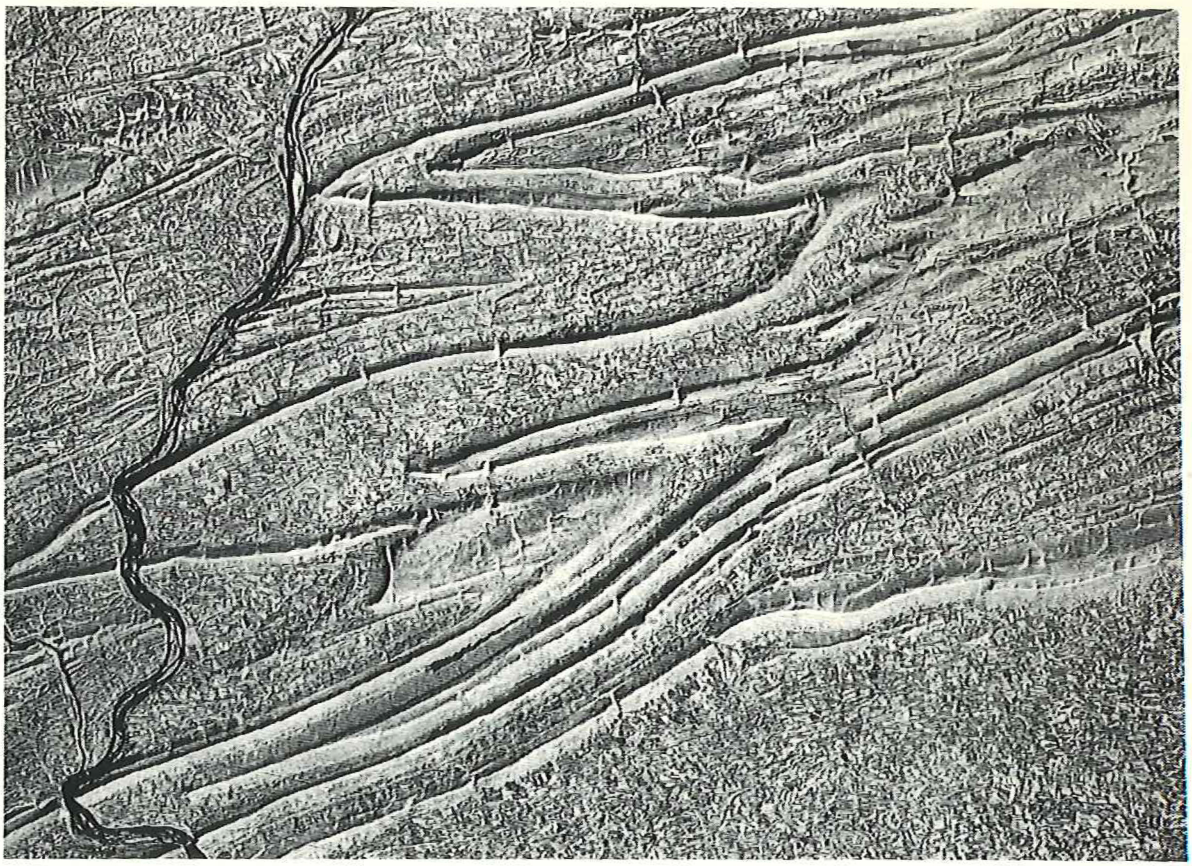
28th International Geological Congress

Sedimentology and Thermal-Mechanical History of Basins in the Central Appalachian Orogen

