (e.g., Bergström, 1973; Fahraeus and Barnes, 1975). Conversely, the species found in the Beekmantown Group (e.g., Hass, in Sando, 1958; Goodwin, 1972; Tipnis and Goodwin, 1972; Repetski, 1977, 1979) are characteristic of the North American Midcontinent conodont Province (NAMP) (e.g., Sweet, and others, 1971) and generally are accepted as having been limited to regions of warmer and/or shallower-water environments. The presence of NATP species in the Hamburg klippe rocks gave Epstein and others (1972) evidence in addition to the sedimentological and structural evidence to propose that these carbonate rocks within the klippe derived from an extracratonic source considerably eastward or southeastward from their present position.

Several other lithologies are present as cobbles in the Spitzenberg conglomerate, although these are quite minor in relative and total volume within the exposed outcrops. One of these lithologies is a pinkish-gray bioclastic lime grainstone to packstone. Some of these discoidal clasts show fragments of brachiopods, including Sowerbyella sp., cf. Hesperorthis sp., and indeterminate dalmanellids (R.B. Neuman, written commun., 1978); most contain echinodermal debris. Except for the color, these clasts closely resemble some of the lithologies found in the Jacksonburg Limestone in the autochthonous sequence. Barnett (1965) described conodonts from the Jacksonburg Limestone at numerous localities; he dated these faunas as post-Blackriveran, pre-Edenian Middle Ordovician. In addition, Barnett showed that the conodonts of the Jacksonburg are typically of the North American Midcontinent Province (NAMP), but that several NATP species appeared and even dominated in rocks mapped as Jacksonburg within the so-called Clinton outlier in northern New Jersey. Those outcrops near Clinton are now included within the Jutland klippe and are considered allochthonous (e.g., Perissoratis and others, 1979). Other collections from the Jacksonburg (USGS collections) are dominated by NAMP species of Rocklandian to Kirkfieldian age (late Middle Ordovician) that inhabited relatively shallow warm water, although some of the samples from eastern outcrops yield remains of Rhodesognathus elegans (Rhodes), a NATP species that appears to have migrated cratonward during the late Middle and early Late Ordovician.

Other lithologies present as Spitzenberg cobbles are medium to light gray dolostone with fine grained textures, and very light to medium gray chert. Lithologies of these types can be found in most of the formations of Beekmantown Group and in the underlying Upper Cambrian sequence in eastern Pennsylvania.

Cobbles were collected for conodont examination during two visits to Spitzenberg. The criteria used for selecting individual cobbles included attempting to get a

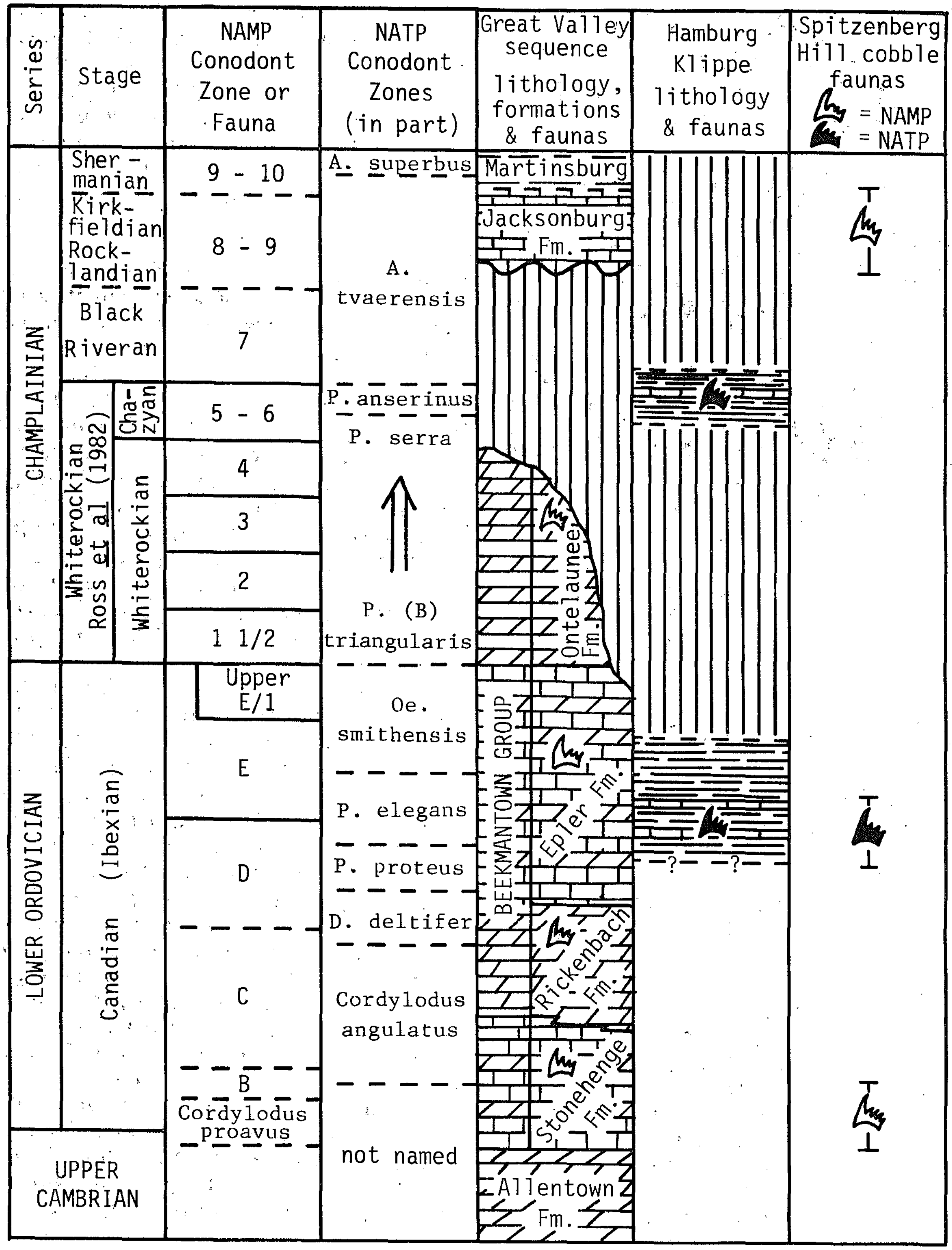


Figure 74. Stratigraphic distribution of Lower and Middle Ordovician rocks and conodont faunas in Lehigh and Berks Counties. Modified from Ross and others (1982).

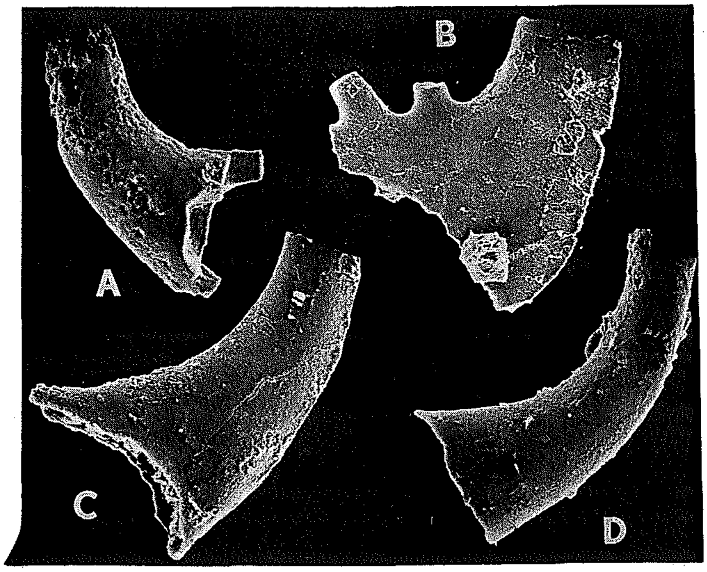


Figure 75. Uppermost Cambrian or lower Lower Ordovician conodonts from cobble sample at Spitzenberg (sample locality USGS 9481-CO). All specimens were photographed using SEM and are reposited in type collections of the Department of Paleobiology, U.S. National Museum of Natural History (USNM), Washington, D.C.

A. Hirsutodontus hirsutus Miller; left lateral view, X 224, USNM 384985.
B. Cordylodus proavus Müller; inner lateral view of rounded element, X 112, USNM 384986.
C. Eoconodontus notchpeakensis (Miller); right outer lateral view rounded element, X 210, USNM 384987.
D. Teridontus nakamurai (Nogami); right lateral view, X 280, USNM 384988.

representative sampling of the lithotypes present, but that was not completely satisfactory due to incomplete exposures, the small size of most of the cobbles, and the physical difficulties of chopping enough of any single cobble from the matrix to produce an adequate-sized unambiguous sample. Experience in searching for conodonts in the Lower Ordovician carbonates in the Appalachians has shown that most of the dolostones, as well as the Hamburg klippe ribbon limestones and calcareous dolostones, generally yield only a few to a few dozen conodont elements per kilogram of rock, if they yield any of the fossils at all. Many of these dolostones were originally deposited high in tidal flat environments, sometimes under conditions of elevated salinity, and apparently thus represent areas unfavorable to most conodont animals. The carbonates of the Hamburg klippe probably formed on a continental slope (Epstein and others, 1972), and perhaps because of cold water or some other factor(s) that environment also saw a generally slow accumulation rate of conodont skeletal material.

Some of these klippe limestones however have yielded fairly abundant collections (per kilogram of rock processed). Perhaps these accumulations resulted in areas of particularly slow carbonate sedimentation. Many of the shallow water limestones, both in the Beekmantown Group and in the Jacksonburg Limestone, also produce good to abundant collections, attesting to the conodonts' general flourishing during most of the Ordovician in warm shallow waters with normal salinity. Because of these differences in original distribution of conodonts therefore, smaller limestone cobbles often yield much better collections than do larger dolostone cobbles.

Even with the relatively small number of cobbles successfully collected and examined, a general distributional trend was noticed. The overwhelming majority of the cobbles are lithically, and the conodonts indicate faunally as well, identifiable as deeper-water limestones derived from the Hamburg klippe. All of the bioclastic grainstone cobbles lithically resembling Jacksonburg Limestone lithologies were found near the base of the exposed conglomerate sequence. Conversely, most of the very light gray dolostone and light gray chert cobbles, most resembling the lower Lower Ordovician and Upper Cambrian formations i.e., Conocheague through Rickenbach Formations, were found near the top of the hill. The ages of the conodonts from the cobbles bear out these observations. All of the Rocklandian and younger ages were derived from cobbles low on the hill, and the oldest ages came from cobbles from high in the sequence. Thus, the distribution of both lithologies and faunas are best explained by the scenario of stripping off of locally-derived coarse sediment from progressively lower formations, with the great majority of the cobbles derived from the surrounding and immediately-underlying Hamburg klippe terrane. The source-formations

Figure 76. NATP conodonts from Spitzenberg (sample USGS 9833-CO). All specimens were photographed with SEM and are reposited in type collections of the Department of Paleobiology, U.S. National Museum of Natural History, USNM.

A - F. Prioniodus elegans Pander;
A, lateral view of Sa ramiform element USNM 384989, X 165;
B, anterolateral view of P element USNM 384990, X 215;
C, inner lateral view of Sc ramiform element USNM 384991, X 190;
D, inner lateral view of M element USNM 384992, X 135;
E, lateral view of asymmetrical quadriramate S element USNM 384993, X 215;
F, lateral view of symmetrical quadriramate Sd element USNM 384994, X 215.
G, N. Prioniodus? sp.;
G, lateral view of symmetrical quadriramate S element USNM 384995, X 305;
N, lateral view of Sa ramiform element USNM 384996, X 266.
H. Fryxellodontus? corbatoi Serpagli; inner lateral view of Sb element USNM 384997, X 135.
I, O. Drepanodus arcuatus Pander; inner lateral views of two drepanodontiform elements, USNM 384998, X 150, and USNM 384999, X 115, respectively.
J, L, M. Paracordylodus gracilis Lindström; J and M, lateral views of two paracordylodontiform ramiform elements, USNM 385000, X 120, and USNM 385001, X 165, respectively; L, inner lateral view of M element USNM 385002, X 190.
K. Oelandodus elongatus (Lindström); inner lateral view of elongatiform element USNM 385003, X 150.

Figure 77. Conodonts from cobble of Jacksonburg Limestone lithology at Spitzenberg (sample USGS no. 9832-CO). All specimens were photographed with SEM and are reposited in type collections of the Department of Paleobiology, U.S. Museum of Natural History, USNM.

A, B. Phosphatized bryozoan fragment (X 35) and steinkern (X 52), USNM 385004, and 385005.
C. Paraprioniodus sp.; lateral view of ramiform (S) element USNM 385006, X 65.
D. Polyplacognathus ramosus Stauffer; upper view of central part of broken platform (P) element USNM 385007, X 87.
E. Protopanderodus sp.; lateral view, USNM 385008, X 227.
F, G. Aphelognathus politus (Hinde); posterolateral view of Sa ramiform element USNM 385009, X 210, and inner lateral view of platform (P) element USNM 385010, X 65.
H. Panderodus sp.; lateral view, USNM 385011, X 114.
I - L. Periodon cf. P. aculeatus Hadding;
I, inner lateral view of falodontiform (M) element USNM 385012, X 262;
J, inner lateral view of digyrate P element USNM 385013, X 157;
K, L, inner lateral view of two ramiform (S) elements, USNM 385014, X 210, and USNM 385015, X 227, respectively.

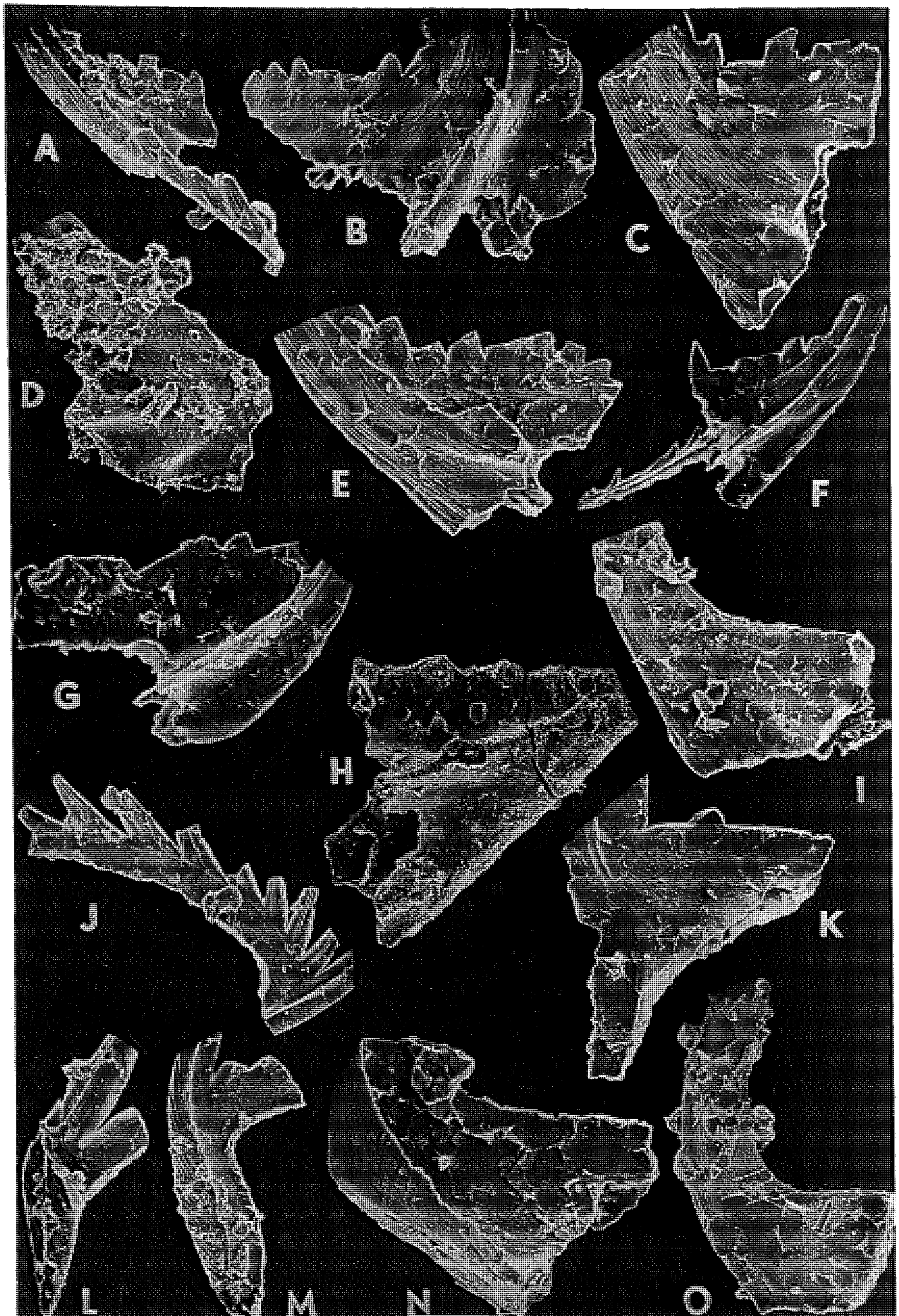


Figure 76.

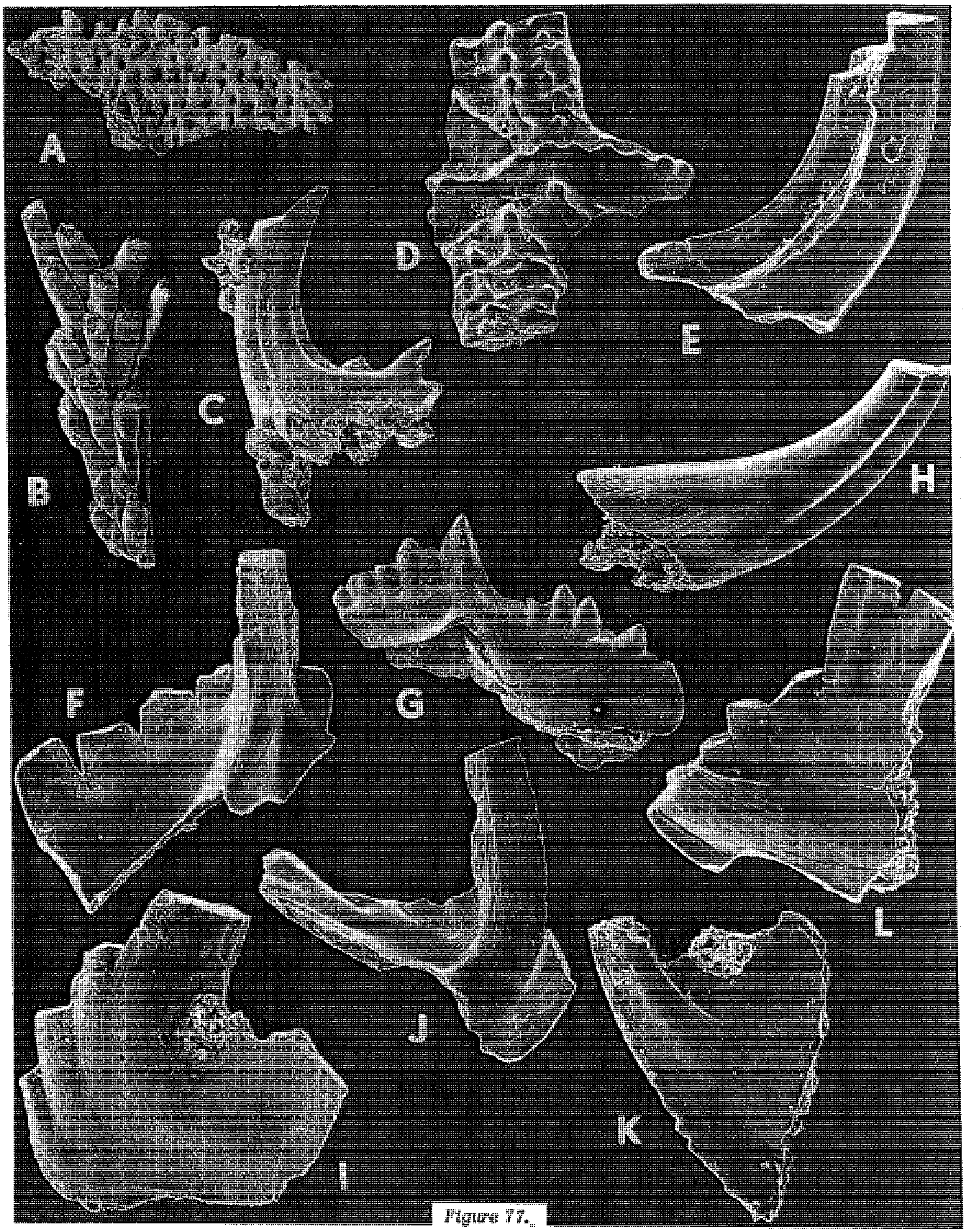


Figure 77.

Table 7. Conodonts from Spitzenberg, Stop 3.

SAMPLE NO. AND COBBLE MASS / CONODONT SPECIES	9830-CO 0.50 kg	9831-CO 0.80 kg	9832-CO 0.92 kg	9833-CO 2.0 kg	9834-CO 0.84 kg	9835-CO 0.73 kg	9836-CO 0.93 kg	9837-CO 1.90 kg	9838-CO 1.10 kg	9839-CO 1.0 kg	9840-CO 1.5 kg	9841-CO 1.2 kg
Aphelognathus sp.	X				barren sample							
A. politus		cf.	X									
Drepanodus sp.	X											
Microzarkodina sp.	X						?					
Panderodus spp.	X		X									
Paracordylodus gracilis	?+		+	X					X	X	X	
Paroistodus parallelus	+			X		X	X		X	X	X	
Plectodina sp.	X											
Polyplacognathus ramosus	cf.		X									
Prioniodus? sp.	X											
Protopanderodus sp.	X	X	X						X			
P. rectus		cf.										
Rhodesognathus elegans	X	cf.										
Paroistodus sp.		X										
Phragmodus undatus		cf.										
Drepanoistodus suberectus			X									
D. forceps			+				?		X	cf.		
Oepikodus smithensis			+									
O.? sp.			X						X	X		
Oneotodus variabilis			+									
Paroistodus proteus			+					X				
Periodon aculeatus			cf.+									
Protopanderodus gradatus			cf.+									
Paraprioniodus sp.			+									
Acodus deltatus				cf.			X	X	X	aff.	X	
Drepanodus arcuatus				X		X		X	X	X		
Oelandodus elongatus				X					X	X		
Prioniodus elegans				X		X	X		X	X		
Scalpellodus sp.				?					?			
Walliserodus australis				cf.								
Scolopodus rex						cf.						
Eucharodus parallelus						*		?*				
Glyptoconus quadraplicatus						*			*			
Scolopodus gracilis						*		*	*			
S. variabilis						*						
Ulrichodina abnormalis						*						
U. simplex						*						
Walliserodus ethingtoni							cf.					
Drepanoistodus deltifer deltifer								X				
D. sp.								X				
Paroistodus amoenus								X				
Scandodus sp.								X				
Scolopodus filosus								?*				
S. peselephantis								X				
S. spp.								X	X			
Walliserodus sp.								?				
Fryxellodontus? sp.									X			
Juanognathus sp.									?			
Triangulodus? cf. T. brevibasis									X	X		
Glyptoconus sp.									X			
Scolopodus carlae									*			
Tropodus comptus									*			
Fryxellodontus? corbatoi				X						X		
Oelandodus costatus										X		
Drepanoistodus inconstans											X	
Oistodus venustus											cf.	
Cordylodus proavus												X
Eoconodontus notchpeakensis												X
Hirsutodontus hirsutus												X
Teridontus nakamurai												X

+ = taxon reworked
* = probably transported

must have had a much wider local distribution than their limited extent seen today.