Field Trip Guidebook T152

Leaders:
Rudy Slingerland and Kevin Furlong



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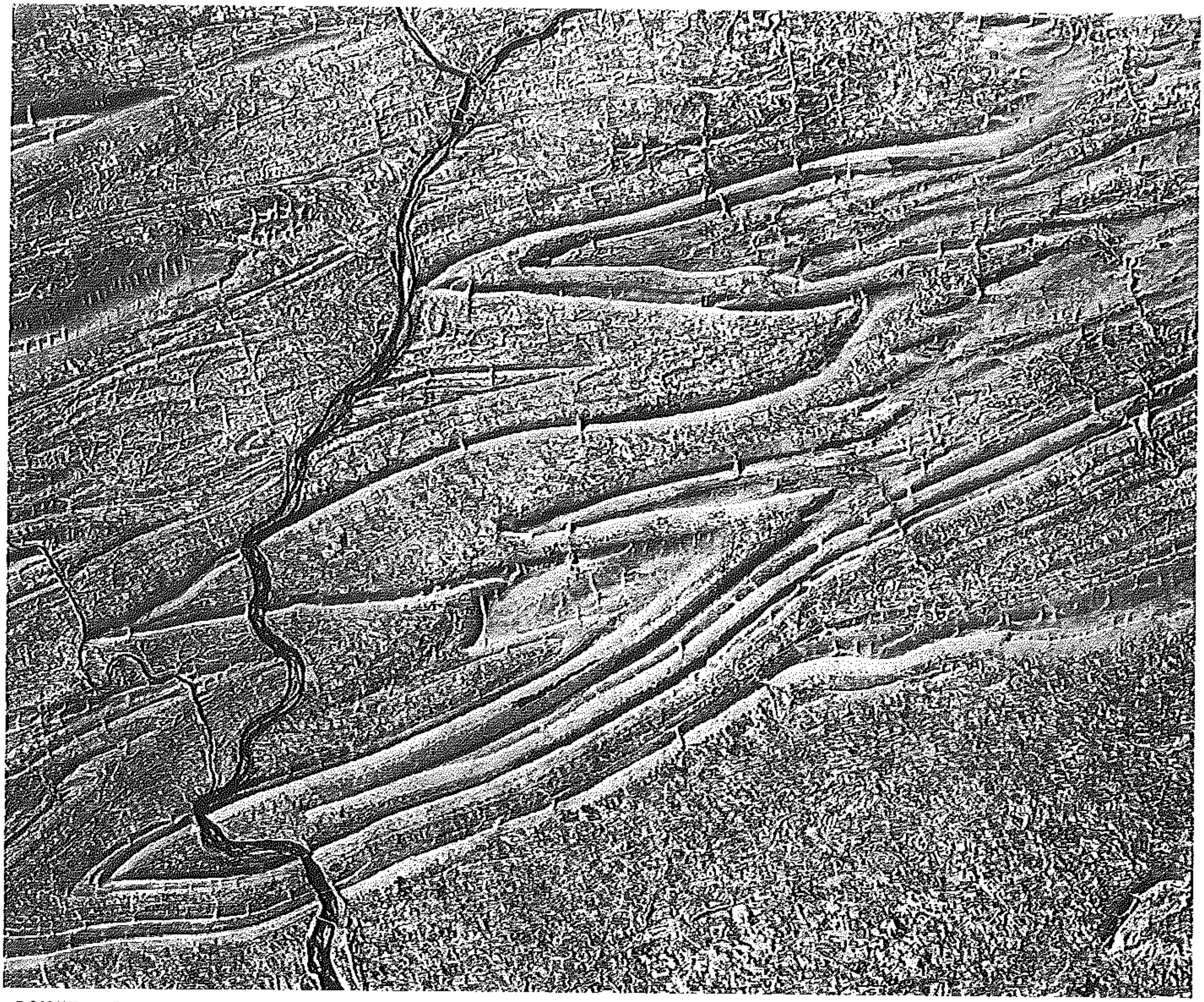

COVER Synthetic-aperture radar image of the Harrisburg, Pennsylvania region (north to the top).

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IGC FIELD TRIP T152: SEDIMENTOLOGY AND THERMAL-MECHANICAL HISTORY OF BASINS IN THE CENTRAL APPALACHIAN OROGEN

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INTRODUCTION

This sedimentary basin workshop and field trip will examine the interplay between basin tectonics and sedimentary deposits in foreland, rift, and, to a lesser extent, passive margin basins of the well-studied central Appalachian orogen (Fig. 1). We will describe the tectonic characteristics of each basin type using thermal-mechanical models of the crust and then discuss the nature of their basin fills in that light. In this way we hope to better understand the relationships between tectonic characteristics---basin geometry, subsidence history, and relative topographic relief, for example---and basin-fill characteristics such as depositional environments and resulting lithofacies, isopach patterns, and on- or off-lap patterns.

The material presented here, limited by publication factors, is arguably the minimum necessary to accomplish these goals. Following these comments, a few paragraphs outline the geologic history of the Appalachian orogen. This is followed by a presentation of a foreland flexural model and its application to the Late Paleozoic foreland basin of the central Appalachians. A similar article presents a rift model and its application to the Mesozoic basins of eastern North America, and in particular, the Newark Basin. No passive margin model is presented as such; the important considerations are presented under the rift model. These are followed by the description of specific field localities in the central Appalachian region, chosen to illustrate the database upon which many of the arguments rest.

Acknowledgments

We thank John S. Bridge, William Duke, Terry Engelder, Edward S. Belt, and Mark Sholes who each reviewed a part of the manuscript. This research was partially supported by The Pennsylvania State University Earth System Science Center, Elf Aquitaine SA, and the Donors of the Petroleum Research Fund, administered by the American Chemical Society (Grant No. 16560-AC2).

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Brief Overview of the Appalachian Orogen

The Appalachian orogen sensu stricto, was created as a result of Late Proterozoic (610-630 Ma) rifting of Gondwana (Africa and South America) and Laurentia (proto-North America) (Cook et al., 1983; Cook and Oliver, 1981), along a trend approximately coincident with the axis of the present Appalachians (see Bouguer gravity gradient in Fig. 18 of Slingerland and Beaumont, this volume). Metamorphic and plutonic rocks of the Grenville Province, with radiometric ages of 1.3 to 1 Ga, were stretched to produce a series of grabens filled with thick sequences of Eocambrian sedimentary rocks such as the Chilhowee Gp. (Fig. 2) and volcanic rocks such as the Catoctin Fm. in southeastern Pennsylvania. As the large, near-surface thermal gradients associated with rifting decayed, a passive margin developed upon which a thin Lower Cambrian transgressive clastic sequence (eg. Antietam Fm.) was succeeded by a 4 km thick sequence of Cambro-Ordovician platform carbonates.

The passive margin in the central Appalachians was disrupted in Caradocian time when eastern Laurentia (North America plus Greenland, Scotland, and northern Ireland) collided with an island arc and a set of microcontinents along an outboard-dipping subduction zone (see Fig. 4, panels I-V in Slingerland and Beaumont, this volume) (for a detailed account in New England see Stanley and Ratcliffe, 1985). The resulting overthrusting event, called the Taconian orogeny, depressed the foreland and allowed accumulation of over 1.8 km of sediment in Pennsylvania between Middle Ordovician and Early Silurian time. This is the first of three eastward-derived clastic wedges, the Taconian wedge, represented in central Pennsylvania by the Antes Shale through Tuscarora Formation (Fig. 2) (Lash, 1987; Lash and Drake, 1984; Rodgers, 1970). Modelling by Beaumont et al. (1988) of the type described in Slingerland and Beaumont (this volume), indicates that between 8 and 12 km of overthrust load is necessary to accommodate the maximum 3 km of Taconian detritus preserved in the basin. As explained later, the outboard region of a rifted cratonic margin can accumulate up to about 20 km of overthrust material before a mountain range of any consequence is created. This arises because seaward of the Bouguer gravity gradient marking the continent-ocean crustal transition, thrust sheets replace water and load an attenuated continental crust and oceanic lithosphere. Thus the Taconian overthrusts loading the Cambro-Ordovician slope and rise probably were of modest subaerial topographic

FIGURE 1 General physiography of the central Appalachian orogen and course of the field trip. Numbers represent days from the start of the trip. Latitude and longitude lines are at 2° intervals; 2° longitude equals approximately 200 km. From southeast to northwest the orogen consists of an unconsolidated coastal plain of Cenozoic passive margin sediments; a low relief piedmont underlain by igneous and metamorphic rocks of late Precambrian to Pennsylvanian Age, occasionally interrupted by lowlands of the Mesozoic basins; a crystalline mountain range (labeled Blue Ridge) of Precambrian and earliest Paleozoic age; ridges and valleys of the Alleghanian fold and thrust belt; and a heavily dissected plateau formed on the more nearly flat-lying rocks of the fold and thrust belt (Modified from Raisz, 1954).