Some of the cobbles even record the occurrence of multiply-reworked or transported faunas. Three of the clasts having klippe lithology and NATP fauna also contained some specimens of NAMP species. The latter probably were redeposited in the off-shelf (slope) environments penecontemporaneously with accumulation of the NATP species specimens because there is no necessary age discrepancy among the species present. One, and possibly two, of the clasts with Jacksonburg Limestone lithology and fauna also contain specimens of Lower Ordovician conodont species typically found in the Hamburg klippe carbonates. These much older elements most likely were derived from pebbles and cobbles of the klippe carbonates that were reworked into the Jacksonburg sediments during the late Middle Ordovician. L. Savoy (1980) also found numerous instances of clasts bearing Lower Ordovician conodont elements having been incorporated into the lower part of the Jacksonburg in easternmost Pennsylvania and northern New Jersey.

Figure 74 shows diagrammatically the stratigraphic distribution of Lower and Middle Ordovician rocks and conodont faunas in Lehigh and Berks Counties. The stratigraphic chart is modified from Ross and others (1982). Conodonts from the cobbles on Spitzenberg derived from at least three general levels: 1) the Jacksonburg Limestone, dominated by late Middle Ordovician NAMP species but also containing some NATP species and occasionally some residual previously-reworked Lower Ordovician species (fig. 77); 2) middle to upper Lower Ordovician from Hamburg klippe carbonates (the dominant cobble clast-type on Spitzenberg) containing NATP species (fig. 76); and 3) uppermost Cambrian or lower Lower Ordovician dolostones from the lower part of the Beekmantown Group (fig. 75). Figures 75 through 77 illustrate some of the condonts from Spitzenberg cobbles that are typical of these three levels, and Table 7 gives the distribution of taxa found in 12 cobbles.

Acknowledgements

I thank Gary Lash, M.E. Taylor, and especially Donna Repetski for their help in finding and chopping out the sampled cobbles. W.R. Brown and Susann Braden (U.S. National Museum, SEM Lab) and Harry Mochizuki aided in specimen illustration, E.R. MacDonald processed the samples and R.T. Lierman drafted the charts; their expert help is gratefully acknowledged.

UPPER SILURIAN BIOSTRATIGRAPHY OF THE ANDREAS QUARRY, PENNSYLVANIA

By Kirk E. Denkler

Thirteen carbonate rock samples from the Poxono Island Formation, Bossardville Limestone, and Decker Formation in the Andreas quarry were collected for conodont and ostracode extraction (table 8).

No conodonts were recovered from the Poxono Island Formation nor from the Bossardville Limestone except for a few biostratigraphically unimportant elements from the uppermost Bossardville (sample I). Other than leperditiid ostracodes, the supratidal and intertidal marine environments of these rocks appear to have been incapable of supporting any significant fossil populations.

In contrast, the four samples taken from the Decker Formation yielded a well preserved conodont fauna consisting almost entirely of complete apparatuses of Ozarkodina remscheidensis remscheidensis (Ziegler). Identical monospecific collections have also been recovered from the Decker at Highland Mills, New York (Barnett, 1971), and from similar strata in the Tonoloway Limestone of West Virginia and Maryland (Denkler, Harris, and Lierman, 1983). Collections dominated by O. remscheidensis remscheidensis appear to be typical of Late Silurian shallow subtidal biofacies of the central Appalachians. In addition, the generally good to excellent preservation of the conodonts recovered from the Decker at Andreas suggests deposition in a quiet, shallow sub-tidal, possibly lagoonal environment.

The stratigraphic interval represented by the presence of O. remscheidensis remscheidensis has recently become the source of a minor debate. The standard North American Silurian conodont zonation of Klapper and Murphy (1974) gives the first occurrence of O. remscheidensis remscheidensis as uppermost Pridolian (= the base of the uppermost eosteinhornesis Zone). However, a subsequent study of the Upper Silurian conodont succession in West Virginia and Maryland by Denkler, Harris, and Lierman (1983) found that this species extended as low as the base of the Pridoli. It appears now that in Late Silurian shallow water carbonate facies such as at Andreas and elsewhere in the central Appalachians the range of O. remscheidensis remscheidensis extends through the entire Pridoli and that only in significantly deeper water is it restricted to the uppermost Pridoli. The effect of this revision is to seriously reduce the biostratigraphic utility of this taxon since it also extends well into the lower Lower Devonian . However, since the Decker elsewhere in eastern Pennsylvania and northern New Jersey has yielded Silurian corals and ostracodes, the presence of O. remscheidensis indicates that the age of the lower Decker at Andreas is also Pridolian (late Late Silurian).

Aside from O. remscheidensis remscheidensis, representatives of only two other species, neither possessing any biostratigraphic utility, were recovered in these samples: three elements of Oulodus sp. indet. from samples M and five specimens of a new species of Ozarkodina from samples J and L.

The Color Alteration Index (CAI) of this conodont collection is 4.5 indicating that the host rock reached about 250^{o} C.

Ostracodes seem to have been less affected than conodonts by the limiting factors associated with shallower marine environments and consequently attained higher levels of diversity and abundance under conditions less favorable to conodonts. Such a relationship is exhibited in the lower twenty feet of the Decker Formation where the diversity, and to a lesser extent, the abundance of ostracodes exceeds that of the conodonts. Like the conodonts, however, they are essentially restricted to this unit except for occasional leperditiids in the Bossardville.

The two lowest samples from the Decker, J and K, yielded only poorly preserved phosphatic steinkerns of Kloedenella sp., Eukloedenella? sp. and Dizygopleura sp. indicting an association characteristic of a shallow water lagoonal environment. The next highest sample, L, however, yielded a diverse association including Leperditia elongata Weller, L. altoides Weller, Zygobeyrichia sp. cf. Z. ventripunctata Ulrich and Bassler, Kloedeniopsis sp. cf. K. hartnageli Berdan, Saccarchites sp , Kloedenella sp. and Dizygopleura sp. Leperditia elongata is a long-

Table 8. Samples collected for paleontologic analysis at Andreas, Pa.

Field No. (An9-16-83-)	*USGS No.*	*Stratigraphic position*	*Lithology*
M	*10851-SD*	*20.0' above base of Decker Formation*	*Slightly silty, shaly medium-bedded wackestone*
L	*10850-SD*	*7.8' above base of Decker Formation*	*Ostracode rich massive wackestone*
K	*10849-SD*	*6.3' above base of Decker Formation*	*Interbedded shale and nodular wackestone*
J	*10848-SD*	*1.0' above base of Decker Formation*	*Thin to medium-bedded bioturbated shaly wackestone*
I		*66.9' above base of Bossardville Ls.*	*Shaly laminated lime mudstone*
H		*48.1' above base of Bossardville Ls.*	*Laminated to thin-bedded massive lime mudstone*
G		*6.3' above base of Bossardville Ls.*	*Massive lime mudstone*
F		*16.0' below top of Poxono Island Fm*	*Massive lime mudstone*
E		*31.0' below top of Poxono Island Fm*	*Laminated shaly calcareous dolomudstone*
D		*69.0' below top of Poxono Island Fm*	*Interbedded dolomitic and calcareous mudstone*
C		*88.0' below top of Poxono Island Fm*	*Interbedded to interlaminated lime mudstone and shaly dolomudstone*
B		*93.0' below top of Poxono Island Fm*	*Interbedded to interlaminated lime mudstone and shaly dolomudstone*
A		*103' below top of Poxono Island Fm*	*Interbedded calcareous mudstone and dolomitic shaly lime mudstone*

ranging species originally described from the Rondout Formation but with occurrences reported as low as the Wills Creek Formation, while L. altoides was described from the lower part of the Decker. The specimens of Zygobeyrichia are very similar to Z. ventripunctata common in the upper member of the Tonoloway Limestone although smaller and more finely reticulate. K. hartnageli was described from the Cobleskill Limestone of New York. This sample, containing the highest levels of diversity and abundance for both ostracodes and conodonts, apparently represents the deepest subtidal environment, although probably still lagoonal, found at Andreas. Two samples collected between L and the highest conodont sample M at 17' and 18' above the base, respectively, yielded poorly preserved specimens of ?Leperditia elongata and ?Kloedeniopsis barretti (Weller), the latter having been described from the upper Decker, the Cobleskill Limestone, and from the Eccentricostra Zone of the Decker. Ostracodes from the highest conodont sample (M) at 20 feet above the base of the Decker, although poorly preserved pyritized steinkerns, provide perhaps the most useful biostratigraphic data. The presence of Bolbiprimitia? sp., although somewhat questionable, is supported by well preserved specimens of the same genus recovered from the same stratigraphic position from a similar section at the abandoned Delago quarry near Schuylkill Haven, Pennsylvania, 21 miles (34 km) to the southwest. Both Bolbiprimitia and Z. ventripunctata are restricted to the upper member of the Tonoloway Limestone of Maryland.

Above sample M the effects of emergence become more pronounced and the section rapidly becomes dominated by clastics and eventually passes into Andreas Red Beds of Swartz and Swartz (1941). At the base of these red beds, 59 feet above the base of the Decker, a distorted impression of Leperditia? sp., Kloedeniopsis? sp., and Dizygopleura? sp. and one poorly preserved pholadomyoid pelecypod Grammysia? sp. were collected.

Apart from conodonts and ostracodes, only scattered traces of brachiopods and bryozoans were observed. A few generically indeterminate byssate pteriomorph pelecypods were recovered from a sample 18 feet above the base of the Decker Formation, and ichthyoliths, including thelodont scales, were quite common in the conodont bearing-samples.

In conclusion, the Upper Silurian rocks at Andreas have a severely reduced fossil population due to enviromental control. However, the presence of a form very similar to Zygobeyrichia ventripunctata and Bolbiprimitia? sp. in the lower twenty feet of the Decker, as well as the presence of probable Pridolian conodonts, allow for a correlation of this unit with the upper member of the Tonoloway Limestone in Maryland. Detailed biostratigraphic studies of this latter unit have established that the base of the Pridoli lies at the base of the middle member of the Tonoloway and that the top of the Silurian lies near the top of the overlying Keyser Formation (Helfrich, 1978). Thus, the upper member of the Tonoloway and consequently the lower twenty feet of the Decker are most probably of middle Pridolian age. The age of the overlying Andreas Red Beds is not constrained by these identifications. It could be Late Silurian or Early Devonian in age.

The assistance of Jean Berdan and John Pojeta, Jr., in the identification of ostracodes and pelecypods, respectively, is gratefully acknowledged.

ROAD LOG AND STOP DESCRIPTIONS

DAY 1

Total Miles	
0.0	Leave parking lot of Reading Motor Inn, making left turn on Park Road. The motel is on the Allentown Dolomite of Cambrian age.
	0.2
0.2	Make right turn at traffic light on US 422 E-US 222 S.
	0.1
0.3	Continue straight ahead on US 222 north.
	0.6
0.9	Make right turn on PA 183.
	0.1
1.0	Follow PA 183 to left.
	0.1
1.1	Traffic light. Make left on PA 183.
	0.2
1.3.	Junction with US 422. Continue straight ahead on PA 183 north.
	1.2
2.5	Reading Airport on right. We are still on Allentown Dolomite.
	0.7
3.2	Descend hill crossing from Allentown Dolomite to rocks of the Beekmantown Group.
	0.7
3.9	Crossing contact of Ontelaunee Formation.
	0.1
4.0	Cross over US 222 and continue straight ahead on PA 183.
	1.0
5.0	Outcrop of Jacksonburg Limestone on right.
	0.2
5.2	Beautiful stone house on right.
	0.2
5.4	Entering rocks of the Hamburg klippe.
	0.1
5.5	Large wind generator on right.
	1.5
7.0	Cross Plum Creek. Outcrop immediately north on the left are sandstones of the Hamburg klippe.
	1.5
8.5	Crossing reservoir.
	0.1
8.6	Red and green shales on right within the Hamburg klippe.
	0.2
8.8	More red and green shales within the Hamburg klippe.
	0.2
9.0	More shales within the Hamburg klippe.
	0.3
9.3	Limestone pebble conglomerates on right.
	0.9
10.2	Red pelagic shales on right.
	0.2
10.4	Cross arm of reservoir again.
	1.1
11.5	Folded shales and sandstones on right.
	0.1
11.6	Red pelagic shales on right.
	0.3
11.9	Disrupted sandstone and shale on right (broken formation). This was Stop 10 of the Pennsylvania Field Conference in 1982.
	0.3
12.2	Town of Bernville
	0.6
12.8	Make right turn on LRO6017 (Shartlesville Road). Continue straight on road through intersection.
	0.1
12.9	Penn Township Consolidated School on left.
	0.8
13.7	Turn left onto side road. Proceed 0.2 miles to end of side road.
	0.2
13.9	Stop 1.

Stop 1. Onyx Cave Member: Sedimentologic Characteristics

Stop 1 displays the major sedimentary characteristics of the Onyx Cave Member of the Virginville Formation. In particular, we will have a chance to look at vertical facies changes related to the progressive abandonment of a submarine channel (fig. 78; Mutti and Ghibaudo, 1972; Martini and Sagri, 1977). Start at the north end and east side of the exposure and

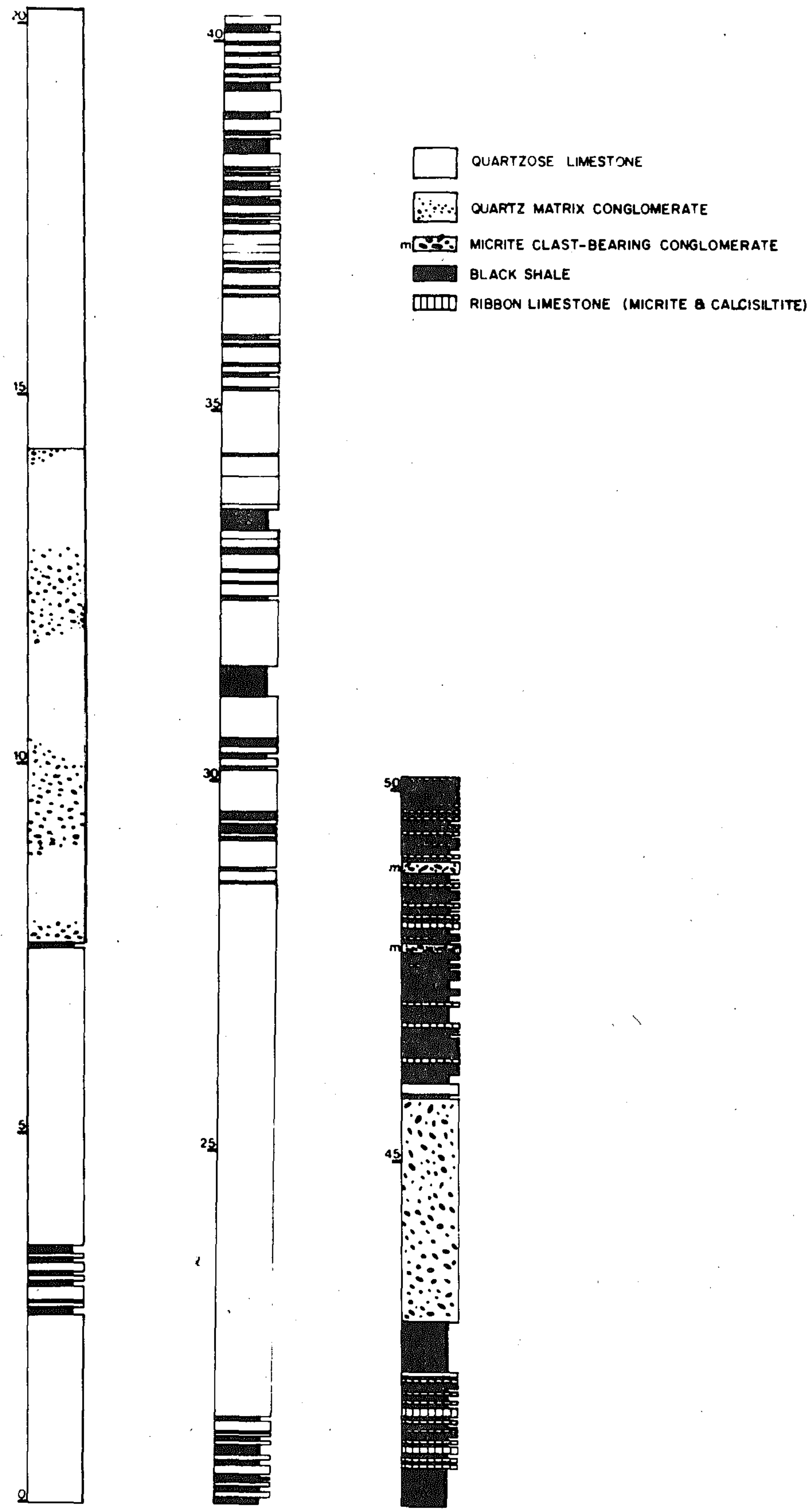


Figure 78. Lithologic log for Stop 1 showing alternations of quartzose limestone, quartz-matrix conglomerate, micrite clast-bearing conglomerate, black shale, and ribbon limestone. Thicknesses in meters.

move south (up-section) referring to Figure 78. Note that beds at this exposure dip gently to the southeast and are cut by joints that are inclined steeply to the northwest.

The lower part of the sequence is characterized by thick to very thick beds of quartzose limestone, some of which appear to be amalgamated. Packages of these very thick beds are separated from each other by thin-bedded, graded and laminated quartzose limestone beds and black shale (e.g., 3 m, 7.5 m, 21 m). Some of the thick quartzose limestone beds contain conglomeratic intervals. Pebbles include quartzose limestone and significantly fewer micrite clasts. The fact that clasts are lithologically similar to associated beds suggests local derivation rather than long-distance transport. Quartz grains in the conglomerate and quartzose limestone beds are well rounded, sometimes frosted (V-indentation pits), and well sorted, suggesting derivation from a nearshore beach or dune environment (Filock and Lash, 1984). Beds in this part of the section can be classified as Facies A and B of the Mutti and Ricci-Lucchi (1975) classification.

About 29 m above the base of the section the very thick, locally conglomeratic quartzose limestone gives way to medium- to thick-bedded graded and laminated quartzose limestone beds. These beds display characteristics of classical top-missing turbidites and can be classified as Facies B and C beds (Mutti and Ricci-Lucchi, 1975). None of these beds exceed 1 m in thickness and interbedded black shale is much more abundant in this part of the section. Although not a common feature, some of these beds are amalgamated (e.g., 34 m, 37 m). Beds become finer grained and noticeably thinner up-section (fig. 78), but are compositionally identical to underlying coarser-grained beds. About 41 m above the base of the section the thin-bedded quartzose limestone beds give way to thin-bedded (typically thinner than the underlying quartzose limestone beds) micritic limestone. These beds are massive to ripple-laminated and locally contorted and can be classified as Facies D beds of the Mutti and Ricci-Lucchi (1975) classification. About 43 m above the base of the section is a 3 m-thick limestone-clast conglomerate bed. Although it is similar to those at the base of the section, it contains significantly more and larger clasts. In addition, micritic clasts are much more abundant. This bed is overlain by ribbon limestone suggesting that deposition of the conglomerate interrupted ribbon limestone sedimentation at this site. There are two conspicuous conglomerates in this part of the sequence (48 m and 49 m). Both are less than 20 cm thick and the dominant pebble lithology is micritic limestone. Penecontemporaneous deformation of these beds evidenced by clasts that project into and deform overlying beds (fig. 78), is related to loading and squeezing of the conglomerate beds and suggests that they were water-rich when deformed. Sedimentary boudinage is common to the ribbon limestone beds and probably reflects differential compaction of limestone and black shale. Indeed, many of the boudins have tapered ends which suggest that compaction took place shortly after sedimentation (e.g., Ramsay, 1967).

The section exposed at Stop 1 illustrates a fining- and thinning-upward sequence that is probably related to avulsion and filling of a submarine channel (e.g., Mutti and Ricci-Lucchi, 1975; Ricci-Lucchi, 1975; Pickering, 1982). In a simplistic sense the very thick beds between the base and 29 m represent beds that plugged the channel. Study of Figure 78, however, suggests a scenario somewhat more complex than simple plugging of a channel by a single or multiple flows. Three small thinning-upward cycles that give rise to a composite thinning-up sequence evident in the section (fig. 78) can be seen within the lower 30 m of the section. The composite thinning-upward sequence suggests plugging of the channel by successive fills, its cross-section becomes wider and thalwegs of flows migrate laterally resulting in the shingling of thalweg fill deposits (e.g., Martini and Sagri, 1977). In this respect each of the thinning-upward cycles represents the filling and lateral migration of a thalweg within the channel. This apparently occurred a minimum of three times within the life of the channel under study here. The last major flow (between 22 and 29 m) probably signaled development of gradients too low to support thalweg flow, resulting in filling of the channel. The thin- to thick-quartzose limestone beds between 29 and 40.5 m are probably

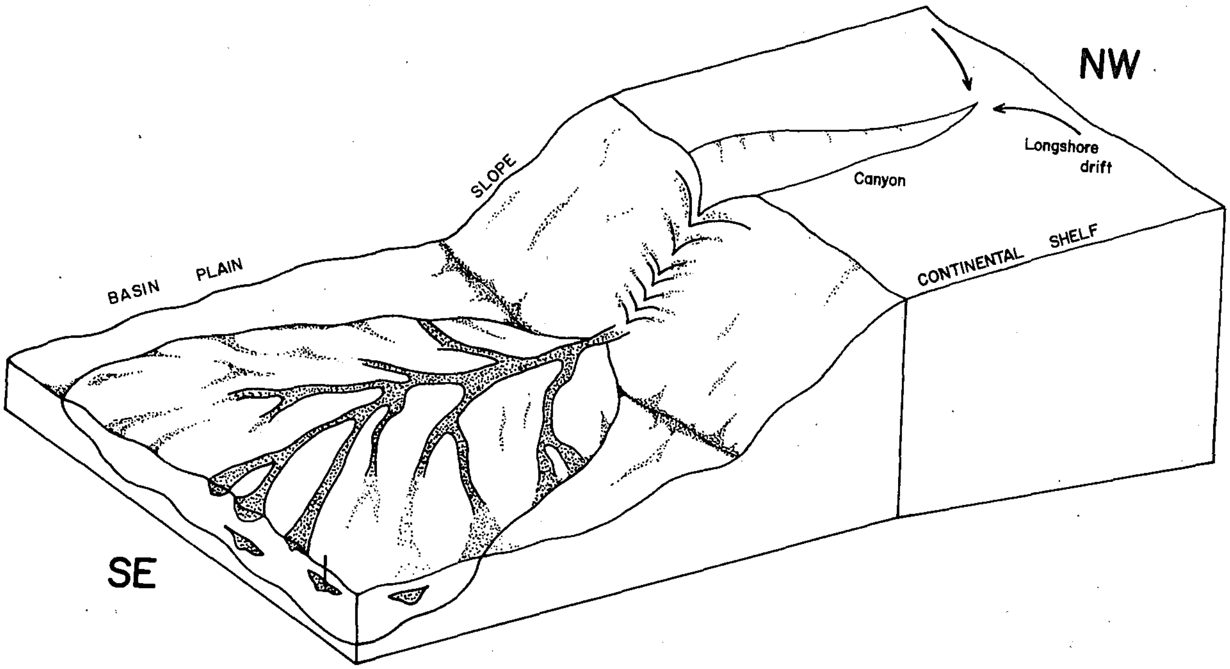


Figure 79. Conceptual deep-sea fan model for the Onyx Cave Member showing abundant distributary channels and lack of differentiation of middle and outer fan areas (modified from Link and Nilson, 1980).

overbank deposits derived from overflow of an adjacent channel. The change from quartzose limestone to ribbon limestone (at about 41 m) marks an important change that may tell of (1) a change in sediment source area, or (2) morphologic changes in this part of the basin caused by channel avulsion. Indeed, the sedimentologic characteristics of the ribbon limestone deposits are similar to those of modern levee deposits associated with clastic submarine channels (e.g., Pickering, 1982). The transition from fine-grained quartzose limestone to micritic limestone may mark the point at which the adjacent channel developed banks that were able to restrict the coarse material to the fines of the channel and allow only the finest material, the lime mud, to escape over bank tops and build levees. The levees eventually migrated over the old channel location giving rise to the section we have been looking at. The thick conglomerate bed 43 m above the base of the section is a levee breach deposit. This is consistent with its stratigraphic position within levee deposits. The significantly thinner conglomerates at 48 and 49 m probably resulted from local slumping on levee flanks, a scenario supported by the predominance of locally derived micrite clasts.

The sedimentology of the Onyx Cave Member is consistent with sedimentation on a submarine fan. In particular, the coarse-grained nature of the unit and the apparent instability of the channels as illustrated at this exposure is suggestive of sedimentation on a small sand-rich fan similar to those off southern California (fig. 79; e.g., Normark, 1978). Although paleocurrent indicators are lacking, the Onyx Cave was probably deposited at the North American shelf edge. Conodont collections from exposures of the Onyx Cave east of this exposure yield a Late Cambrian age indicting that it is coeval with the Allentown Dolomite, a platform sequence. The abundant round, well-sorted quartz sand was apparently funnelled from near-shore or littoral areas, possibly by long-shore currents into the head of the submarine canyon that fed the part of the fan seen at this stop (fig. 79). Refer to the text section "Depositional model for the Virginville Formation" for more details on Onyx Cave sedimentation.

Return to buses, make right turn on Shartlesville Road.

0.2

14.1 Pass through roadcuts of Stop 1.

0.2

14.3 Make left turn on Irish Creek Road.

The entire route between Stops 1 and 2 will be on rocks of the Hamburg klippe.

1.9

16.2 Road intersection. Continue straight on Irish Creek Road.

0.1

16.3 Intersection with Molasses Hill Road. Continue straight on Irish Creek Road.

0.2

16.5 Intersection with Scull Hill Loop. Continue straight on Irish Creek Road.

0.4

16.9 Cross under power line.

2.0

18.9 Pretty farm on left.

0.4

19.3 "T" in road. Continue left heading toward Centerport.

0.2

19.5 Thin bedded siltstone and shale in the Hamburg klippe on right. Irish Creek on left.

0.8

20.3 Cross Irish Creek

0.8

21.1 Village of Centerport. Stop sign, make right turn. Make right turn on Main Street.

0.05

21.15 Fork in road; bear left heading toward Mohrsville and Shoemakersville.

0.35

21.5 Cross center township boundary line.

0.2

21.7 Make left turn on Shoey Road heading toward Shoemakersville.

0.8

22.5 Precambrian rocks hold up the Reading Prong in the skyline to right. Blue Mountain underlain by quartzites of the Tuscarora Sandstone on left. The conical hill straight ahead is Spitzenberg which will be Stop 3 today.

1.0

23.5 Just before ConRail Railroad Tracks pull into parking lot of Van Mar Feeds Inc., distributor of farmers premium fertilizer. Disembark and walk north along railroad tracks for 2000 feet to Stop 2. Be careful! Trains still use these tracks.

Stop 2. Moselem Member: Sedimentologic Characteristics

CAUTION: TRAINS OCCASIONALLY USE THESE TRACKS.

Stop 2 illustrates the distinctive characteristics of the Moselem Member of the Virginville Formation and will serve to distinguish it from the Onyx Cave Member seen at Stop 1. Conodont collections from the Moselem Member to the south of this exposure yield Middle Ordovician ages (J. E. Repetski, pers. commun., 1983), significantly younger than the Onyx Cave Member. The Moselem is somewhat similar to the Onyx Cave Member in that both units contain ribbon limestone and black shale but that is where the similarity ends. Major differences include the lack of thick quartzose limestone and thick coarse-matrix conglomerate beds and the abundance of black shale in the Moselem Member. Although ribbon limestone is common to both units it is the predominant carbonate lithology of the Moselem Member. Note that the ribbon limestone beds at this exposure occur as distinct bundles separated by black shale. Interbedded with the ribbon limestone and black shale are irregular calcisiltite laminations that are similar to modern contour current deposits recovered from continental margin areas. The presence of grading in some of the more continuous laminations, however, is suggestive of deposition from silt turbidites (e.g., Piper, 1978). Analysis of polished and stained slabs of ribbon limestone beds, many of which appear massive in the field, suggests that most were deposited from dilute turbidity currents. These beds which are typically flat and ripple laminated are identical to Facies D beds of Mutti and Ricci-Lucchi (1975).

A conspicuous feature of the Moselem Member at this exposure and most others is the abundance of folding and boudinage. Most deformation at this exposure, particularly that illustrated by the ribbon limestone, is probably soft-sediment in origin. Much of the deformation appears to be restricted to specific zones or horizons characterized by various degrees of boudinage and folding. In some cases folds have been completely rotated or rolled into ball-like masses of limestone. Note that the majority of boudins are characterized by tapered ends suggesting that they were not well lithified when deformed. Much of this deformation is apparently related to submarine sliding. Slip-line analysis of slump folds from this exposure and others in the Moselem Member suggests that the submarine sliding reponsible for the folding was directed to the northwest (fig. 80). This can be explained by one of two models: (1) slumping on a submarine slope inclined to the northwest (fig. 81A), or (2) sliding to the northwest in response to southeast-directed growth faulting (fig. 81B). The Moselem Member was deposited on a subsiding passive margin in Middle Ordovician time concomitant with uplift of the Beekmantown shelf to the northwest (refer to text section "Sedimentologic Evolution of the Richmond Slice"). The fact that the ribbon limestone deposits are arranged in bundles suggests that the turbidites were emplaced at specific times whereas other periods were characterized by accumulation of black shale. The major control of the turbidite limestone sedimentation was probably tectonic instability and seismic generation of turbidites in the source area, presumably in the eroding North American carbonate platform. A modern analogue of Moselem sedimentation is carbonate sedimentation on the Sahul Shelf of northern Australia which is currently subsiding into the Timor Foredeep along numerous normal fault (Veevers, 1971).

Return to bus, turn right on Shoey Road, cross railroad tracks heading into Shoemakersville.

0.2

23.7 Cross Schuylkill River

0.1

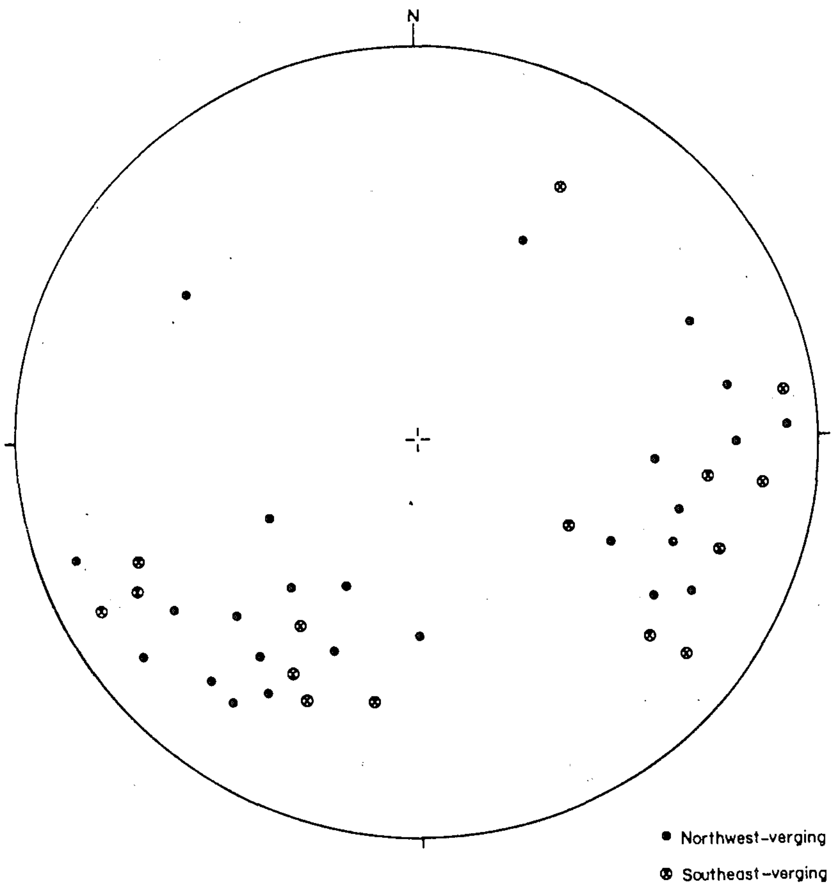


Figure 80. Vergence of slump folds in the Moselem Member.

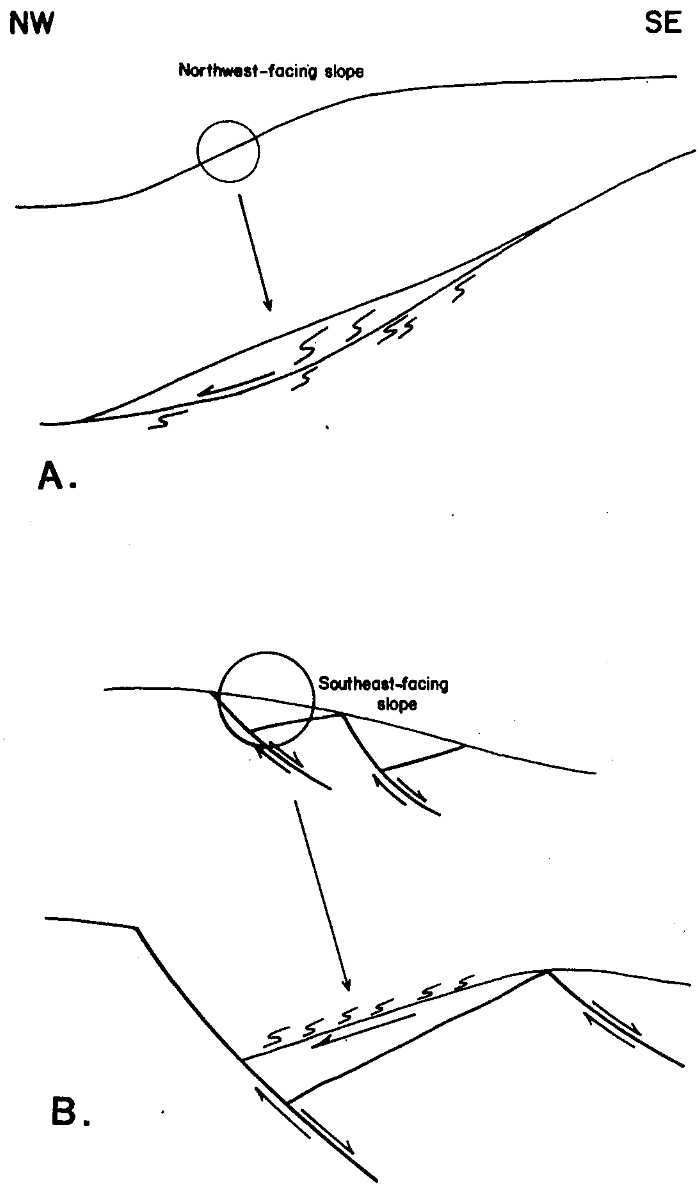


Figure 81. Possible explanations of northwest-directed slump folds of the Moselem Member.

23.8 Village of Shoemakersville. Make left turn at heading toward PA 61. Make left turn on Main Street.
0.1
23.9 Globe Underwear Co. Inc. on left.
0.1
24.0 Cross the Reading Railroad tracks. (Do not pass GO, do not collect $200.)
0.1
24.1 Make right turn on Noble Avenue
0.4
24.5 Make left turn on PA 61 (Pottsville Pike) heading north.
0.3
24.8 Blue Mountain straight ahead.
1.6
26.4 Overturned flysch of the Windsor Township Formation.
0.2
26.6 Right side up flysch of the Windsor Township Formation.
0.3
26.9 Famous Schmeck's Family Restaurant on left.
0.4
27.3 Town of Hamburg. Continue north on PA 61 north.
0.3
27.6 Schuylkill Gap straight ahead. Cross railroad tracks.
0.8
28.4 Cross Schuylkill River.
0.3
28.7 Make right turn onto Interstate 78 east towards Allentown.
0.4
29.1 Cross Schuylkill River.
0.6
29.7 Continue straight on I-78. Outcrops at the Hamburg exit are in shales of the Windsor Township Formation.
0.5
30.2 Hamburg Sanatorium smokestack on right. At present it is a mental institution. It once was a tuberculosis institution where you could acquire TB.
2.1
32.3 Sharps Mountain at 10 o'clock, underlain by the Tuscarora Sandstone. Spitzenberg at 11 o'clock, the site of Stop 3.
1.2
33.5 Cross into Greenwich Township.
0.1
33.6 Inverted sandstone and shale of the Windsor Township Formation.
0.4
34.0 Same rocks but right-side-up.
0.6
34.6 Red pelagic shale exposed on left.
0.1
34.7 Exit to right to PA 143.
0.2
34.9 Stop sign. Make left turn on PA 143 north. Maiden Creek on right.
0.5

35.4 Sandstones of the Windsor Township Formation.
0.1
35.5 Red shale and sandstones of the Windsor Township Formation on left.
0.1
35.6 Thick-bedded sandstone of the Windsor Township Formation on left.
0.2
35.8 Zettlemoyers Bridge on right. Sandstone and shale with abundant sole marks in the Windsor Township Formation on left. These rocks are in an overturned syncline whose axial plain dips to the southeast. There is also abundant tectonic wedging of the overturned limb.
0.5
36.3 Overturned sandstone and shale of the Windsor Township Formation on left, with abundant load casts.
0.3
36.6 Right side up rocks of the same.
0.1
36.7 Spitzenberg straight ahead.
0.5
37.2 Red and gray shale grading up into sandstone of the Windsor Township Formation on left. Entire sequence inverted.
0.4
37.6 Local winery on left.
0.05
37.65 Turn right.
0.05
37.7 Cross Maiden Creek.
0.05
37.75 Turn right and cross Greenawald Bridge. Turn right onto Spitzenberg Road.
0.05
37.8 Abandoned red shale quarry formerly quarried for paint pigment to left.
0.4
38.2 Steeply dipping rocks of the Windsor Township Formation on left. The gently dipping Spitzenberg Conglomerate caps Spitzenberg Hill to left.
0.4
38.6 Turn left on Nursery Road.
0.1
38.7 "Quack". Duck Crossing.
0.1
38.8 More sandstone of the Windsor Township Formation exposed on right.
0.5
39.3 Turn left onto dirt road of farm.
0.2
39.5 Stop 3. Park in farm courtyard. Ascend slope of cow pasture to north.

Stop 3. Spitzenberg and Lunch

At Stop 3 we will have a chance to study the lithologic characteristics of the Spitzenberg Conglomerate exposed at Spitzenberg and to consider timing of deformation in this area. A well exposed sequence of conglomerate and sandstone beds at the east side of the hill displays a number of important sedimentologic features of the Spitzenberg Conglomerate. Note that conglomerate beds generally lack well defined primary structures or grading. Also take note of variations in the amount of matrix relative to clasts. The conglomerate beds are continuous and were probably deposited by sheetflood on an alluvial fan rather than braided fluvial transport (refer to section titled "Spitzenberg Conglomerate" for detailed discussion). Clasts include laminated calcisiltite and dolosiltite, chert, and red sandstone. Conodonts from limestone clasts collected at this exposure indicate that these deposits were derived from Lower Ordovician shelf carbonate units to the south rather than from immediately underlying allochthonous carbonate rocks. Interbedded with the conglomerate beds are trough and tabular cross-bedded sandstone beds. The sand is typically coarse-grained but does contain some gravel layers which may be lag deposits. The sand in these beds is compositionally similar to the matrix of the conglomerates.

The Spitzenberg Conglomerate is gently inclined to the northwest. Stratigraphically beneath these rocks are rocks of the Greenwich slice that include the Windsor Township Formation and red shale units. These rocks dip steeply to the southwest (fig. 82), and are separated from the Spitzenberg by a profound angular unconformity (e.g., Willard and Cleaves, 1939). This relation is particularly important in differentiating Taconic orogenesis that deformed the allochthon, from post-Taconic deformation in this area. The local occurrence

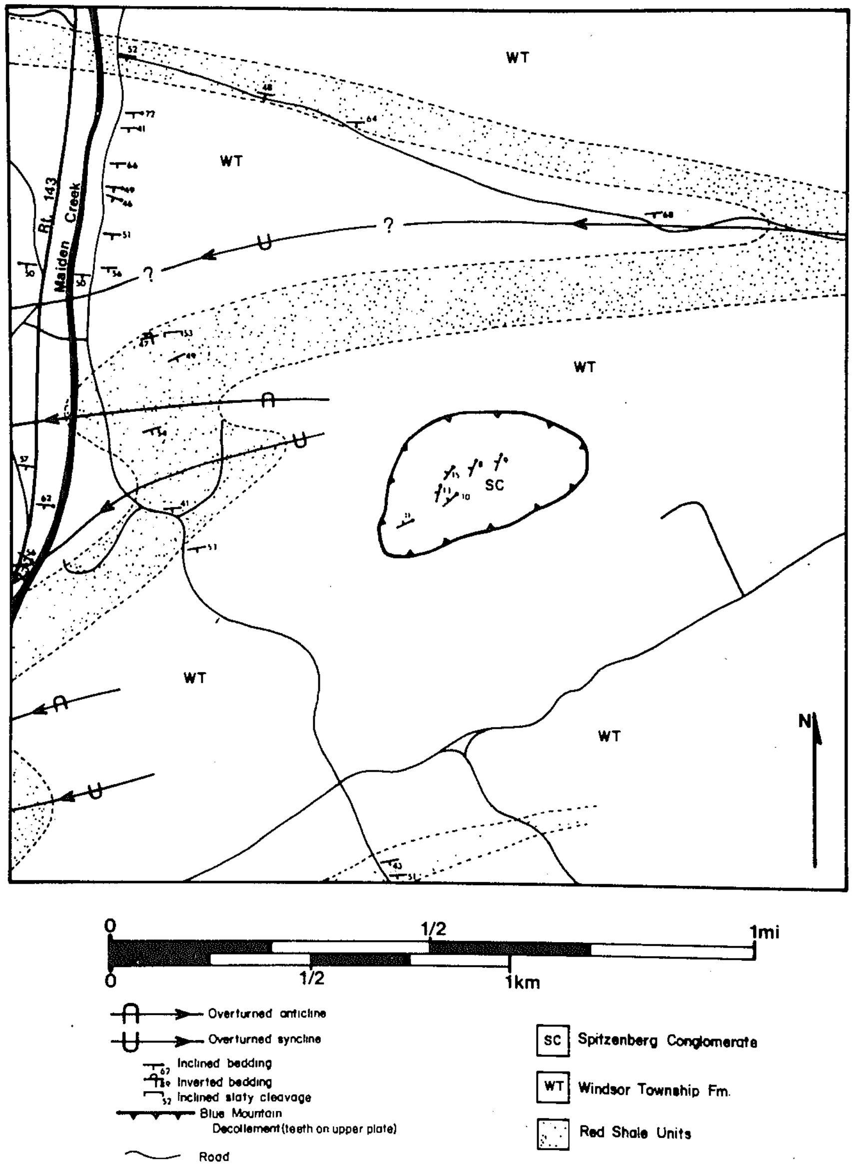


Figure 82. Geologic map of Spitzenberg and surrounding area (from Lash, 1980).

of slaty cleavage that is axial planar to tight folds in the Greenwich slice suggests that the pre-Late Ordovician deformation event (Taconic) may have resulted in formation of slaty cleavage in the parautochthonous Martinsburg Formation to the east and south. Indeed, some workers (e.g., Drake and others, 1960; Maxwell, 1962) maintain that slaty cleavage in the Martinsburg formed during the Taconic orogeny whereas others (e.g., Epstein and Epstein, 1969; Lash, 1978) proposed that it is Alleghenian in age. Work in the area around Spitzenberg (Lash, 1980) suggests that slaty cleavage here is a Taconic feature (pre-Spitzenberg Conglomerate). Recently, Drake and Lyttle (1980) suggested that concepts of cleavage formation discussed by Mitra and Elliot (1980) may explain the apparent age diachroneity of slaty cleavage in eastern Pennsylvania and western New Jersey. According to Mitra and Elliot (1980) penetrative cleavage formation within an orogen may migrate progressively toward the foreland during a continuum of deformation. Additional work is required before this mechanism can be fully evaluated in this area. It is, however, apparent that slaty cleavage formation in eastern Pennsylvania may not be simultaneous throughout the extent of the Great Valley.

Retrace route out of farm.
0.3

39.8 Intersection with Nursery Road, turn right.

0.6

40.4 Caution. Duck Crossing

0.1

40.5 Intersection with Spitzenberg Road. Turn right.

0.2

40.7 View of Sharps Mountain and the Pinnacle straight ahead.

0.6

41.3 Turn left and cross Greenawald Bridge.

0.1

41.4 Turn right on PA 143 north.

0.3

41.7 Hahn Mountain Ski area down road to left.

0.3

42.0 Beginning of exposures of Stop 4. Continue straight.

0.4

42.4 Stop 4. Park in field to right.

Stop 4. Deformation of the Greenwich Slice and Convergent Margin Sedimentation

CAUTION: ROUTE 143 IS HEAVILY TRAVELLED AND DANGEROUS TO ALL LIVING THINGS. EXERCISE EXTREME CAUTION AT THESE EXPOSURES.

Stop 4 consists of two exposures along Route 143 that display numerous structural and sedimentologic characteristics that have been documented from detailed studies of active convergent margins around the world. The north exposure illustrates the style of deformation of the Greenwich slice and the important influence of vertical variations in pore fluid content on deformation of sediment undergoing accretion at a convergent margin. The style of deformation displayed here can be seen in other exposures of the Greenwich slice and has been noted in lithologically similar rocks as far west as Harrisburg. It should be noted that although there is more than one phase of deformation illustrated at this stop, our discussion will be limited to the mechanism of formation of the earliest features. Pay close attention to variations in style of deformation of sandstone beds as one proceeds north along the exposure. Initial exposures, especially those on the left hand side of the road proceeding north, illustrate a broken formation style of deformation. Individual sandstone beds have been disrupted and are surrounded by scaly cleaved mudstone. Pinch-and-swell and complete boudinage of sandstone beds is common. Boudins generally lack primary structures, range from a few centimeters to at least 5 m in length, and are oriented parallel to a plane defined by scaly cleavage. Zones of deformed sandstone beds and scaly cleavage are adjacent to mesoscopically non-deformed sandstone beds that display numerous types of primary structures such as graded bedding, parts of the Bouma sequence, and groove and load casts on the soles of beds. Zones of mesoscopically deformed sandstone coincide with horizons of relatively low sandstone:shale ratio whereas non-deformed zones are characterized by higher sandstone:shale ratios. Non-deformed zones display a slaty cleavage that dips to the southeast at a lesser angle than that of associated bedding. Slaty cleavage is axial planar to folds in this area that have folded some of the deformation zones, indicating that scaly cleavage occurred well before folding and slaty cleavage formation.

Petrographic analysis of mesoscopically deformed sandstone indicates that stratal disruption was accommodated by particle movement. Analysis of mesoscopically non-deformed sandstones, on the other hand, indicates a higher degree of grain size reduction characterized by grain breakage and pressure solution. This suggests that mesoscopic deformation was accommodated on a microscopic scale by high fluid pressure whereas non-deformed sandstone was deformed by cataclastic grain movement fostered by low pore fluid. The fact that almost all boudins lack internal primary structures and some have irregular bottom and top surfaces similar to load casts suggests that the sand was deformed while water-rich.

Minor folds are rare but can be found within some deformed horizons. Petrographic examination of hinges of these folds indicates that scaly cleavage is axial planar to the folds and may have formed in soft sediment (fig. 83; e.g., Moore and Geigh 1974). Although the vast majority of boudins in this exposure are tapered

Figure 83. Minor folds in scaly cleaved mudstone. Scaly cleavage is axial planar to these folds.

and internally structureless there is one that displays excellent primary structures (Tbce Bouma sequence) and is characterized by square, blunt ends (fig. 84). The morphology of this rare but conspicuous boudin suggests that it was deformed after it was lithified. Unlike most deformed sandstone, the sediment that made up this boudin was lithified enough to retain its internal integrity while undergoing deformation. Disruption of this bed probably occurred during the late stages of deformation as part of a continuum that was initiated in soft sediment and eventually ended with brittle deformation (e.g., Nelson, 1982). Indeed, Carson (1977) pointed out that most deformation of accreted sediment occurs at the time of offscraping and emplacement of a tectonic overburden and decreases with time. In any event, the association of disrupted sandstone beds and horizons of low sandstone:shale ratio and the fact that the mesoscopically deformed sandstone was accommodated by high fluid pressure when deformed is consistent with liquefaction of sand interlayered with water-rich hemipelagic mudstone horizons during the early stages of accretion. Zones of high sand:mud ratio, on the other hand, deformed by cataclastic shear as a result of lower fluid pressure.

The zones of deformation are indirect evidence that sedimentation rates of the sandstone and mudstone were high (approx. 250 mm/TY or higher). Porous hemipelagic mud is generally water-rich immediately after deposition (e.g., Faas, 1982). Rapid sedimentation of sand over a water-rich mud horizon would tend to trap pore fluid in the mud whereas slower accumulation of sediment on the mud would allow for natural dewatering and strengthening of the mud. The deformation zones, then, represent zones in which water was trapped by rapid sedimentation of overlying sand-rich horizons (fig. 85). These water-rich mud horizons were, upon accretion-related deformation, areas of fluidization of interbedded sand intervals, on the other hand, by virtue of their high sand:mud ratio, did not have enough water and deformed in a more brittle manner. A sequence of events that resulted in the structural variations seen at this

Figure 84. Blunt-ended boudin that illustrates well defined internal primary structures.

exposure and others of the Greenwich slice is described below:

1. Accumulation of hemipelagic mud and lesser sand beds in a near-trench or trench axis location (fig. 85A);
2. Rapid accumulation of a thick sequence of sand and minor hemipelagic mud causing an increase in pore fluid pressure in the underlying mud layer (fig. 85B);
3. Initiation of offscraping and accretion-related deformation;
4. Tectonic squeezing caused by the sudden imposition of a tectonic load (structurally overlying accretion complex) resulting in a reduction of pore fluid in the mud and migration of pore fluid into nearby sand beds (fig. 85C);
5. Continued deformation resulted in continued pore reduction and lithification of mud (scaly cleavage formation) and stratal disruption of the water-rich and overpressured sand (fig. 85D);
6. Concomitant deformation of zones of high sand:shale ratio resulted in internal grain size reduction (i.e., pressure solution, microgouge) and reflects significantly lower pore pressure during deformation (fig. 85D).

The variations in style of deformation illustrated at this stop and in other exposures of the Greenwich slice, then, are not controlled by local variations in crustal convergence and strain rate but rather variations in sedimentation rate and lithologic characteristics (refer to the section titled "Accretion-Related Deformation in the Greenwich Slice" for details).

The southern exposure of Stop 4 illustrates a coarsening-upward sequence typical of those described from exposed and active accretionary complexes around the world (fig. 86). Start at the north end on the east side of the exposure and proceed south or up-section. The lower 25 m of the section consists of pelagic sediments that include red mudstone and white dolomitic chert. Note that some of the dolomitic chert beds are graded. Petrographic examination of these rocks indicates the presence of round chlorite bodies that McBride (1962) suggested are psuedomorphs of radiolaria. Some of these beds are characterized by a high degree of folding and boudinage. Although both tectonic and penecontemporaneous folding appear to be present, the majority of deformation is probably sedimentary in origin. This is supported by undeformed beds that drape folds (e.g., Naylor, 1981) and rounded outer limb perimeters and tight interlimb angles. Early Ordovician conodonts have been collected from micritic limestones interbedded with red mudstone that lie stratigraphically beneath the rocks exposed at this part of the section but to the east along strike. Proceeding upsection the red shale

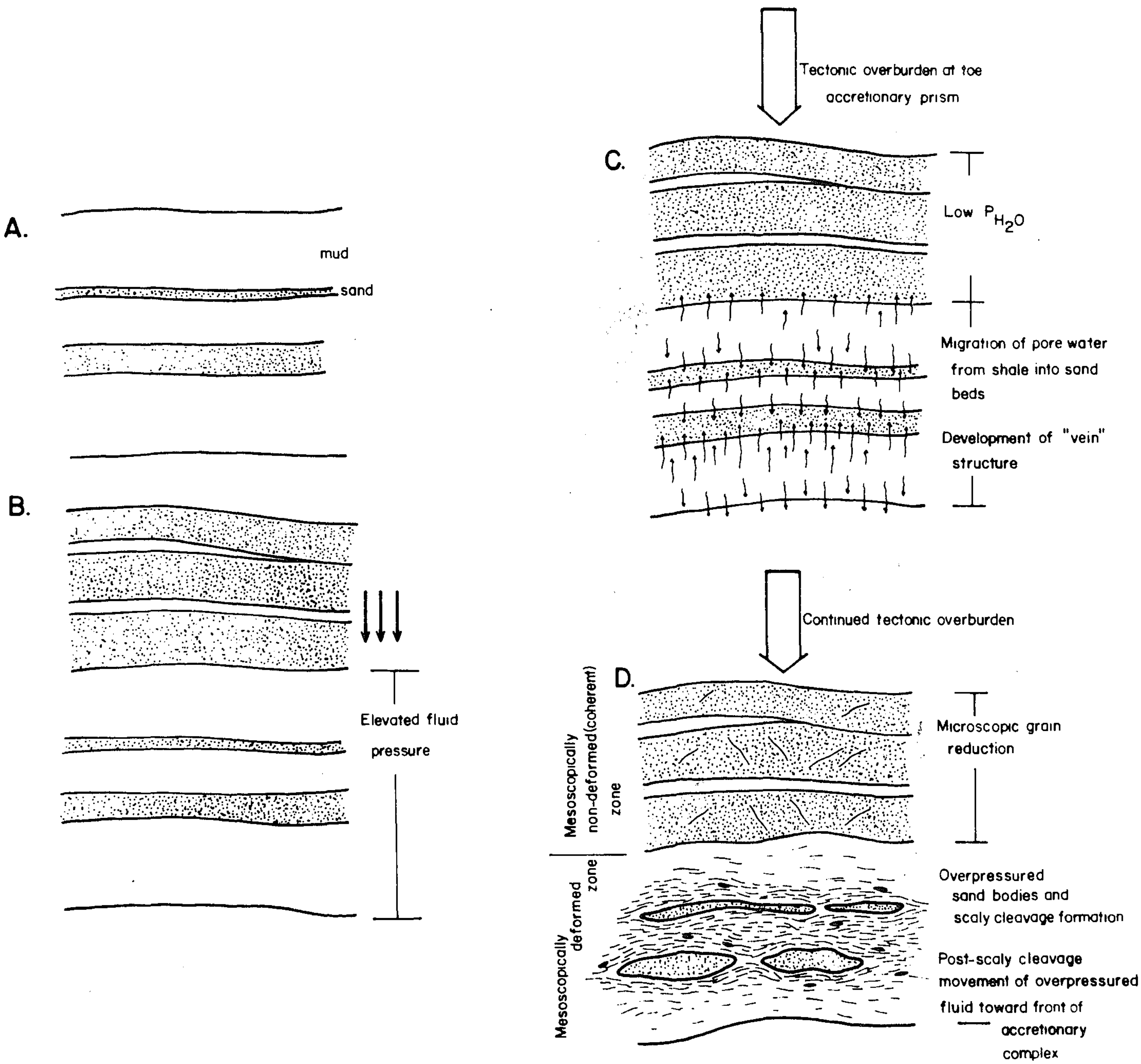


Figure 85. Conceptual model illustrating the important influence of pre-deformation vertical variations in pore fluid content on the manner in which sediment deforms in the early stages of accretion-related deformation. Refer to text for discussion.

passes into approximately 3 m of laminated light-green mudstone and shale. These deposits are similar in composition to the red shale except for lower Fe+3/Fe+2 ratios and lower Mn contents. The geochemical characteristics of these rocks are suggestive of sedimentation under reducing, possibly "near shore", conditions (refer to text section titled "Red Shale Units" for more discussion). The light green shale grades into about 40 m of olive-green mudstone, silty mudstone and thinly-bedded siltstone. These deposits are identical to hemipelagic sediments recovered from continental slope and rise settings at modern active and passive margins. Note that these rocks are cut by a pervasive scaly cleavage. The hemipelagic deposits grade into approximately 25 m of mudstone and sandstone, some of which is

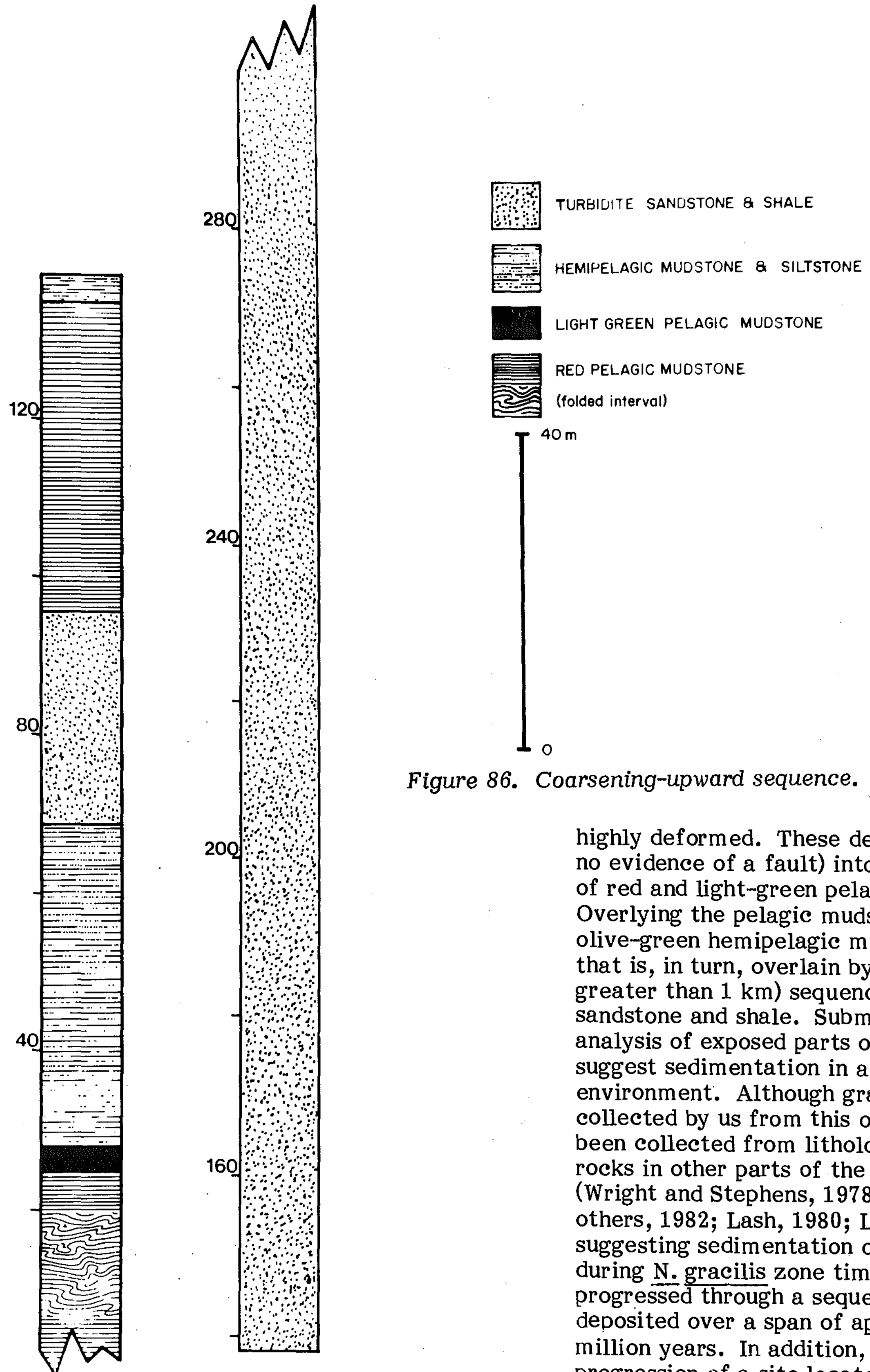


Figure 86. Coarsening-upward sequence.

highly deformed. These deposits "grade" (e.g., no evidence of a fault) into approximately 40 m of red and light-green pelagic mudstone. Overlying the pelagic mudstone is about 3 m of olive-green hemipelagic mudstone and siltstone that is, in turn, overlain by a thick (possibly greater than 1 km) sequence of interbedded sandstone and shale. Submarine fan facies analysis of exposed parts of this sequence suggest sedimentation in a middle fan environment. Although graptolites were not collected by us from this outcrop, they have been collected from lithologically identical rocks in other parts of the Greenwich slice (Wright and Stephens, 1978; Stephens and others, 1982; Lash, 1980; Lash and Drake, 1984) suggesting sedimentation of these deposits during N. gracilis zone time. Thus, we have progressed through a sequence of rocks that was deposited over a span of approximately 25 million years. In addition, this sequence records progression of a site located on oceanic lithosphere from an area of pelagic sedimentation in Early Ordovician time to a near-trench area subject to pelagic-hemipelagic sedimentation probably by early N. gracilis time (Middle Ordovician) followed by movement into the trench axis and the area of submarine fan sedimentation within and possibly near the end of N. gracilis time. This sequence was then accreted to the hanging wall of the opposing southeasterly plate.

The presence of a thick pelagic sequence (between 95 and 135 m) above the first turbidite and hemipelagic mud deposits is somewhat puzzling. Most thick flysch sequences of the Greenwich slice are generally found stratigraphically above pelagic sediment suggesting that they were probably confined to near-trench and trench axis areas by the outer trench slope (e.g., Piper, 1972). If the outer trench slope is breached at some point along its length, or if sediment supply is high enough in a particular area to overflow the outer trench slope a thin blanket of clastic sediment would be deposited on the oceanic floor away from near-trench and trench-axis areas (e.g., Leggett, 1980b). Accurate biostratigraphic data are required to substantiate or modify this scenario.

Sections such as that exposed here, that is, sequences of relatively young hemipelagic and trench clastic sediments overlying older pelagic deposits, can be found in other areas of the Greenwich slice and they are quite similar to cores recovered from eastern Pacific trenches where they can clearly be related to crustal convergence. Refer to text section "Sedimentologic Evolution of the Greenwich Slice" for more details.

Leave parking area and turn left on PA 143 south. Be careful in exiting bus; PA 143 is a very dangerous road.

3.7

46.1 Overpass of US 22 and I-78; continue south on PA 143 into Lenhartsville.

0.2

46.3 Little One Room School House on left.

0.1

46.4 Stop sign, turn left following PA 143 south toward Virginville.

0.1

46.5 Bear right at Y following PA 143 south.

0.2

46.7 Interbedded sandstone and shale of the Windsor Township Formation on right.

0.9

47.6 Long outcrop of interbedded sandstone and shale of the Windsor Township Formation (braided midfan facies). Beds are right side up.

0.4

48.0 Same rocks, but overturned.

0.2

48.2 Dreibelbis Station Covered Bridge on left. Built in 1869.

0.1

48.3 Pelagic green and red mudstone and deep water limestone of early Ordovician age of the Windsor Township Formation on right.

0.2

48.5 "Y" in road, stay left on PA 143 south.

0.9

49.4 Bear left continuing on PA 143. We have just crossed the contact between the Greenwich Slice and overlying Richmond Slice.

0.3

49.7 Quarry in folded shale, siltstone, and ribbon limestone of the Moselem Member of the Virginville Formation.

0.2

49.9 Outcrops of Moselem Member on right.

0.3

50.2 Stop sign. Intersection with Ontelaunee Trail. Continue straight on PA 143 south.

0.3

50.5 Cross Maiden Creek. Enter Virginville.

0.1

50.6 Make sharp left heading towards Kempville.

0.05

50.65 Cross Sacony Creek.

0.05

50.7 Make left turn at "Y" in road. Outcrops of south-dipping Sacony Member of the Virginville Formation straight ahead. Old Reading Railroad tunnel to right.

0.1

50.8 Passing through the better parts of Virginville.

0.1

50.9 Stop 9. Pull off road to right.

Stop 5. Exposure of Fault Separating the Sacony-Onyx Cave Sequence from the Moselem Member

Stop 5 illustrates the fault that separates the Sacony and Onyx Cave Members from the younger Moselem Member. This fault is visible at a number of other areas including exposures to the immediate south of Stop 9 of the 1982 Field Conference of Pennsylvania Field Geologists but this exposure is the most accessible and best exposed. In addition, we will get a glimpse of the Sacony Member of the Virginville Formation, a unit we have not seen yet. As noted at Stop 1, conodont collections from the Onyx Cave Member yield Late Cambrian ages thereby suggesting a to Late Cambrian age of the depositionally underlying Sacony Member. This exposure and Stops 1 and 2 illustrate the stratigraphic and structural relations of the three members of the Virginville Formation/Richmond slice (fig. 87). Start at the south end of the exposure and move north. Black shale of the Moselem Member dips gently to the southeast at the southern part of the exposure. Excellent exposures of south-dipping Sacony Member clastic rocks that apparently overlie the younger Moselem Member can be found just to the south of this stop. At the north end of the exposure, Moselem beds can be seen to change attitude abruptly and are inclined to the northwest. The fault, too, is inclined to the northwest and is visible at the north end of the exposure in a small borrow pit (fig. 88). The fault is marked by a highly deformed chaotic zone of brecciated limestone and cleaved pelitic rocks of the

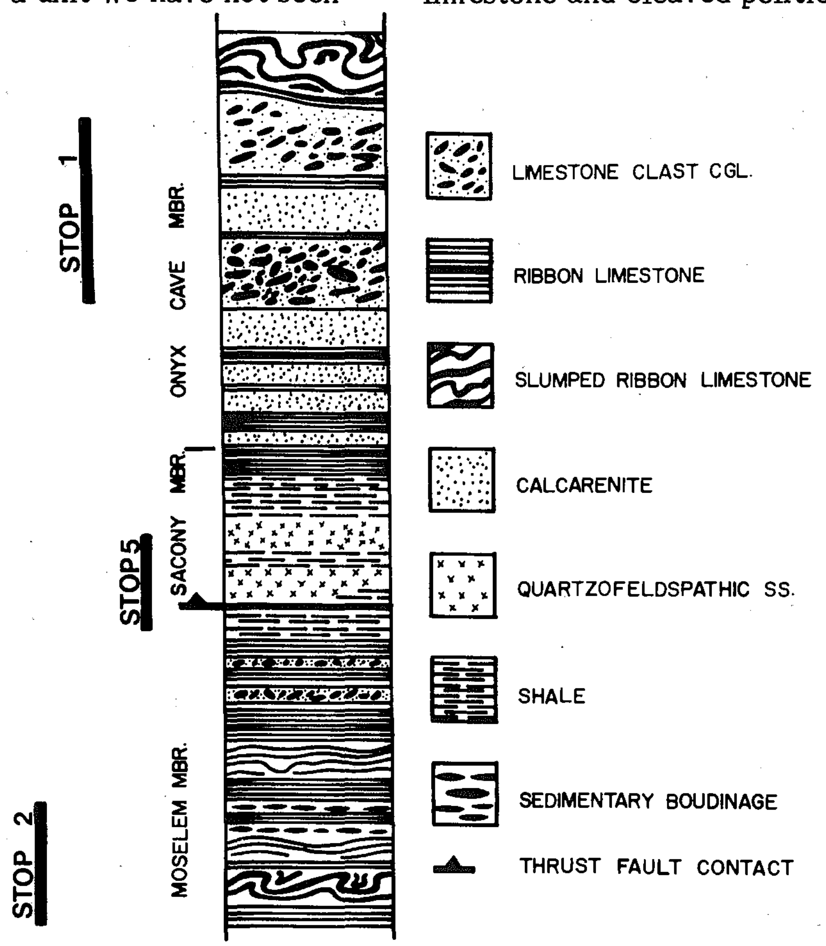


Figure 87. Generalized stratigraphic column of the Richmond slice showing stratigraphic section seen at Stops 1, 2, and 5 within the framework of the slice.

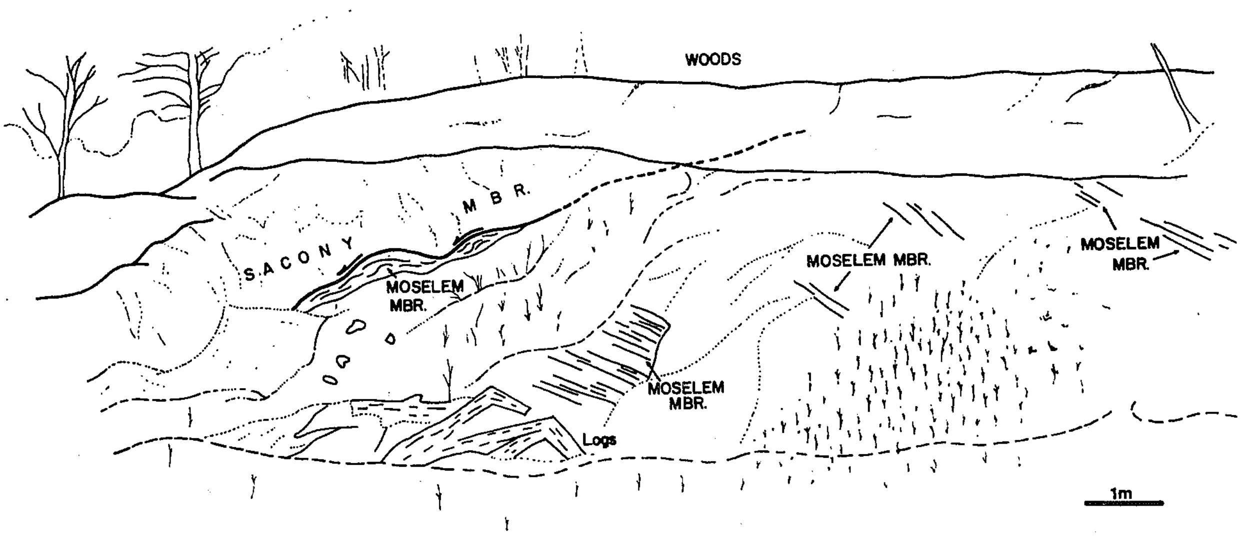


Figure 88. Sketch of the north end of Stop 5 showing the fault separating the Sacony and Moselem Members and the attitude of Moselem beds beneath the fault.

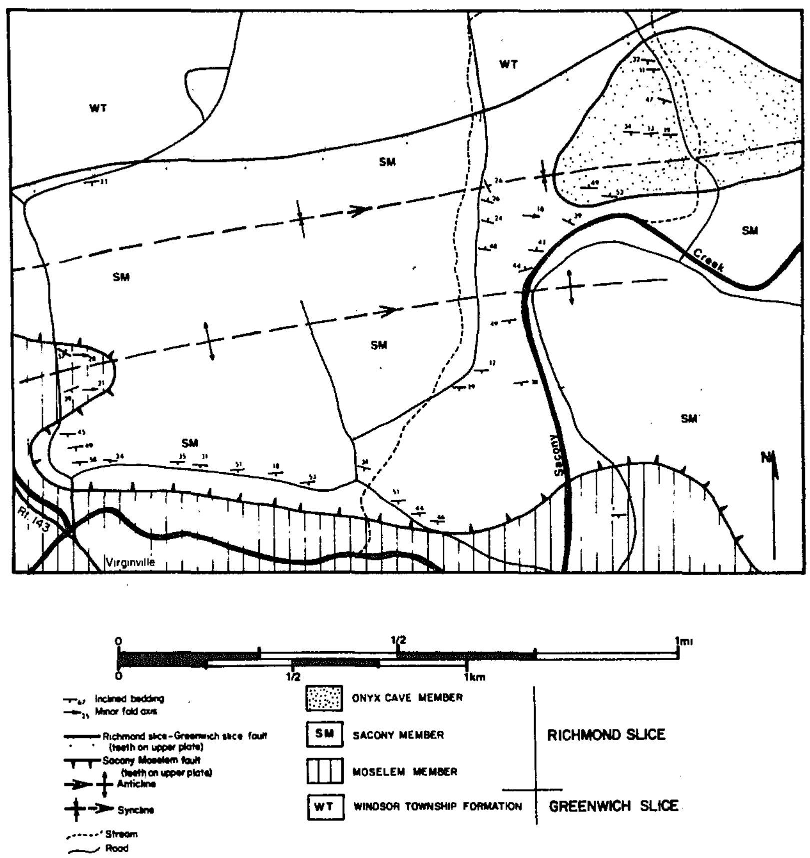


Figure 89. Geologic map of area surrounding Stop 5 (from Lash, 1980).

Moselem Member. The overlying Sacony Member is also deformed and contains clasts of itself (autoclastic melange). It was apparently deformed as a result of movement on the fault plane. Although beds of the Moselem Member are parallel to the fault, close to the fault this parallelism may merely be the result of drag along the fault plane. Evidence for this includes the presence of south-dipping Moselem beds beneath the north-dipping fault plane near the north part of the exposure (fig. 88).

Minor folds related to movement on the fault plane are present at this exposure as well as others. Strain-slip analysis of these folds (e.g., Hansen, 1971) suggests that the upper plate composed of the Sacony and Onyx Cave Members was thrust from the southeast over the Moselem Member (Lash, 1980). The fault plane was subsequently folded as illustrated at this exposure as well as by mapping in the area (fig. 89). Minor folds related to movement on the fault plane verge to the northwest regardless of what fold limb they are found on indicating that the minor folds formed prior to the major fold phase. The folded thrust fault plunges to the east (fig. 89). The Onyx Cave Member which overlies the Sacony crops out in a rather large quarry within the keel of a syncline to the north of the anticline exposed at this stop (fig. 89).

Retrace route back to Virginville.

0.2

51.1 Bear right at "Y" in road.

0.1

51.2 Cross Sacony Creek. At stop sign turn left on PA 143 south heading toward US 222.

2.2

52.4 Siltstone and shale of the Sacony Member on left.

0.7

53.1 View of the Reading Prong, underlain by Precambrian rocks, straight ahead.

0.3

53.4 Cross Moselem Creek.

0.1

53.5 Stop sign. PA 143 ends. Continue south on PA 662 toward Moselem Springs.

0.1

53.6 Abandoned quarry within Ontelaunee Formation, the uppermost part of the Beekmantown Group, to right. These rocks have been thrust northward as the Irish Mountain nappe as mapped by MacLachlan (1979) onto rocks of the Hamburg klippe which we have just left.

Traveling South on PA 662 and southwest on US 222 we will be descending stratigraphically within the Beekmantown Group, eventually winding up within the Allentown Dolomite. Note the typical rolling topography on these Cambrian and Ordovician carbonate rocks.

1.2

54.8 Village of Moselem Springs. Make right turn on US 222. Note historic Moselem Springs Inn, established in 1852, on right. They have very good duck in here. Perhaps some drivers did not heed the "Duck Crossing" sign at Spitzenberg.

2.3

57.1 Irish Mountain at 11 o'clock, underlain by Precambrian gneisses which form the core of the Irish Mountain Nappe.

1.2

58.3 Entering village of Maiden Creek.

0.6

58.9 Junction with PA 73. Continue on US 222 south.

1.4

60.3 Bear right onto US 222 South heading towards Sinking Spring. We are on the Allentown Dolomite.

1.4

61.7 Scenic junkyard on right.

0.

62.3 Turn right on PA 61 south. We are still on the Allentown Dolomite.

0.4

62.7 Turn right onto PA 61 south.

1.0

63.7 Exxon Tuckerton Gas Terminal on right.

0.5

64.2 Reading Railroad (now Conrail) overpass.

0.5

64.7 Stromatolitic Allentown Dolomite outcrops on right.

0.9

65.6 Turn right onto US 222 South heading to I-176 and US 422.

0.6

66.2 Cross Schuylkill River on US 222.

0.9

67.1 Intersection with PA 183 (Schuylkill Avenue). Continue straight on US 222. Stay in left lane.

0.6

67.7 Continue straight on US 422 West, do not turn off onto US 222. Bear left at traffic lights ahead.

0.1

67.8 Turn left at traffic left onto Park Road. Outcrops of carbonate are in the Allentown Dolomite.

0.1

67.9 Turn right into parking lot of Reading Motor Inn.

END OF FIRST DAY.

DAY 2

Leave parking lot of Reading Motor Inn turning left onto Park Road.

0.1 Turn right at traffic light onto US 222.

0.1

0.2 Continue straight on US 222 north. Allentown Dolomite on right.

0.2

0.4 Cross Tulpehocken Creek.

0.4

0.8 Intersection of US 222 with PA 183. Continue straight on US 222 south.

0.9

1. Cross Schuylkill River.

0.4

2.1 Junction with PA 61 north. Do not turn onto PA 61 south. Follow PA 61 north for 15 miles until Hamburg, PA.

1.2

3.3 Stromatolitic Allentown Dolomite on left.

2.3

5.6 Village of Cross Keys.

0.3

5.9 Outcrops of Allentown Dolomite.

1.0

6.9 Cross Maiden Creek.

0.6

7.5 Junction with PA 73 east. Continue north on PA 61. Cross contact between Allentown Dolomite and Beekmantown Group.

1.2

8.7 Town of Leesport. Contact of Beekmantown Group carbonates and structurally overlying rocks of the Hamburg klippe.

0.6

9.3 Entering Ontelaunee Township. Ontelaunee is the Delaware Indian word for maiden or virgin.

0.5

9.8 Siltstones of the Sacony Member of the Virginville Formation on right.

0.5

10.3 More of same on left.

0.3

10.6 Moselem Member of the Virginville Formation in pit on left. This was described as the Leesport Limestone by Stose and Jonas (1927).

0.5

11.1 Siltstones in the Sacony Member of the Virginville Formation on right.

0.7

11.8 Glen-Gary brick plant on right.

0.4

12.2 View of Blue Mountain, underlain by quartzites of the Tuscarora Sandstone of Silurian age, to north.

0.6

12.8 Shoemakersville.

1.9

14.7 Overturned sandstone and shale of the Windsor Township Formation.

0.2

14.9 Right side up sandstone and shale of the Windsor Township Formation.

0.3

15.2 Schmeck's Family Restaurant on left.

0.6

15.8 Town of Hamburg. Continue north on PA 61.

0.9

16.7 Cross Schuylkill River.

0.3

17.0 Turn east onto US 22 and I 78 heading towards Allentown.

0.6

17.6 Cross Schuylkill River.
0.4
18.0 Turn right at Hamburg exit. Shale exposed in clover leaf is in the Windsor Township Formation.
0.1
18.1 Stop sign. Turn left onto North Fourth Street.
0.3
18.4 Sharp right turn onto Port Clinton Avenue.
0.2
18.6 Pass under I 78.
0.2
18.8 Braided mid-fan sandstone and shale of the Windsor Township Formation.
0.5
19.3 More of same.
0.6
19.9 Beginning of outcrop for Stop 6.
0.2
20.1 Stop 6. Pull off on the right shoulder.

Stop 6. Submarine Fan Deposits of the Windsor Township Formation, Greenwich Slice

Stop 6 illustrates some of the sedimentologic characteristics of the Windsor Township Formation. These rocks are gently to moderately inclined to the south at this exposure. Approximately 500 m north of this stop are exposures of the Tuscarora Sandstone (Late Ordovician-Early Silurian; Epstein and Epstein, 1972) inclined steeply to the north in apparent angular unconformity with the Windsor Township Formation. Fossils have not been collected from rocks of this exposure by any of the authors. However, lithologically and compositionally similar rocks to the south yield graptolites of the N. gracilis zone. Figure 90 illustrates a measured section of part of the total section exposed here. Start at the north end of the exposure and move south or upsection. We will begin by progressing through a thinning-upward megasequence that is characterized in its lower part by thick to very-thick sandstone beds that can be classified as Facies B beds of the Mutti and Ricci-Lucchi (1975) classification. Some of these beds are amalgamated (fig. 90) and display erosive features such as rip-up clasts and minor channeling. Although a number of these beds display coarse-tail or content grading most are not graded or poorly graded at best. Beds in this part of the section attain a maximum thickness of approximately 3 m. The thick-bedded sandstone is overlain by about 11 m of turbidite-shale facies beds that become noticeably thinner up-section (fig. 90). Sandstone beds in this part of the section are generally less than 1 m thick and are graded (Facies C of Mutti and Ricci-Lucchi, 1975). Note the associated decrease in sand:shale ratio and the lack of amalgamation in this part of the section. Thus from 6 m above the bottom of the section to 26 m above the base, a vertical distance of about 20 m, we see a gradual but conspicuous decrease in bed thickness and sandstone:shale ratio. This sequence probably records abandonment of a submarine channel (refer to section titled "Submarine Fan Facies Associations of the Greenwich Slice" for detailed discussion). In particular, beds of the thick-sandstone facies tell of plugging of laterally migrating channel thalwegs whereas overlying turbidite-shale facies beds are probably interchannel deposits derived from an adjacent "new" channel. This sequence is overlain by about 7 m of predominantly turbidite-shale facies beds that form a thickening-upward cycle (fig. 90). Sequences such as these are characteristic of prograding depositional lobe deposits. The close association of channel and depositional lobe deposits of the Windsor Township Formation is characteristic of sedimentation on a suprafan (Walker, 1978; Ricci-Lucchi, 1981). A suprafan setting is also supported by the coarse-grain size of the sediment which would tend to promote a more "inefficient" sand transport system resulting in the close association depositional lobe and channel deposits. Thick-shale facies deposits (about 9 m) overlie the lobe sediments (fig. 90) indicating sudden abandonment of the channel that was feeding the lobe. Lobe abandonment may have occurred as a result of channel avulsion farther upslope. The only turbidite deposits in this sequence are fine sand and silt beds that display Tce partial Bouma sequences. The shale facies beds are overlain by turbidite-shale facies beds (41 m) that mark a return to active sand sedimentation (fig. 90).

This exposure illustrates the variability of sedimentation on the Windsor Township submarine fan(s). As noted above, the vertical

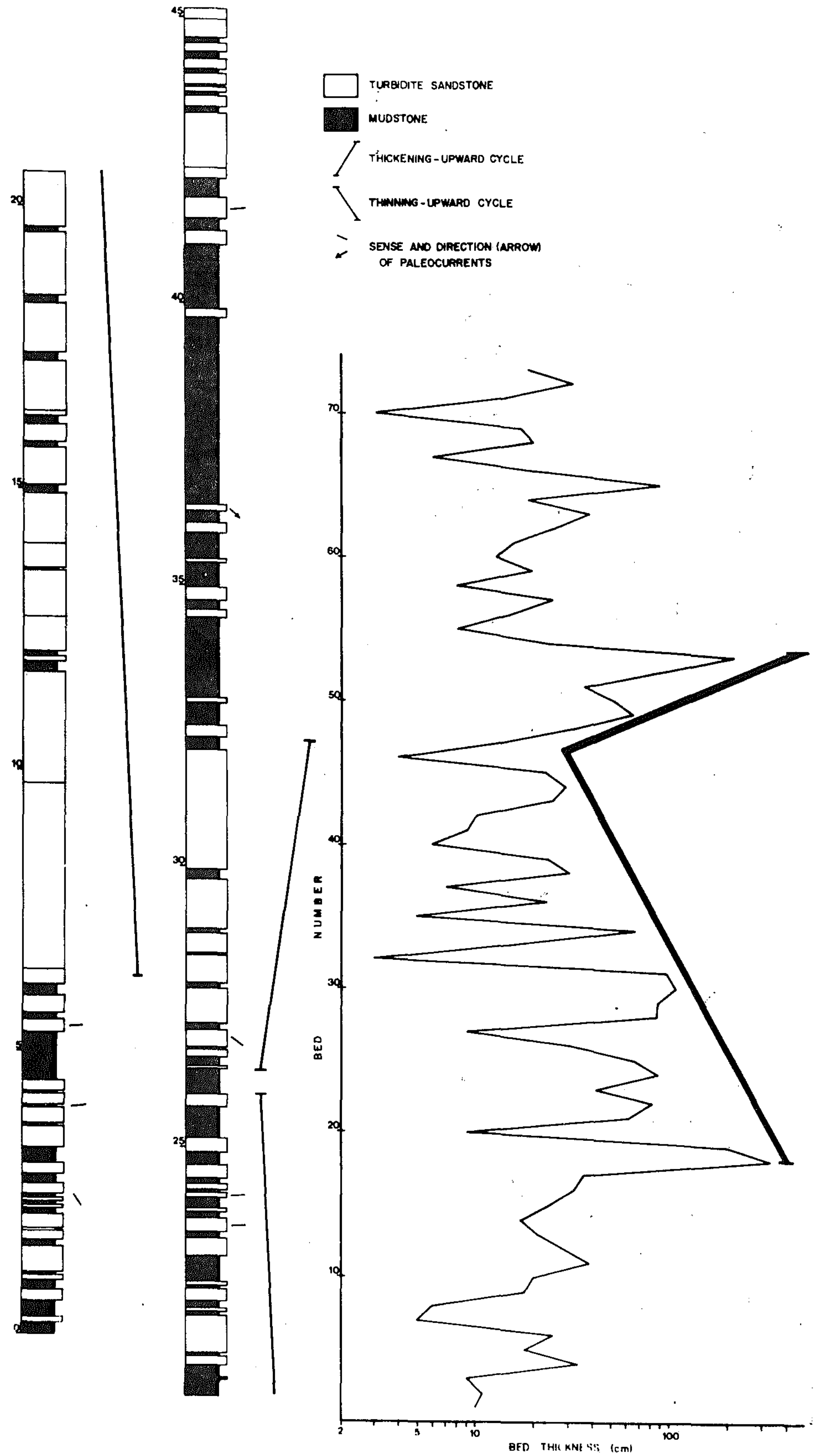


Figure 90. Lithologic log showing alternations of sandstone and shale and thickness log defining vertical variations in sandstone layer thickness. Thicknesses in meters.

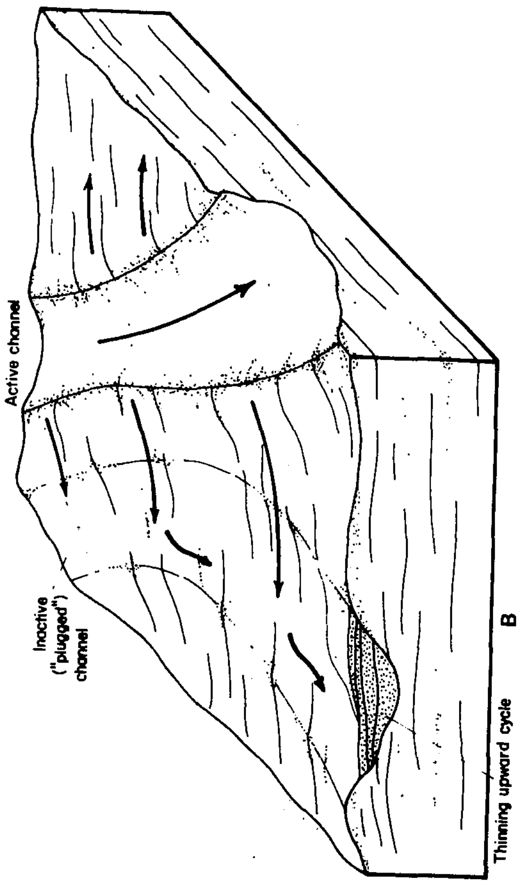
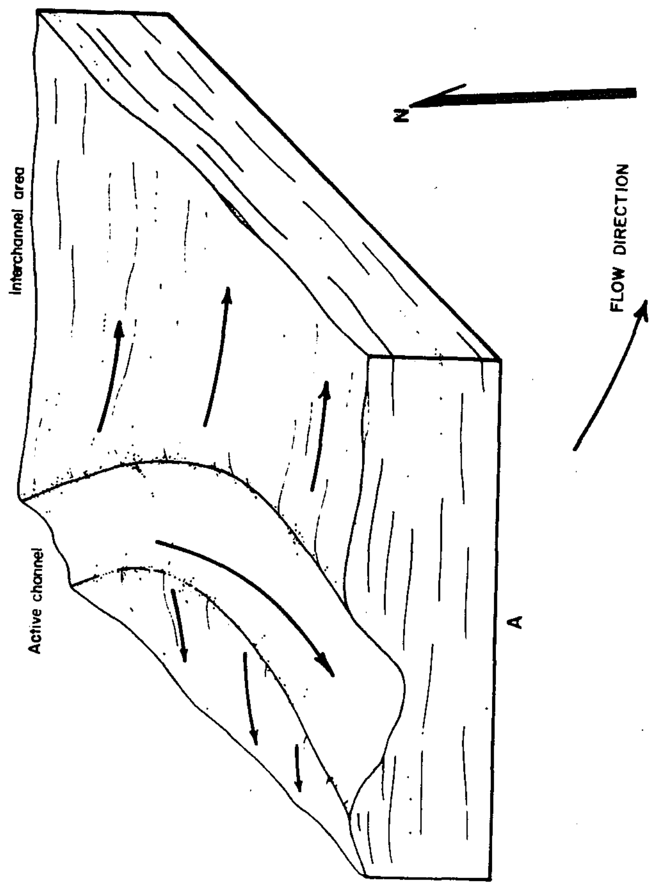
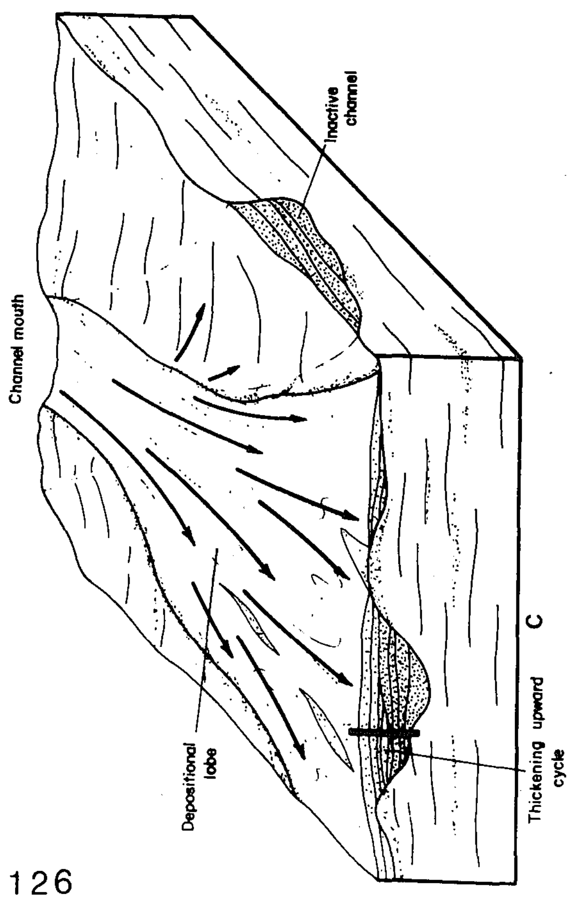


Figure 91. Conceptual model illustrating vertical facies illustrated at *STOP 6.* *Refer to text for discussion.*

juxtaposition of channel and deposition lobe deposits and the coarse-grain size of these deposits is suggestive of sedimentation on a suprafan rather than the more stable type of fan described by Mutti and Ricci-Lucchi (1972, 1974). In addition, the conspicuous absence of levee deposits (turbidite-shale facies) in this sequence suggests that the sediment required for levee building was not present in sufficient quantities, a characteristic typical of "inefficient" suprafan systems (e.g., Normark, 1978).

Figure 91 illustrates a possible scenario for the sedimentological evolution of the section exposed at Stop 6. Deposition of the thick sandstone facies beds in the lower part of the section plugged and filled a channel that had been cut into interchannel sediments (fig. 91A,B). The overlying turbidite-shale facies were probably deposited from overspill of an adjacent channel (fig. 91B). Many of these deposits may have been funnelled along the "old" channel giving rise to E-W paleocurrent trends (fig. 90). The "new" channel then shifted position and the area became the site of accumulation of lobe sediments that were





probably deposited at the mouth of a nearby channel and funnelled along the old channel. The "old" channel morphology may have, to a certain degree, constrained the direction of lobe building (fig. 91C). The sudden cessation of lobe sedimentation in this part of the fan probably resulted from a major channel avulsion upslope that isolated this part of the fan from sand sedimentation allowing it to be covered by a blanket of hemipelagic mud. Subsequent

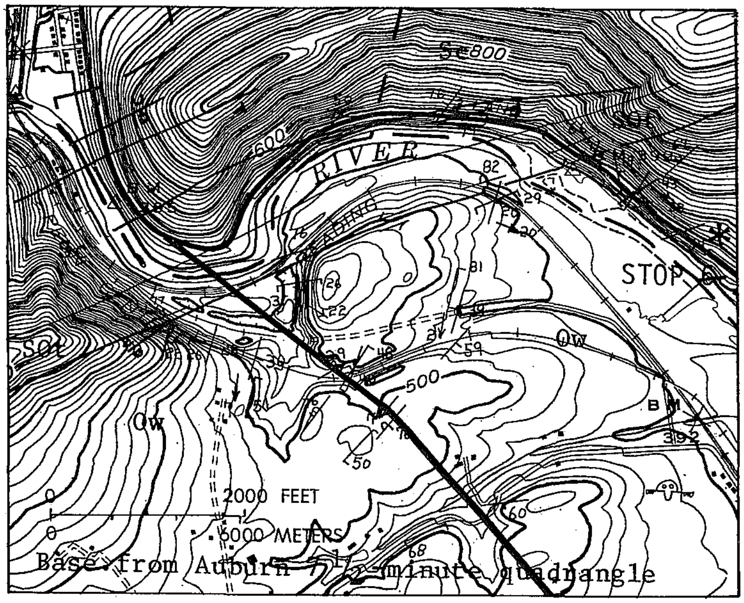


Figure 92. Geologic map of the southern part of Schuylkill Gap including Stop 6.

Explanation

Ow, Windsor Township Formation
SOt, Tuscarora Sandstone
Sc, Clinton Formation.

Contact

Thrust fault. T on upthrown plate.

Fault: u, upthrown; d, downthrown.

Strike-slip fault.

Upright Overturned
Strike and dip of bedding.

Strike and dip of cleavage.

Bearing and plunge of intersection of bedding and cleavage.

migration of a channel led to renewed sedimentation of sand in this area. Refer to the sections entitled "Submarine Fan Models" and "Submarine Fan Facies" for a more detailed discussion of submarine fans and sedimentation of fan sediments of the Windsor Township Formation.

Bedding in the Windsor Township Formation about one-half mile (0.8 km) south of the Tuscarora contact parallels the northeast structural grain (fig. 92). As the contact is approached, bedding in the Windsor Township strikes more northerly until finally it is nearly normal to the contact. The intersection of bedding and cleavage, approximately parallel to the trend of the fold axes, likewise is rotated counterclockwise as the contact is approached. Two interpretations of these relationships (or combinations, thereof) are possible. One is that the Windsor Township was folded during Taconic deformation prior to deposition of the Tuscarora. The other is that the beds are dragged into a fault along the contact. Hoskins (in Wood and others, 1963) discussed some of the aspects of faulting near the contact. Additionally, there is gouge and sheared shale 5 to 30 cm (2 to 12 in) thick at the contact with slickensides indicating an earlier northwest translation of the overlying Tuscarora, and a later set indicating left-lateral strike-slip movement. For this reason, a strike-slip component is shown on Figure 92. It is puzzling, however, that the cleavage in the Windsor Township reverts to a northeast strike at the contact and is not rotated as is the bedding. Locally, a northwest-dipping crenulation cleavage is developed at the contact. Another puzzling consideration is that if the contact is only an angular unconformity, and if we rotate the Tuscarora back to the horizontal, the the underlying Ordovician rocks would contain cleavage that dips very steeply to the northwest, a strange pre-Silurian orientation indeed. This is a common problem all along the sub-Tuscarora/Shawangunk contact in eastern Pennsylvania. The complexly faulted Clinton Formation north of the Tuscarora contact along Pa. 61 has been described by Burtner and others, 1958.

Continue north on Port Clinton Avenue.

0.1

20.2 Steeply dipping conglomerates and quartzites of the Tuscarora Sandstone of Silurian age.
0.4
20.6 Red sandstone and siltstone of the Clinton Formation which overlie the Tuscarora Sandstone.
0.5
21.1 Stop sign, intersection with PA 61. Turn left heading south.
0.1
21.2 Cross Schuylkill River. There are excellent exposures of the Tuscarora Sandstone and sandstone of the Windsor Township Formation on both sides of the road along the railroad cut.
1.6
22.8 Pass over I-78, bear right onto I-78 east on US 222 east heading towards Allentown. Continue east on I-78 and US 222 for 21 miles until you reach PA 100 at Fogelsville.
1.8
24.6 Smokestack to right is at the Hamburg Sanitarium.
2.4
27.0 Sharpes Mountain at 10 o'clock and Spitzenberg at 11 o'clock. The tops of both hills are capped by coarse sandstones and conglomerates primarily derived from the immediately underlying Hamburg klippe.
1.0
28.0 Overturned sandstone and shale of the Windsor Township Formation.
0.4
28.4 Upright beds of the same.
0.9
29.3 Pass over Maiden Creek.
0.9
30.2 Limestone-clast conglomerate and green shale on right descibed by Epstein and others (1972).
1.8
32.0 Shale and sandstone of the Windsor Township Formation on left.
1.3
33.3 Same stuff.
2.4
35.7 Interbedded sandstone and shale of the Windsor Township Formation.
0.9
36.6 Same rocks.
0.6
37.2 Same stones.
1.1
38.3 Sheared shale of the Windsor Township Formation marking the eastern boundary of the Hamburg klippe. To the east lie rocks of the Martinsburg Formation.
0.9
39.2 Overpass of PA 863.
0.8
40.0 Exposures in roadcut on both sides of road are of the Bushkill Member of the Martinsburg Formation.
0.9
40.9 Exposures of the Bushkill Member of the Martinsburg Formation on left. Straight ahead are rocks of the Martinsburg Formation and older rocks in the overturned limb of the Northampton nappe of Sherwood (1964).
1.6
42.5 Flat-floored valley underlain by carbonates of the Lehigh Valley sequence. Precambrian rocks of the Reading Prong in the distance to the right.
0.7
43.2 Strohs Brewery on right.
0.5
43.7 Turn right onto clover leaf, then turn right onto PA 100 toward Fogelsville.
0.6
44.3 Traffic light. Continue north on PA 100. The break in slope about 500 ft north marks the boundary between the carbonate rocks of the Beekmantown Group and the Jacksonburg Limestone.
0.4
44.7 Climb onto the Jacksonburg Limestone. Quarries to the immediate right are in the Epler Formation of the Beekmantown Group.
0.2
44.9 Large quarry to right in Jacksonburg Limestone.
0.1
45.0 Turn right on Haasadahl Road.
0.3

45.3 Overturned and faulted contact between the Jacksonburg Limestone and Martinsburg Formation.

0.2

45.5 Intersection of Haasadahl Road and Hilltop Road. Turn left on Hilltop Road.

0.1

45.6 Stop 7. Pull buses off on the shoulder of Hilltop Road.

Stop 7. Bushkill Member of the Martinsburg Formation: Fault-related structures and sedimentology

This stop allows us to look at the basal part of the Bushkill Member of the Martinsburg Formation several hundred feet north of the steep to overturned contact with the underying Jacksonburg Limestone. It is important to compare the Bushkill at this stop with the rocks at the next two stops. Numerous workers have stated that the rocks at all three localities are the same unit, and we hope to demonstrate the many and varied differences. The Bushkill is a laminated to thin-bedded ribbon slate containing thin beds of quartzose slate, some carbonaceous slate, and siltstone and sandstone turbidites. The sandstone beds rarely exceed 12 inches in thickness anywhere and here rarely exceed 6 inches. The slate beds usually are less than 2 inches thick but range up to 6 inches. On slaty cleavage surfaces, the sand and silty beds in the Bushkill characteristically weather to a moderate yellow brown or an orange brown color giving the unit a diagnostic striped appearance.

There is considerable shearing and thrust faulting at and near the contact of the Bushkill and Jacksonburg Limestone which clearly affects the rocks exposed here (A. A. Drake, Jr., unpublished maps and personal communication, 1984). As shown on the accompanying geologic map, the Jacksonburg appears to be carried as discontinuous slivers along the leading edge of a thrust that can be mapped almost continuously across the entire map. To the northeast along strike in the Nazareth quadrangle, this fault appears to die out, and the contact between the Bushkill and Jacksonburg appears to be a normal conformable sedimentary contact. Not far from this stop, there are areas where the Jacksonburg is entirely missing and the rocks of the Beekmantown Group are in direct contact with the Bushkill. Although much farther to the northeast in New Jersey and New York, this relation can be explained by the deposition of the Jacksonburg on a very irregular Beekmantown erosional surface, here the explanation is most likely structural. Although there may be a considerable amount of movement parallel to bedding along this thrust, there is very little stratigraphic offset where it is exposed.

Structures visible at this stop that result from movement along the fault at the Bushkill/Jacksonburg contact, are both open and tight folds in slaty cleavage and sets of strain-slip crenulation cleavage that are axial planar to these folds. Along the entire length of the outcrop there are large open warps in slaty cleavage. The axes of these late very open folds are oriented roughly east-west. Slip along slaty cleavage planes produced by this late warping of cleavage is accompanied by the infilling of vein quartz.

Approaching the bend in the road about 200 feet south of the road intersection, a large recumbent fold in bedding is visible. At this point slaty cleavage dips to the north. Regionally, slaty cleavage dips consistently south or southeast. It is only when there is late folding of slaty cleavage generally associated with thrust faulting that the cleavage is rotated to a northward dip.

Right at the bend in the road, there are very tight north-verging, overturned folds in slaty cleavage. Axial planar to these folds is an excellent strain-slip cleavage oriented N. 40° E., 57° SE. This cleavage appears to cut another strain-slip cleavage oriented N. 53° W.,65° NE., but these relations are ambiguous and it is left up to the field tripper to decide. On the whole, bedding dips north here and cleavage dips south. From this point to the contact with the limestone several hundred feet to the south, bedding remains steeply dipping to the north, or slightly overturned dipping steeply to the south. Slaty cleavage continues to be complexly folded between here and the contact. Near the contact, in a very shaly limestone outcrop that we will not be visiting, there is an excellent lineation seen on slaty cleavage surfaces suggesting that transport along the thrust fault is N35°W. The amount of vein quartz and the degree of contorted bedding and crenulation cleavage increases as the fault is approached.

Both Plate 1 and Figure 70 show the unnamed fault immediately south of this stop. It is immediately apparent that the thrust must be gently to moderately dipping and that it is folded into rather large scale roughly east-west trending folds. Figure 70 also shows that this fault is offset by the Kistler Valley fault that we will be visiting at Stop 8. The Kistler Valley thrust fault dips moderately to the south for its entire length and remains quite straight. Aaron (1969) has noted large scale (1 mi wavelength) east-west trending open folds in the Nazareth quadrangle not far to the east. Lyttle and Drake (1979) and Lyttle, Lash, and Epstein (1985) note that similar open folds are present throughout the Slatedale quadrangle. It seems very likely that these folds are directly related to blind thrusts, which like the Eckville and Kistler Valley faults are fairly straight and east-west trending (see cross section on Plate 1 for examples).

It is difficult to prove that the Kistler Valley fault cuts the unnamed fault exposed near this stop. However, the sheared contact between the Jacksonburg and the Bushkill takes an abrupt east-west bend in the southwest Cementon quadrangle. In addition, the fact that the unnamed thrust is much more folded than the Kistler Valley fault suggests that the latter is younger. This is in keeping with the popular younging-toward-the-foreland model for thrust faults. If the Kistler Valley fault truly offsets Silurian age rocks at Hawk Mountain, we can perhaps be comfortable assigning that fault an Alleghenian age. The age of the fault immediately south of this stop is more of a problem. Based on similarities between these faults (extensive tight folding of slaty cleavage and the presence of vein quartz in the hinges of these folds) we feel that this fault may also be Alleghenian in age. However, it clearly could be older.

Return to buses, turn right on PA 100 heading north.
0.8

46.4 Village of Claussville. We are still on the Bushkill Member of the Martinburg Formation.
0.7

47.1 Turn right on Kernsville Road heading toward Orefield.
0.7

47.8 Entering village of Leather Corner Post.
0.3

48.1 Bear left at "Y" in road at grain elevators.
0.1

48.2 Stop sign at Leather Corner Post Hotel (home of the world champion Oom-pah band), make left turn on Highland Road.
0.2

48.4 On a clear day you can see both Lehigh Gap and Delaware Water Gap cut through Blue and Kittitinny Mountains to the north.
0.2

48.6 Intersection of Highland Road and Church Road. Bear right on Highland Road.
0.8

49.4 To the left in the distance you can see Shochary Ridge trending westward where it eventually plunges beneath rocks of the Tuscarora Sandstone in Hawk Mountain.
0.1

49.5 Field to left is full of vein quartz marking the trace of the Kistler Valley fault that we will see at the next stop.
0.1

49.6 Stop sign, turn left on Rhueton Hill Road.
0.1

49.7 Valley to left follows the the Kistler Valley fault.
0.1

49.8 Pavement outcrops of the Bushkill Member of Martinsburg Formation exhibit E-W trending late folds of the slaty cleavage.
0.3

50.1 Road intersection. Turn left on Weidasville Road.
0.2

50.3 Beginning of Stop 8. Jordan Creek on right.

0.1

50.4 Stop 8. Park along right shoulder of road and walk back to roadcut.

Stop 8. Kistler Valley fault zone

BEWARE OF POISON IVY—PARTICULARLY AT SOUTH END OF THE OUTCROP

The main purpose of this stop is to show the deformation in a fault zone at least 600 feet wide and to illustrate differences in rock type and bedding character between the New Tripoli Formation exposed here and both the Bushkill and Pen Argyl Members of the Martinsburg (Stops 7 and 9 respectively). Both of these goals can be achieved at a number of spots in these outcrops, so it is not important to see every inch of the exposure. This stop consists of two long outcrops that are separated by a tiny stream entering Jordan Creek. We will begin at the larger road cut at the south end and work our way northward along the road (fig. 93).

The Kistler Valley fault can be traced about 2 miles east from here into the Cementon quadrangle and about 18 miles west through the Kistler Valley of the New Tripoli and New Ringgold quadrangles. There are few places along its length where all of the structures produced by this fault are so nicely exposed. However, even when the fault crosses cultivated farm lands and there are no exposures, it is almost always possible to find abundant float of vein quartz, which is so commonly produced by late Alleghenian thrust faults, and which is abundantly present at the north end of this stop. The Kistler Valley fault zone of probable late Alleghenian age marks the northern edge of the Hamburg klippe for much of its length, but at this stop thrusts the New Tripoli Formation over itself greatly increasing the thickness of that unit locally. Regionally the fault trends east-west, but locally trends east-northeast and east-southeast.

This stop is located very near the eastern end of the synclinally folded Shochary Ridge thrust sheet and is on the more steeply dipping to overturned southern limb of the Shochary

Figure 93. Kistler Valley thrust fault zone exposed at southern part of STOP 8. At this locality the New Tripoli Formation shows large number of fault splays. Bedding is severely broken up, but the remnants of north-verging slightly overturned folds show clockwise sense of rotation suggesting this locality is located on the south limb of the Shochary syncline. Counterclockwise sense of rotation of folds in cleavage, as well as slickenline orientations, suggest thrusting to the northwest (left).

synform. At the east end this limb has been faulted and repeated by the Kistler Valley fault. Just east of this locality the fault cuts the Game Preserve fault of probable Taconic age. There is an apparent right lateral offset of approximately one mile on the Game Preserve fault. However, the dominant component of movement on the moderately south-dipping Kistler Valley fault is vertical and may amount to something on the order of 2000 feet. It is possible that several splays of the Kistler Valley fault, at its western termination in the New Ringgold quadrangle, may cut and offset the Silurian Tuscarora Formation. This is the way relations are shown on the cross section on Plate 1. However, it is also possible that much of the movement of the Kistler Valley fault is taken up on the Alleghenian Blue Mountain decollement. The talus along the slopes of Blue Mountain prevent the actual tracing of the Kistler Valley fault westward into the thrusts that cut the Silurian rocks.

It is impossible to estimate the amount of movement on the many splay faults visible at this locality due to lack of marker beds. The aggregate amount of shortening within the New Tripoli Formation near the fault may be considerable, but movement on each splay may be very small. And indeed, even though this stop is definitely within the Kistler Valley fault zone, the main fault is impossible to define. It may well be located along the tiny stream valley dividing this stop in two. When bedding can be recognized, sedimentary facing directions suggest that the remnants of northward-verging overturned folds are still present. Shearing disrupts bedding severely at several places in this stop, producing an incipient broken formation and disrupting the folds, but not entirely destroying them (see Figure 71 which shows similar structures 18 miles to the west along the same fault). Whether these folds were produced during the earliest stages of development of the Kistler Valley fault or during a previous structural event is unknown. The clockwise sense of rotation of bedding seen at several spots at this stop also suggests that we are indeed on the southern limb of the regional Shochary synform. These folds are not to be confused with the tight kink folds in slaty cleavage that are clearly associated with many of the imbricate thrust splays. These kink folds show a counterclockwise sense of rotation that is appropriate for south-dipping thrust faults whose transport direction is to the north or northwest.

Many, but not all, of the thrust splays found at this outcrop are oriented roughly east-west, and thus parallel the overall strike of the Kistler Valley fault zone. This fault zone is roughly parallel to, and presumably of the same age as, the Eckville fault to the north (see Figure 70). It is possible that both structures represent rare views of an imbricate fan or splays that generally are blind in this part of the Great Valley and that may produce the large wavelength, east-west-trending, late open folds noted at Stop 7. Both of these faults formed later than the Game Preserve fault of probable Taconic age that carried the Shochary sequence, and possibly the piggybacked slices of the Hamburg klippe as well, over the Martinsburg Formation. This older fault is synformally folded along with the Shochary sequence, perhaps during the formation of the Kistler Valley fault.

Things to note at south end of Stop 8:

1) The number of faults and overall disruption of this outcrop by faulting can best be seen from across the road nearer Jordan Creek. Although it is impossible to climb up to a number of the faults, several can be seen at ground level. At least one of these will be labelled by a sign placed directly on the outcrop.

2) Note that nowhere in this outcrop can you find the well-developed and diagnostic ribbon slate with its characteristic orange striping so common in the Bushkill at Stop 7. Also nowhere can you find the very thick-bedded carbonaceous slates to be seen in the Pen Argyl at Stop 9. Instead it is much more common to find finely laminated beds of calcisiltite that typify the New Tripoli.

3) In several places, difficult to find, there is significant pressure solution along slaty cleavage surfaces that give the appearance of minor normal faults in bedding. These offsets are generally less that a centimeter.

4) A close examination of slickenline orientations shows that a number of faults in this outcrop have a component of strike slip motion. Recent work (Steven Wojtal, personal

communication) has documented that on some major thrusts, up to 25% of the movement is strike slip roughly perpendicular to the thrust transport direction.

Things to note at the north end of Stop 8:

1) The important structures to note here are fault related: a) complex folding of slaty cleavage, b) the intensive development of strain-slip cleavage, c) the concentration of vein quartz with minor calcite in the hinges of folds in slaty cleavage as well as parallel to both slaty cleavage and crenulation cleavage planes. It is this development and concentration of quartz during faulting and strain-slip cleavage development that provides one of the most helpful markers for faults in areas of little outcrop. By itself, the presence of quartz should not be the basis for postulating a fault, but combined with other structures it is very helpful.

2) The orientation of some of the folds in slaty cleavage is highly variable and is at a high angle to the strike of the thrust faults in the south half of the stop. Either the folds are getting rotated in the plane of the fault, or perhaps early movement along the Kistler Valley fault was dominated by north or northwest transport and late movement was dominated by more east-west strike-slip faulting.

	Return to buses and continue westward on Weidasville Road.
	0.4
50.8	Weidasville. Road intersection, continue straight towards Lyons Valley.
	0.2
51.0	Cross Jordan Creek.
	0.1
51.1	New Tripoli exposures on right.
	0.3
51.4	Kistler Valley fault crosses near the bend in road here.
	0.1
51.5	Sheared New Tripoli rocks on right.
	0.2
51.7	Stop sign at "T". Turn left on Lyons Valley Farm road heading towards PA 100.
	1.2
52.9	Cross Lyons Creek. Exposures of New Tripoli Formation. Bear left on Raber Road.
	0.2
53.1	Intersection with PA 100. Turn right heading north towards Pleasant Corners. For about 3 miles going north we will be traveling across the Shochary syncline.
	0.5
53.6	New Tripoli Formation exposed on right in the overturned south limb of the Shochary syncline.
	1.0
54.6	New Tripoli Formation.
	0.4
55.0	More New Tripoli Formation.
	0.9
55.9	Shochary Sandstone exposed on right in the core of the Shochary syncline.
	0.4
56.3	Cross Jordan Creek. We are in the upright north limb of the Shochary syncline.
	0.7
57.0	Stop sign, intersection with PA 309. Turn right on PA 309 south. New Tripoli exposed in borrow pit north of PA 309.
	0.3
57.3	Village of Pleasant Corners. Turn left on Germansville Road.
	0.2
57.5	New Tripoli exposed to right.
	0.4
57.9	Bear left at "Y" in road heading toward Germansville.
	0.1
58.0	Crossing trace of south-dipping Eckville fault, which marks the northern limit of the Shochary Ridge rocks and thrusts the New Tripoli Formation on the south over the Bushkill Member of the Martinsburg Formation on the north. This fault was first recognized by Behre (1933).
	0.4
58.4	Village of Germansville.
	0.4
58.8	Bushkill Member of Martinsburg Formation on right.
	0.1

58.9 Road intersection in Germansville. Continue straight.
0.2

59.1 Bear left at "Y" heading toward Tamaqua.
0.3

59.4 The break in slope in the hill straight ahead marks the contact between the Bushkill Member of the Martinsburg Formation to the south with the Ramseyburg Member.
0.3

59.7 Bushkill-Ramseyburg contact. Siltstone and sandstone of the Ramseyburg Member exposed to the right just north of the contact.
0.6

60.3 Contact between the Ramseyburg Member and the overlying Pen Argyl Member of the Martinsburg Formation. Bake Oven Knob on Blue Mountain to the distance.
0.7

61.0 Slates of the Pen Argyl Member exposed on the right. Slaty cleavage dips steeply to the southeast. Road intersection, stop sign. Turn right toward Palmerton. For the next several miles we will be riding in the strike valley underlain by the Pen Argyl Member. Overturned and faulted Tuscarora Sandstone, Shawangunk Formation, and Clinton Formation underlie Blue Mountain to the left.
2.1

63.1 Exposures of slate in Pen Argyl Member to right.
0.3

63.4 Village of Lehigh Furnace.
0.4

63.8 Turn right towards Slatedale.
0.2

64.0 Dumps of slate quarry in the Pen Argyl Member to the right.
0.3

64.3 Entering village of Slatedale. Hills to right are underlain by sandstone and slate of the Ramseyburg Member. Several slate dumps can be seen north of the Ramseyburg hill. Continue straight on Main Street in Slatedale until the fire station.
0.6

64.9 Slatedale Post Office.
0.2

65.1 Turn right into parking lot of Citizens Fire Co. No. 1 Fire house.
0.1

65.2 LUNCH

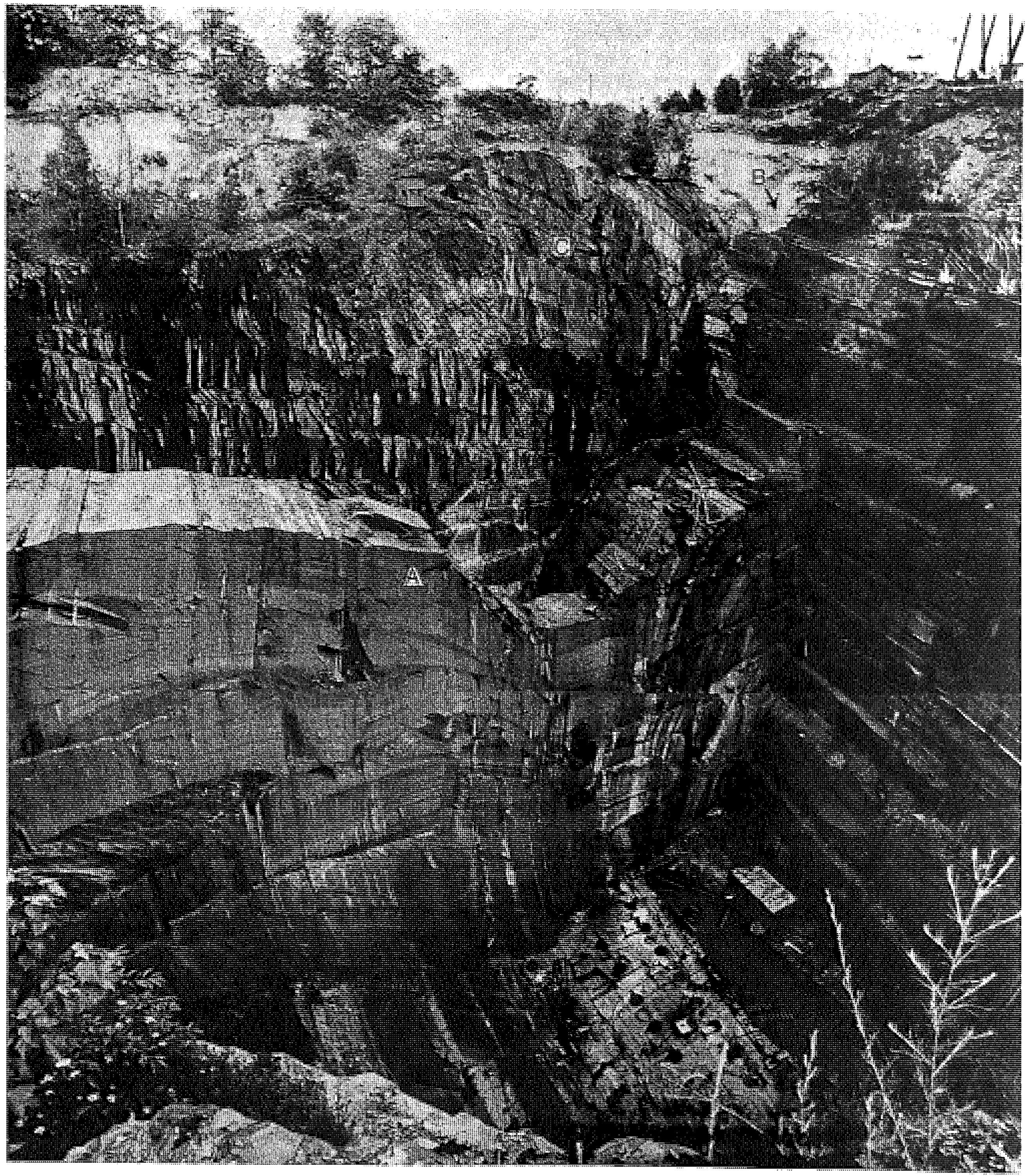
Return to Main Street and turn left.
0.2

65.4 Turn right.
0.2

65.7 Turn left into the Penn Big Bed Slate Co. quarry and drive to the top of the quarry.
0.2

65.9 Stop 9.

Figure 94. Overturned anticline in the Pen Argyl Member of the Martinsburg Formation in the quarry of the Penn Big Bed Slate Company, 600 m (2000 ft) north of Slatedale, Pa. View looking eastward. A rib about 30 ft wide separates the presently active quarry from a hole beyond. Cleavage dips steeply southward (to the right). The face in the center of the photo (at A) is parallel to the cleavage. A syncline (at B) is seen in the southeast corner of the quarry. A steeply dipping reverse fault (at C) is located in the upright limb of the fold. Figure 95 depicts the main structural features in the quarry. Overhead cables, with a breaking strength of 490 tons and supported by the derricks to the upper right, hoist blocks of slate out of the quarry. The smooth faces in the quarry have been cut with a wire saw. Compare these with the rough blasted beds just under the shed on top of the east wall. The shed is an old signalman's house. The signalman directs the hoisting operations upon vocal instructions from men in the quarry below. These instructions include such gems as "YOO HOO! HOIST AWAY MATEYS!" or "STUFF IT UP YOUR DERRICK, JACK"). The engine houses and mill are located to the south beyond the brink of the quarry. Three foot wide calyx holes can be seen in the floor of the quarry. These are about 15 feet deep and are sites where the wire saw standards and sheaths are placed and between which the wire cuts the rock. The Pen Argyl is thick bedded here, and one bed, the Penn Big Bed, is 3.7 m (12 ft) thick. The top of that bed is located just under the signalman's shed.



Stop 9. Penn Big Bed Slate Company Quarry (Manhatten Mine); Stratigraphy, Structure, and Economic Geology of the Pen Argyl Member of the Martinsburg Formation

WARNING: AVOID DEATH--KEEP BACK FROM VERTICAL QUARRY WALLS!

Avoid interfering with operations in the mill.

The 32nd Field Conference of Pennsylvania Geologists visited this quarry 17 years ago (Epstein and Epstein, 1967), but it is still an excellent locality to compare the stratigraphic characteristics of the upper (Pen Argyl)

Member of the Martinsburg Formation with those of the lower (Bushkill Member) seen at Stop 7. We can also compare typical folds in the Martinsburg with those of younger Silurian and Devonian rocks to be seen at the next stop. If the quarry and mill are in operation at the time of our visit, we can observe quarry and milling practices. Please be sure to avoid interfering with operations in the mill.

Stratigraphy

Many beds in the Pen Argyl Member of the Martinsburg Formation in this quarry exceed 10 feet (3 m) in thickness (fig. 94), although laminated slates are not uncommon. As discussed under the section on stratigraphy and at Stop 7, this is in sharp contrast with thicknesses of beds in the Bushkill Member which have never been seen to exceed 6 inches (15 cm). In general, medium-dark-gray to dark-gray evenly bedded slate grades up into thinner grayish-black carbonaceous slate. Laminae to beds of graywacke as much as four feet (1.2 m) thick may form the base of some of the cycles. Examples of graywacke, some with intraformational convolutions, may be seen on the dumps west of the mill. You will note that several hundred feet of rock are exposed in a continuous section in this quarry alone. The outcrop width of the Pen Argyl is 13,000 feet (4000 m) at this locality. Excellent exposures of the Pen Argyl Member are found along the Lehigh River, about 2.6 miles (4.1 km) to the northeast, and many folds similar to those in this quarry have been mapped. Few faults have been recognized. Thus, it is not surprising that we have estimated the thickness of the Pen Argyl Member to be more than 5,000 feet (1500 m) in this area.
Petrographic characteristics of the Pen Argyl are given in Epstein, Sevon, and Glaeser (1974).

Structure

The overturned anticline and syncline to the south seen in this quarry (figs. 94 and 95) are typical of the folds mapped in the Martinsburg Formation of the Lehigh Valley. Folding was dominately passive and slaty cleavage is the dominant secondary structure. The cleavage forms a fan with an angle of about 24°. Much evidence indicates that the cleavage formed under conditions of low-grade regional

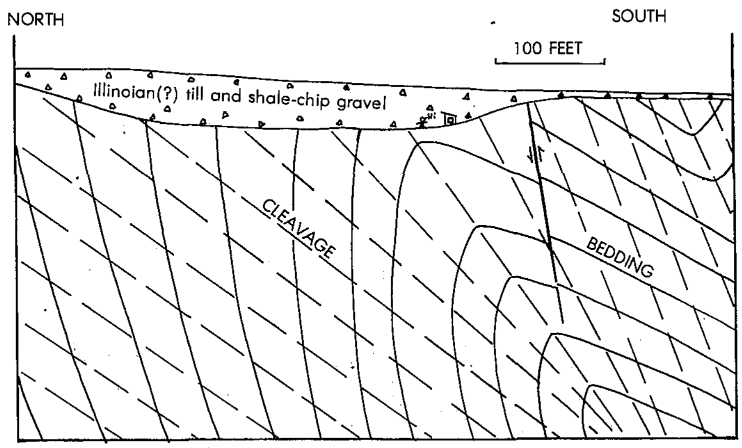


Figure 95. Diagrammatic cross section showing overturned folds, a reverse fault and fanning of cleavage in the Penn Big Bed Slate Company quarry.

metamorphism, by processes involving pressure solution, new mineral growth, and some mineral reorientation, but that is a subject beyond the scope of this stop. Slickensides of calcite and quartz fibers up to 2 inches (5 cm) thick on many bedding surfaces indicate that flexural slip was a subordinate process in producing the folds. The slickensides both cut and are cut by the cleavage, indicating that cleavage formation both preceded and followed bedding slippage.

Economic geology of slate

The first slate quarry in Pennsylvania was opened in 1812 near Bangor, and since then more than 400 quarries have been opened in the slate belt in the Martinsburg Formation of Northampton and Lehigh Counties. At present, only about half a dozen quarries are active. Nevertheless, Pennsylvania is the third largest producer of slate in the United States, following Vermont and Virginia, according to the U.S. Bureau of Mines.

The Penn Big Bed Slate Quarry, the only active quarry in Lehigh County, is divided into two parts that are separated by a rib in the middle (fig.94). The eastern section is about 350 feet deep, and the western part, now being worked, is about 200 feet deep. An abandoned pit, now flooded, lies immediately to the west.
The slate removed from the quarry is used for roofing shingles, structural slate, floor tile, and fireplace facing. In the past, additional uses

included blackboards, flagging, aquaria bottoms, and billiard table tops. About 4,000 squares of roofing slate are produced each year.

Quarrying operations and profitability are contolled by the geologic setting of a quarry. Removing large quantities of "top" (weathered slate, colluvium, and till) might be prohibitively costly because stripping operations and extensive cribbing might be required.. The present active hole might not have been opened had the owners known that the overburden thickened from about 8 feet (2 m) in the older western hole to about 45 feet (14 m) in the present opening. Had an exploratory drilling program been conducted, the quarry would have been located at another site.

The shape and depth of the quarry is controlled by the steepness of bedding. For the most part, thick clear "runs" are followed down the near vertical beds in the north limb of the anticline. (The deepest slate quarry in the United States, by the way, the collapsed Parsons quarry in Pen Argyl, Pa., 25 miles (40 km) to the northeast, is reported to have been about 900 feet (274 m) deep. It also followed vertical beds).

Cleavage is the feature that makes a rock a slate. It should be continous through the rock, and it should not be curved or irregular ("curled"). Warped slaty cleavage is generally associated with a second-generation crenulation cleavage. Minor northwest-dipping crenulation cleavage was seen in the syncline in the southeast corner of the quarry. The length of a piece of slate that can be removed is determined by the thickness of a clear bed and the angle between bedding and cleavage. If the angle between cleavage and bedding is high, the length of a piece of slate derived from that bed is relatively short. Conversely, if the angle between the two is low, the piece of slate is long. The thickest bed in the quarry, the Penn Big Bed, is 12 feet (3.7 m) thick (measured orthogonally) and is 21 feet (6.4 m) thick along the "split" (cleavage). The dip of cleavage is also important because it generally forms the floor of the quarry and may be inconveniently steep. In the Penn Big Bed quarry, cleavage averages about 55°, and other fractures in the rock are used to form the quarry floor.

Joints may facilitate quarrying if their orientation and spacing are favorable for the size of block being removed. If, however, the joints are too close together or are at low angles to each other, large blocks of slate cannot be obtained and the rock is worthless.

Grain ("sculp" of quarrymen) is important in removing slate from quarries because fractures readily form parallel to it during blasting, and the sides of many quarries therefore parallel the grain. This direction of splitting is believed to be controlled by the elongation of prismatic minerals, mainly quartz, in the direction of tectonic transport, approximately at right angles to cleavage and bedding, generally in the downdip direction of slaty cleavage.

Most of the rock removed from the slate quarries in eastern Pennsylvania wind up on slate dumps that form conspicuous hills, or are dumped in adjacent abandoned holes. Approximately 80 percent of the rock removed from the Penn Big Bed quarry is waste.

Hazards associated with slate quarries are numerous. Drownings of swimmers and scuba divers have been reported in flooded quarries. A car was accidently driven into an quarry just south of the Penn Big Bed quarry. Rockslides have occurred, especially along "rotten ribbons" (fault zones), killing and maiming workers below. Many abandoned quarries have been used for dumping of garbage, creating the possibility for ground water pollution.

Quarry methods, including the use of the calyx drill and wire saws, will be explained at the stop. For additional information, refer to Behre (1933), Stickler, Mullen, and Bitner (1951), and Epstein (1974).

	Retrace route out of quarry.
	0.1
66.0	Turn left out of quarry property.
	0.8
66.8	"T" in road, turn left heading to Lehigh Furnace. Talus-covered slopes of Blue Mountain underlain by quartzites of the Shawangunk Formation. For the next several miles we will be traveling in the strike valley of the Pen Argyl

Member of the Martinsburg Formation.
6.9
73.7 Turn right on PA 309 heading north toward Tamaqua.
2.2
75.9 Approximate location of covered contact between the Shawangunk Formation and underlying Martinsburg Formation. View of Great Valley to left.
0.8
76.7 Crest of Blue Mountain. Appalachian trail crosses the road.
0.5
77.2 Sandstones of the Clinton Formation exposed on the left.
0.7
77.9 Faulted contact of steeply dipping Bloomsburg Red Beds and underlying Clinton Formation.
1.1
79.0 Abandoned quarry in the Bossardville Limestone and Decker Formation to the left.
0.5
79.5 Nis Hollow Siltstone Member of the Mahantango Formation exposed behind barn to right.
0.1
79.6 Turn right on PA 895 toward Andreas. We will be driving in the strike valley underlain by siltstones of the Mahantango Formation. To the left the hills are underlain by siltstone and very fine-grained sandstones of the Trimmers Rock Formation.
2.4
82.0 Siltstone of the Mahantango Formation in borrow pit to left.
0.5
82.5 Nis Hollow Siltstone Member exposed to left.
0.4
82.9 The quarry in Chestnut Ridge to right is the site of Stop 10. The taller mountain well beyond is Blue Mountain held up by the Tuscarora Sandstone.
0.4
83.3 Village of Andreas.
0.5
83.8 Road intersection at Andreas Post Office. Turn right.
0.5
84.3 Blacktop batching plant to left.
0.2
84.5 Stop 10.

Stop 10. Structure and Stratigraphy of Upper Silurian to Middle Devonian rocks, Huss Stone Quarry, Andreas

Rocks of Late Silurian through Middle Devonian age are miserably exposed between Bossardsville and Schuylkill Haven, Pa., a distance of 51 miles (82 km). Only by mapping float and scattered outcrops could the stratigraphic section in fig. 72 be compiled. The quarry at Andreas is a geological oasis (fig. 96) in the midst of this desert of poor exposure. Here we can examine the stratigraphic characteristics of rocks from the Poxono Island Formation through the Selinsgrove Limestone and compare typical Alleghenian folds with those in the Martinsburg Formation seen at stop 9. Rocks from the Bossardville Limestone and some of the Poxono Island Formation in the quarry are crushed for construction stone and the Palmerton Sandstone is exploited for sand.

The units exposed in the quarry are generally a transgressive sequence, having been deposited in a wide variety of environments, including supratidal and intertidal flats, barrier bar or beach, subtidal, and possibly fluvial (see section on stratigraphy and sedimentologic history of Silurian and Devonian rocks). The characteristics of the units are given in table 9. Some petrographic details are given in Epstein and others (1974).

The quarry area has been mapped in detail (fig. 97). The sand quarry in the Palmerton Sandstone southeast of the stone quarry is in an open syncline which becomes overturned to the southwest. The stone quarry is in an overturned anticline, nicely exposed on the east wall (fig. 101). Small-scale satellitic folding is common in the Bossardville Limestone and Poxono Island Formation. The northwest overturned limb is contorted in a 50-foot (15 m)- wide shear zone which contains some backlimb thrusts (cross section in fig 97; fig. 102). Bedding-plane

Table 9. Stratigraphic units exposed in the Huss Stone quarry at Andreas, Pa. Thicknesses are in parentheses.

MARCELLUS SHALE (not measured)

Laminated dark-gray shale. Deep-water.

SELINSGROVE LIMESTONE (40 ft; 12 m)

Grayish-orange- and moderate-yellowish-brown-weathering, medium-gray to medium-dark-gray, very fine-grained, medium-bedded, very fossiliferous (brachiopods, bryozoans, corals, and gastropods) silty and shaly limestone and dark-gray pyritic shale. Dark-gray shale in lower 8 feet (2.4 m) may be the Needmore Shale. Siderite of the Hazard Paint ore, near the middle of the Selinsgrove, has been mined in the surrounding area for paint pigment. The Hazard is not exposed and is probably less than 2 feet (0.6 m) thick. Subtidal.

PALMERTON SANDSTONE (50ft; 15 m)

Very pale-orange to dark-yellowish-orange very coarse-grained to conglomeratic (quartz pebbles as much as one inch (2.5 cm) long, but averaging about 1/4 inch (6 mm) long), medium- to thick-bedded, (some beds are more than 4.5 feet (1.4 m) thick), indistinctly to planar bedded sandstone with siliceous siltstone and white to light-gray chert replacing some sandstone at the lower transitional contact with the Esopus-Schoharie Formation, undivided. Rare low-angle crossbedding, ripples, and scoured surfaces. Probable barrier bar.

SCHOHARIE-ESOPUS FORMATIONS, UNDIVIDED (16 ft; 5 m)

Weathered very light-gray to light-gray, thin- to medium-bedded quartzose siltstone with rare floating coarse sand grains and very fine-grained sandstone. Some chert at bottom. The trace fossil Taonurus, which is characteristic of these rocks farther east, was not seen in the quarry. Subtidal.

ORISKANY GROUP (38 ft; 12 m)

Weathered pale-yellowish-orange to dark-yellowish-orange and very light-gray to yellowish gray, partly hematitic, planar bedded to partly crossbedded, coarse-grained, slightly conglomeratic sandstone with rounded quartz pebbles as much as one inch (2.5 cm) long and spiriferid molds in lower 8 feet (2.4 m), and fossiliferous very light-gray to medium light gray chert, with siltstone and coarse-grained sandstone in upper 30 feet (9 m) Lower contact abrupt and irregular and believed to be unconformable. Beach and shallow subtidal.

STORMVILLE FORMATION(?) (35 ft; 11 m)

Very light-gray to light-gray, and very pale-orange to grayish-orange and light-brownish-gray, planar bedded and crossbedded (trough crossbedding), lenticular (beds may disappear over a distance of a few hundred feet, as seen on southeast wall of quarry), thin- to thick-bedded, partly burrowed, coarse-grained sandstone and conglomeratic sandstone with rounded to subrounded quartz pebbles as much as 1 inch (2.5 cm) long and lesser chert clasts as much as 1/8 inch (3 mm) long, with many scoured basal surfaces. The cross bedding is unidirectional, suggesting a fluvial environment, but thin conglomerates are interbeded with weathered, fossiliferous, light- to bluish-gray, dark-yellowish-orange to moderate-yellowish-brown siltstone, very fine-grained, thin-bedded and rippled sandstone, and rare laminae of medium-light-gray shale at the top, so the upper part at least, is marine. The brachiopods and crinoids are poorly preserved and are not useful in pinning down the age of this unit (John Pojeta, oral communication, 1984).

ANDREAS RED BEDS OF SWARTZ AND SWARTZ (1941) (65 ft; 20 m)

Grayish-red, dusky-red, to moderate-reddish-brown, and lesser light-grey to pinkish-gray, laminated to thick-bedded shale, siltstone, and fine- to coarse-grained sandstone, slightly conglomeratic in places. Planar bedded, crossbedded, rippled, and burrow mottled. Some channels. Crossbeds trend to the northwest, perpendicular to strike of beds. Channels trend parallel to strike of beds. Some upward-fining cycles (fig. 98), but lacks distinctive epsilon crossbedding to suggest a meandering fluvial origin, probably a tidal deposit. The lowest bed contains ostracodes and a pelecypod (see paper by Denkler, this volume).

DECKER FORMATION (59 ft; 18 m)

Laminated to thick-bedded, light- to dark-gray, fine- to medium-grained shaly and silty limestone and calcareous siltstone and shale, and a few thin beds of fossiliferous (ostracodes, conodonts, bryozoans, brachiopods, and pelecypods). coarse-grained limestone. Partly nodular, ripple laminated, graded, and burrowed. Weathers grayish orange to moderate-yellowish brown. Fossils suggest an uppermost (Pridolian) age (see paper by Denkler, this volume). Carbonate has been leached to depths of several tens of feet in most surface exposures. Shallow subtidal.

BOSSARDVILLE LIMESTONE (70 ft; 21 m)

Medium-gray to dark-gray, laminated and irregularly thin-bedded, very fine-grained, graded, cross-laminated, partly mudcracked, shaly limestone, with scattered leperditiid ostracodes (fig. 99). Intertidal.

POXONO ISLAND FORMATION (100+ ft; 30+ m)

Cyclical micro-laminated (algal?) to thin-bedded medium-gray (with a purplish tinge) limestone, locally with dolomite and shale intraclasts (fig. 100), laminated greenish gray to medium-gray shale and greenish-gray dolomite, grading up into massive to partly laminated greenish-gray mudcracked dolomite which may be irrgularly replaced by medium-dark-gray dolomite. The cycles are from one foot (30 cm) to nearly 20 feet (6 m) thick. Contains rare leperditiid ostracods. Grades downward into the Bloomsburg Red Beds through interbedded green and red shale and siltstone. Supratidal.

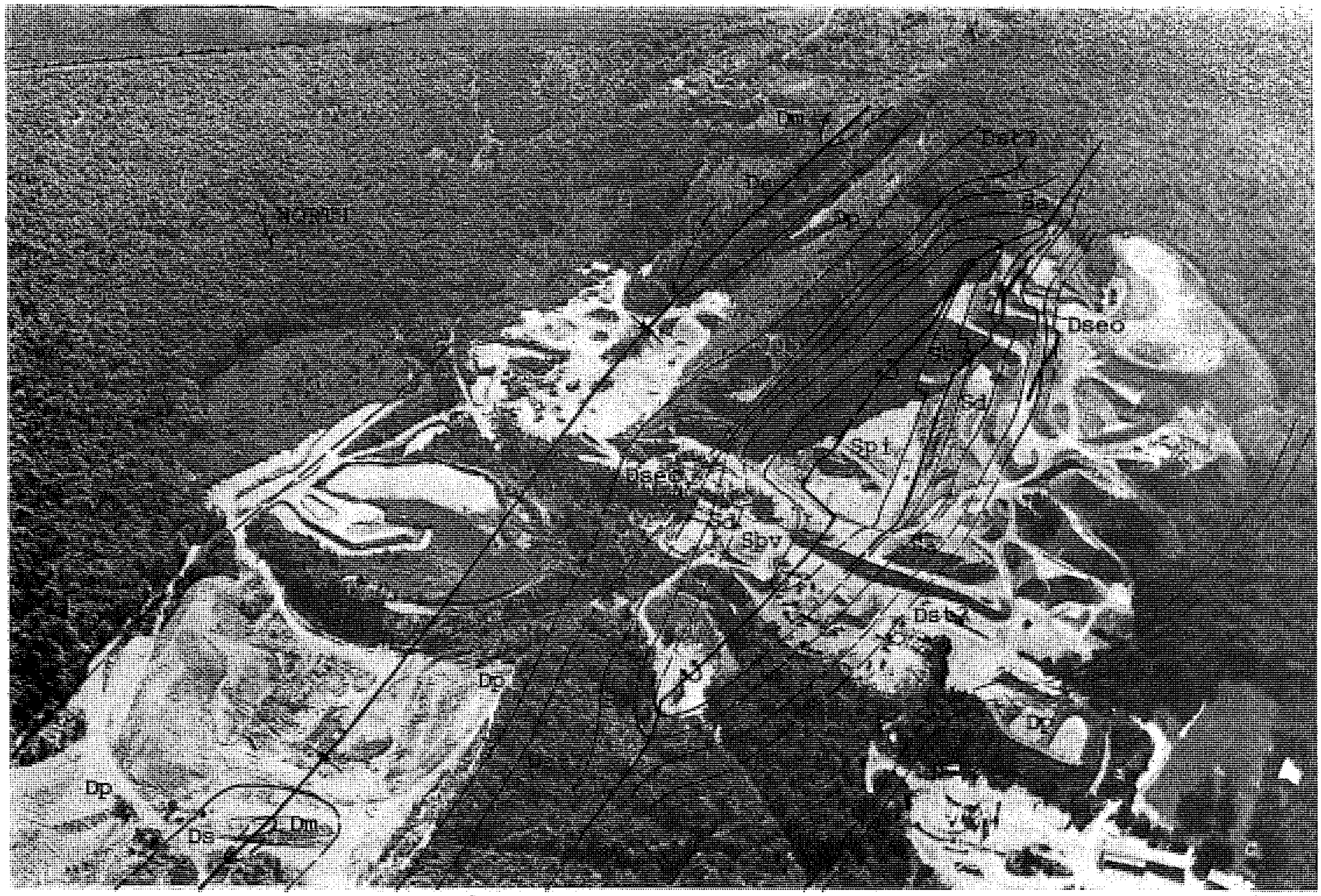
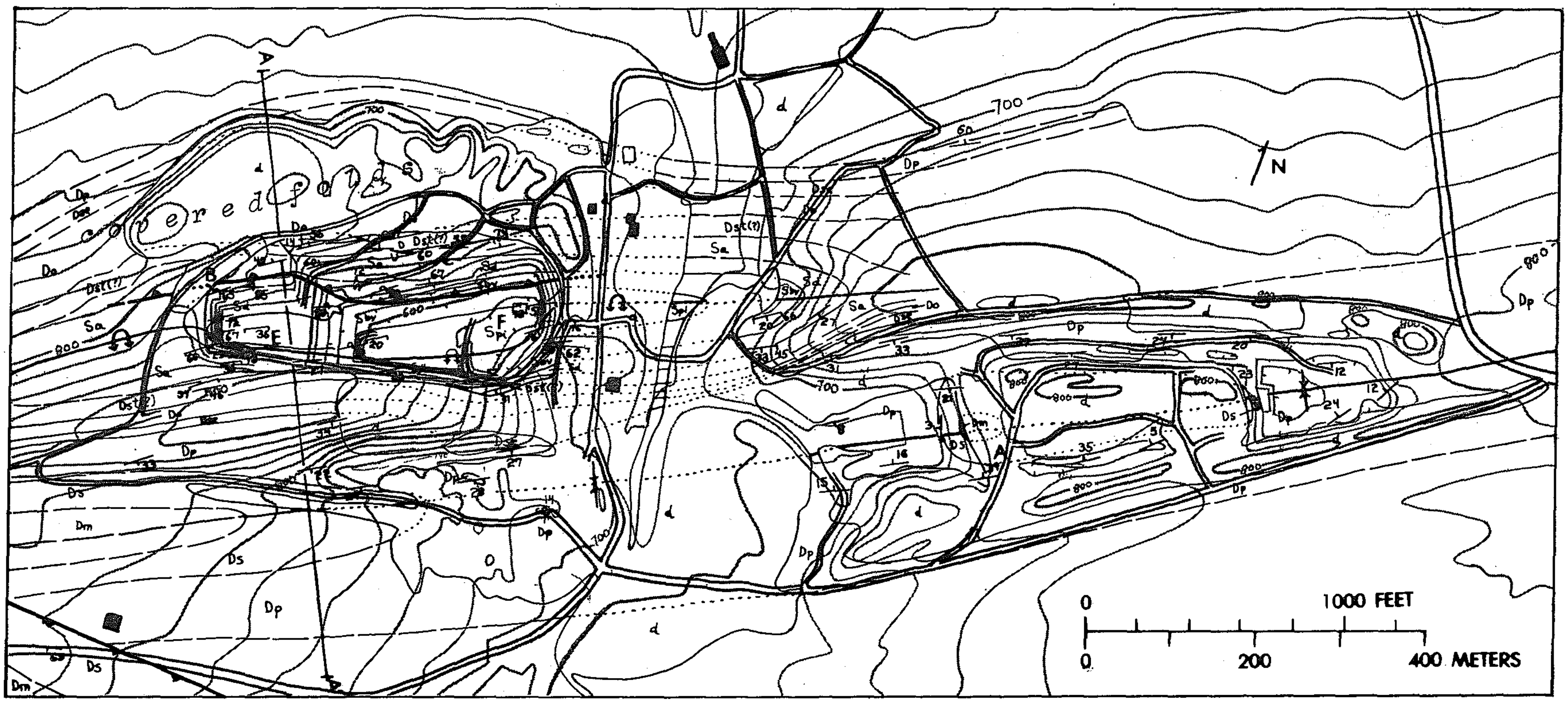


Figure 96. Oblique aerial photograph and geology of Huss Stone quarry at Andreas, Pa., looking southwestward. Dm, Marcellus Shale; Ds, Selinsgrove Limestone; Dp, Palmerton Sandstone; Dseo, Oriskany Group and Schoharie and Esopus Formations, undivided; Dst(?) Stormville Formation(?); Sa, Andreas Red Beds of Swartz and Swartz (1941); Sbv, Bossardville Limestone; Spi, Poxono Island Formation.

slickensides, indicating flexural slip folding, is cut in many places by cleavage, but in other places cleavage is sigmoidally dragged along slickensides. In the north corner of the quarry cleavage has been rotated to a gentle northwest dip, and a second-generation crenulation cleavage has developed. The sequence of structural development therefore appears to be 1) flexural slip folding, 2) continued folding with development of pressure-solution cleavage, 3) additional flexural slip that dragged cleavage, 4) faulting of the northwest limb of the fold with rotation of some of the cleavage, and 5) development of local crenulation cleavage.

Cleavage is the dominant secondary structure in the quarry. It consists of dark folia up to a several millimeters wide composed of the residue of the original carbonate rock (micas, some quartz, iron oxides, and dark (organic?) material) separating interfolial areas (microlithons) that exhibit little strain, and which are less than 1 mm to more than 10 cm wide. Two trends of cleavage folia that crosscut at low angles indicate that there were at least two generations of cleavage development (fig. 103). The micas appear to have grown parallel to the cleavage folia because they are mostly larger than micas in the microlithons. This is a pressure solution cleavage, which is abundantly described in the literature, and which shows such typical features as truncation of fossils (fig. 104) and offset of bedding laminae. The cleavage did not



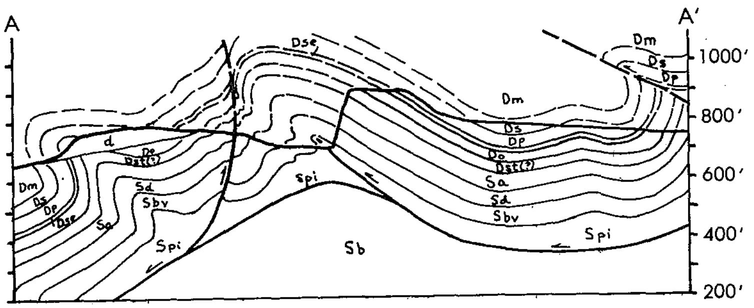


Figure 97. Geologic map and section of the Huss Stone Quarry at Andreas, Pa. d, dump; Dm, Marcellus Shale; Ds, Selinsgrove Limestone; Dp, Palmerton Sandstone; Dse, Schoharie and Esopus Formations, undivided; Do, Oriskany Group; Dst(?), Stormville Formation(?); Sa, Andreas Red Beds of Swartz and Swartz (1941); Sd, Decker Formation; Sbv, Bossardville Limestone; Spi, Poxono Island Formation; Sb, Bloomsburg Red Beds. Standard structural symbols are used for bedding, cleavage, folds, faults, and contacts. Folds, faults, and contacts are dashed where approximately located or inferred, and dotted where concealed. Letters A-F are possible stop localities described in the text. Topographic map prepared from altimetry on low altitude aerial photograph. Contour interval, 20 feet.

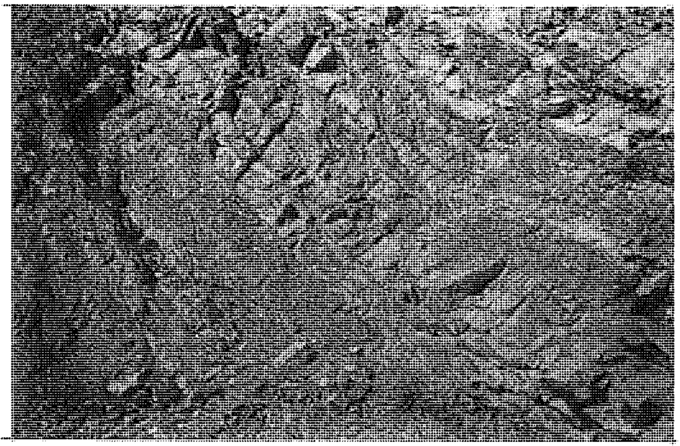


Fig. 98. Upward fining cycles at top of Andreas Red Beds and lower Stormville Formation(?). Contact in middle of photograph at base of massive sandstone.

Fig. 99. Negative print of acetate peel showing laminated, graded, and cross laminated shaly limestone in the Bossardville Limestone. Cleavage dips gently to the right.

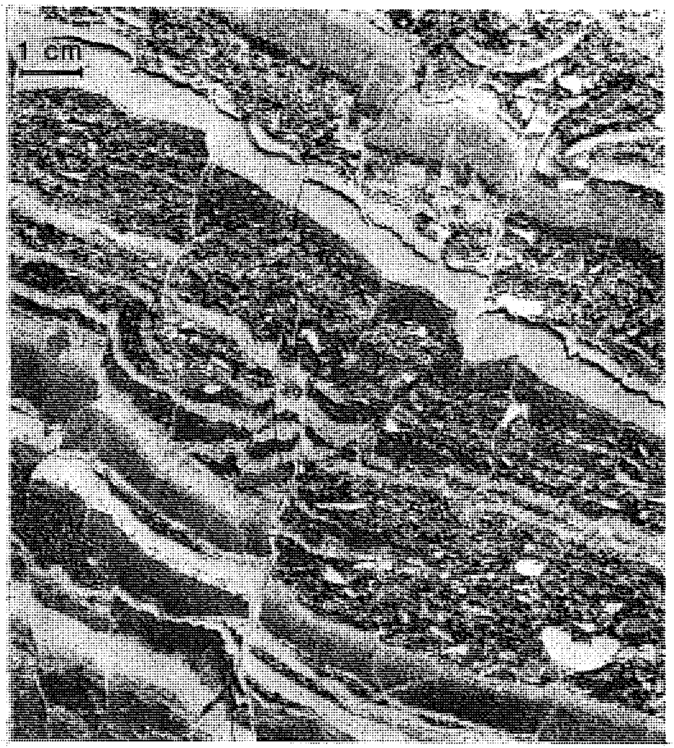


Fig. 100. Negative print of acetate peel stained with alizarin red and potassium ferricyanide showing laminated to thin-bedded intraclastic and micritic calcium dolomite (gray) and limestone (black) in the Poxono Island Formation. Some intraclasts are dolomite (white), apparently ripped up from beds farther onshore during storms.

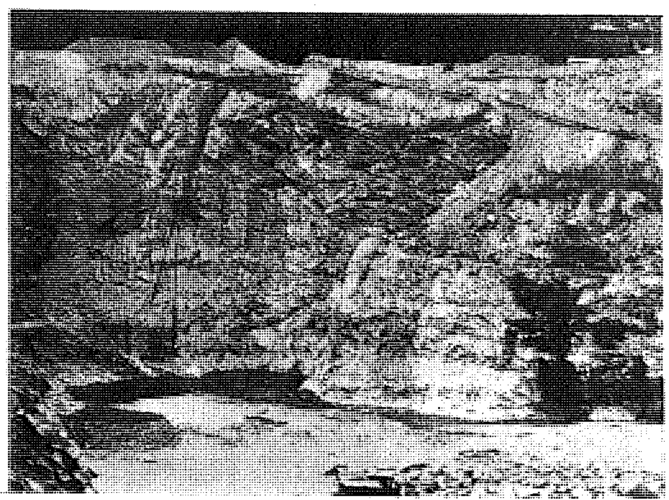


Fig. 101. Overturned anticline in the Poxono Island Formation and Bossardville Limestone in the northeast wall of the Andreas quarry. The axial plane of the fold dips 37° SE. and cleavage fans the fold through an angle of about 30°.

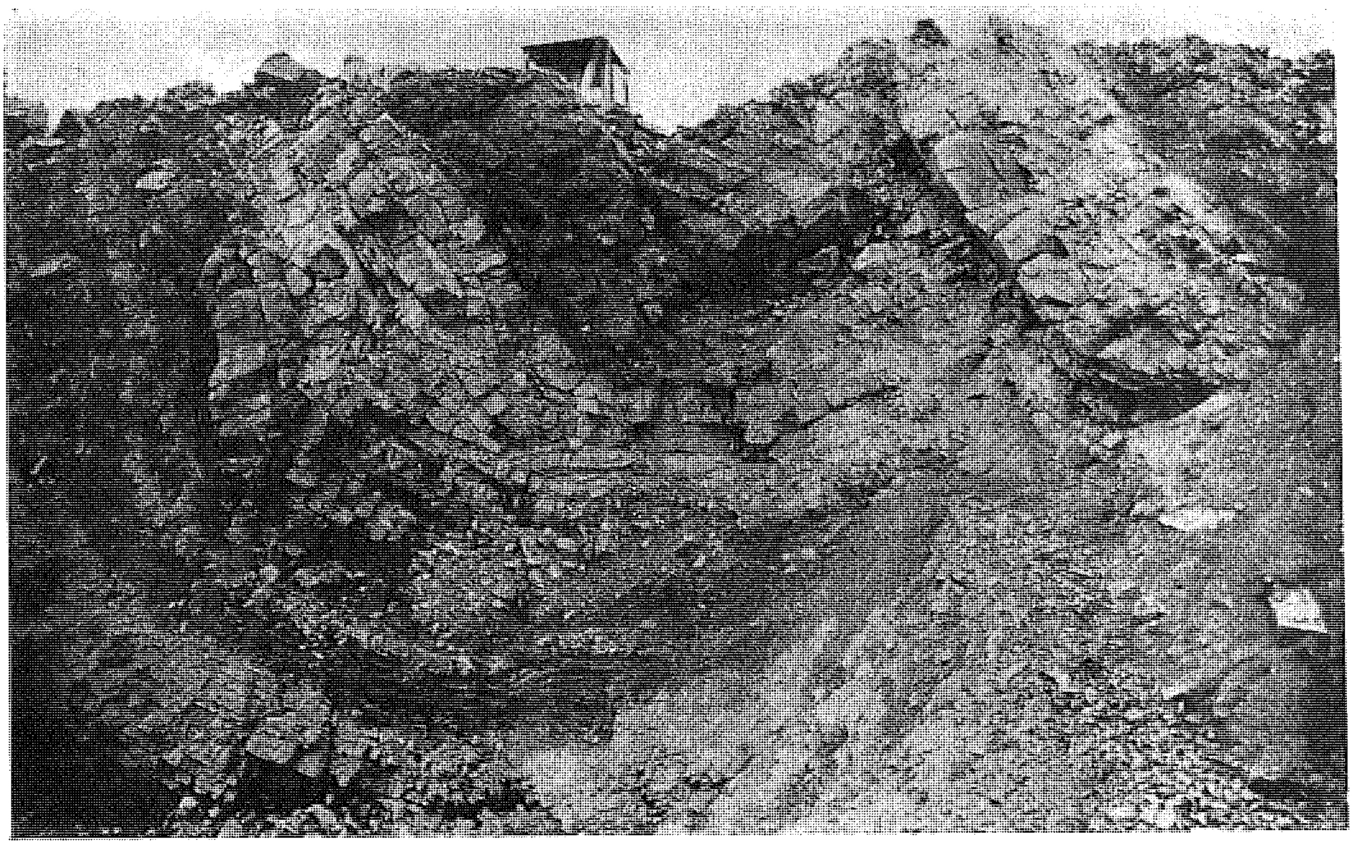


Fig. 102. Backlimb thrust in the overturned limb of the Andreas anticline. The Andreas Red Beds and Decker Formation have been dragged into a small syncline by the upward movement of the more steeply dipping Andreas Red Beds on the right. The fault zone comprises gouge and breccia about three feet (1 m) wide. Slickensides corroborate the upward movement of the northwest plate. Some shortening in the synclinal trough in the southeast limb has been taken up by the development of wedges. Quarry operations has removed this outcrop since this photo was taken.

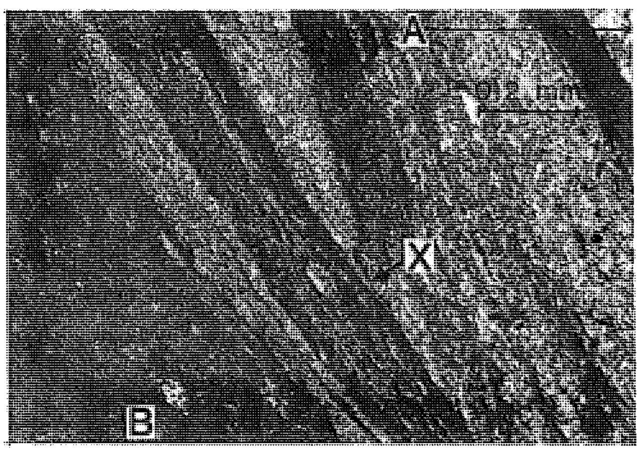


Fig. 103. Photomicrograph showing abrupt contact between cleavage folia (A) and micrite country rock (microlithon, B) in the Poxono Island Formation. Note elongate mica grains parallel to cleavage and angular intersection (11^{o}) of two generations of cleavage at X.

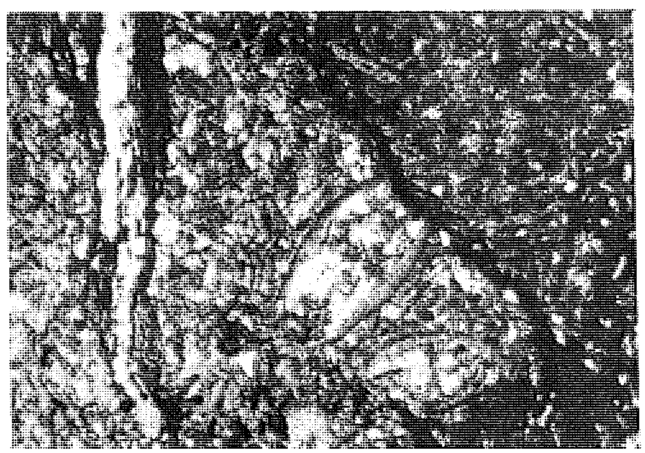


Fig. 104. Photomicrograph of ostracodes truncated at cleavage folia by pressure solution in silty limestone, Decker Formation. Ostracode in center is 0.5 mm long.

Fig. 105. Negative print of acetate peel stained with alizarin red and potassium ferricyanide showing laminated limestone (dark), calcareous dolomite (medium gray), and dolomite (light gray) in the Poxono Island Formation with cleavage developed in steep limbs of crinkles.

develop without some distortion in the microlithons, however, because some cleavage appears to be best developed in the limbs of flattened folds, especially in thinly interlayered rocks of varied lithology (fig. 105). Rotation of beds in the microlithons may be considerable (fig. 106). The amount of shortening due to by pressure solution is great, possibly more than 30 percent, judging from the length-width ratios of compressed mudcrack polygons. One nagging question, common to pressure solution cleavage, is where does the dissolved material (mainly calcite and quartz) go? It apparently may leave the parent rock entirely to be deposited nearby or far away in veins, it may be precipitated as overgrowths, or it may be deposited in voids in opened cores of minor folds (fig. 107). Also, thin zones of quartz (approximately 0.05 mm wide) were seen in many thin sections to have been precipitated between folia and neighboring microlithons.

Because this is an active quarry, new faces are exposed almost every week, and marvellous

Fig. 106. Rotation of microlithons in laminated Poxono Island Formation.

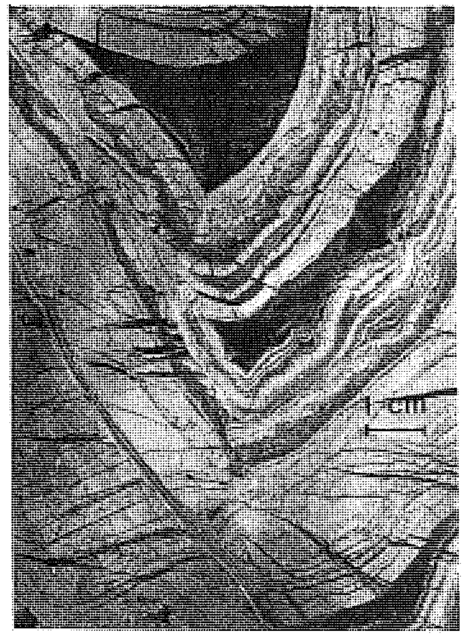


Fig. 107. Negative print of thin section showing sparry calcite and quartz (dark areas) precipitated in opened cores of flattened folds in laminated micrite and calcareous shale. Chalcedonic quartz and carbonate are also precipitated in many tension veins at high angles to cleavage.

outcrops may disappear in a flash with the shovel. Depending on available exposures, the possible stops that we may make in this quarry are shown on fig. 97. These include:

A. Examine the Palmerton Sandstone in the apparently upright open syncline and collect fossils from the Selinsgrove Limestone. Abundant fossils were collected at this locality and were identified by J. T. Dutro and Jean M. Berdan, U.S. Geological Survey. They indicate a late Emsian (Schoharie of New York State) age, and include:

- Pleurodictyum cf. P. lenticulare Hall
- fenestrate bryozoans, indet.
- Pholidops sp.
- Eodevonaria sp. (abundant)
- Parachonetes sp.
- spiriferoid, indet.
- Tentaculites elongatus (Hall)
- Ranapeltis sp. cf. R. trilateralis Swartz and Swain, 1941
- Bollia sp. cf. B. diceratina Swartz and Swain, 1941
- Reticestus? sp. cf. R.? altireticulatus (Swartz and Swain, 1941)

B. View the anticline in the northeast wall; examine sandstones (Ridgeley Sandstone) and cherts (Shriver Chert) of the Oriskany Group.

C. Examine Andreas Red Beds and Stormville Formation(?)

D. Try to find the backlimb thrust shown in fig. 102. Quarry operations have already removed the exposure shown in the figure!

E. Tight fold with satellitic folds in the Poxono Island Formation and Bossardville Limestone.

F. Sedimentary structures in the Poxono Island Formation and Bossardville Limestone; rotated cleavage in the Poxono Island; abundant features of pressure solution cleavage, including effects on mudcracks.

Retrace route back to Andreas.

0.8

85.3 Stop sign in Andreas, turn left heading west on PA 895.

4.2

89.5 Stop sign, intersection with PA 309. Continue west on PA 895 toward New Ringgold.

0.6

90.1 Note that Chestnut Ridge to the left (south) becomes lower and finally disappears as we travel westward. This is the position where the folds we saw at the quarry in Andreas die out and Upper Silurian and Lower Devonian rocks are sandwiched conformably between younger and older rocks. A ramp fault may separate the two rock sequences here. For the next several miles we will be traveling down the strike valley of siltstones of the Mahantango Formation with siliceous siltstone and very fine-grained sandstone of the Trimmers Rock holding up the up the ridge to right.

2.3

92.4 Siltstone of the Mahantango Formation in borrow pit to right.

1.3

93.7 Ridge in middle ground to left is underlain by Upper Silurian-Lower Devonian rocks that dip conformably between rocks above and below.

3.3

97.0 Exposures of the Mahantango Formation on the right. Note southeast dipping cleavage. The beds dip gently to the northwest.

0.5

97.5 Stop sign, New Ringgold. Continue west on PA 895 and PA 443.

0.2

97.7 Cross Little Schuylkill River.

0.8

98.5 Cross Sweet Arrow Fault, along which the Marcellus Shale has been thrust northward onto the Mahantango Formation.

0.2

98.7 Intersection of PA 895 and PA 443. Bear left on PA 895. From this point for several miles south we will be cutting across the regional strike and traversing rocks of the Bloomsburg Red Beds.

0.9

99.6 Bloomsburg Red Beds exposed on left.

2.2

101.8 Village of Drehersville. Exposures of Bloomsburg Red Beds. Hawk Mountain Sanctuary is located 2 miles to the east on the crest of Blue Mountain. The sanctuary is located along a flyway used by birds of prey, such as hawks, osprey, falcons, and eagles, in their northward migration from mid-February through May, and their southward migration during August through November. Thousands of these migrating raptors were shot prior to 1934 when the sanctuary was established to protect them. A visit to this attraction is very rewarding.

0.6

102.4 Village of Molino. The mountain to the left is underlain by the Tuscarora Sandstone in folds that plunge southwestward towards us.

2.0

104.4 Stop sign intersection with PA 61. Turn left on PA 61 south.

0.3

104.7 Cross Little Schuylkill River and enter Schuylkill Gap which cuts through folded and faulted rocks of the Bloomsburg Red Beds, Clinton Formation, and Tuscarora Sandstone.

0.6

105.3 Steeply dipping rocks of the Clinton Formation on left.

0.1

105.4 Tuscarora Sandstone exposed on left.

0.1

105.5 Rocks of the Clinton Formation on left.

0.7

106.2 Village of Port Clinton. Synclinal axis in the Bloomsburg Red Beds.

0.5

106.7 Steeply dipping, faulted, and repeated rocks of the Clinton Formation to left described by Burtner and others (1958).

0.5

107.2 Cross Schuylkill River. Continue south on PA 61 into Reading.

1.4

108.6 Intersection with I 78 and US 222, continue south on PA 61.

4.6

113.2 Traffic light in Shoemakersville, continue south on PA 61.

4.1

117.3 Traffic light, village of Leesport.

3.3

120.6 Pass over US 222. Continue south on PA 61.

2.9

123.5 Turn right onto US 222 south.

2.3

125.8 Turn left on Park Road.

0.1

125.9 Turn right into parking lot of Reading Motor Inn.

END OF TRIP

Have a pleasant trip home.

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