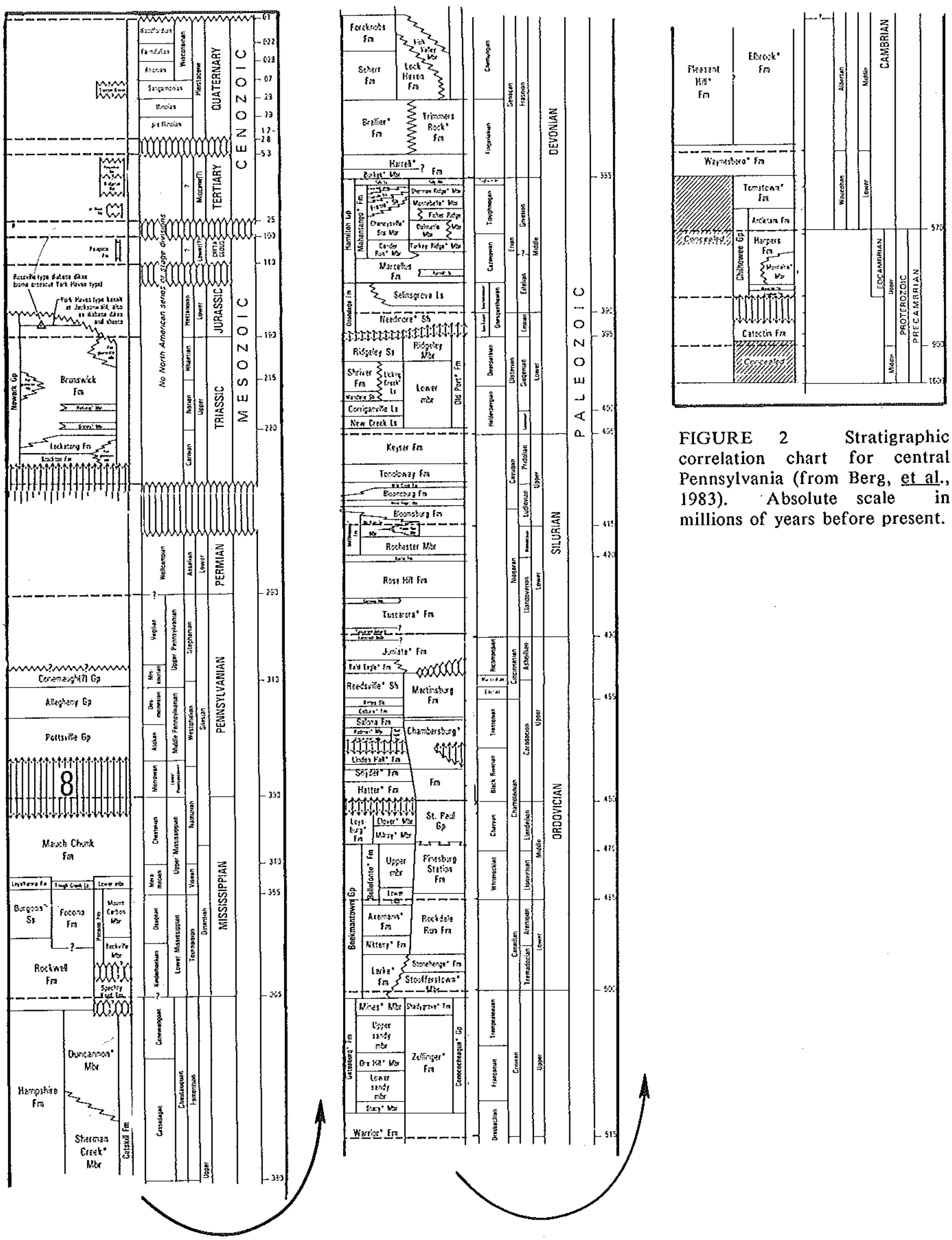


FIGURE 2 Stratigraphic correlation chart for central Pennsylvania (from Berg, et al., 1983). Absolute scale in millions of years before present.

relief. Our interpretation is that the orogen varied along strike among states III-V in Figure 4 of Slingerland and Beaumont (this volume) by the end of the Taconian orogeny.

Following the Taconian orogeny, sedimentation rates declined in the basin. Approximately 900 m of carbonates, salt, fine-grained clastics, and thin, mature shelf sandstones were deposited during Middle Silurian to Early Devonian time (Fig. 2), reflecting relative tectonic quiescence along the orogen. Although plate convergence continued along the eastern Laurentian margin during this interval (Van der Voo, 1988), crustal loading by overthrusting apparently was minor.

Commencing in the Early Devonian in New England and ending in the Early Mississippian in Pennsylvania, convergence between Laurentia and an unspecified plate (Ferrill and Thomas, 1988) produced a metamorphic, plutonic, and loading event called the Acadian orogeny. The resulting foreland basin fill in the central Appalachians is called the Catskill-Pocono clastic wedge (Marcellus through Pocono Formations, Fig. 2), and is the subject of our field trip on days 3 and 5.

Closing of the proto-Atlantic continued during the Mississippian to Permian, culminating in the collision of Gondwana with eastern North America and the third Paleozoic deformation event, the Alleghanian orogeny. Outboard loading rejuvenated the Acadian foreland basin, and it received a minimum of 7.5 km of sediments from the orogenic highlands to the east (Mauch Chunk through Conemaugh Fms. of Fig. 2 seen on field trip days 4, 5, and 6). Subsequently the whole eastern half of the orogen was subjected to folding and thrusting, and, to a lesser extent, metamorphism and plutonism from relative transpression. (see Slingerland and Beaumont, this volume for details).

The Permian and Early Triassic history of the Appalachian orogen is uncertain, because there are no preserved deposits of that age. It is clear however (Fig. 2), that by the Carnian or late Landinian (230-225 Ma) sediments had begun accumulating in basins along reactivated strike-slip and thrust faults (Manspeizer and Cousminer, 1988; Traverse, 1987), recording the initial breakup of Pangea (days 6 and 7). Rupture occurred roughly along the present continental shelf edge (see Manspeizer and Huntoon, this volume, for details) and sea-floor spreading began between late Early to Middle Jurassic (190-175 Ma)(Klitgord and Schouten, 1986, p.364).

A second passive margin developed, of broad platforms having fairly thin sediment cover and basins whose margins probably mark the sites of transform faults active during the initial breakup (Folger et al., 1979). Jurassic sediments of the passive margin tend to be terrigenous lagoonal, fluvial, or deltaic nearshore lithosomes ponded behind widespread carbonate build-ups at the shelf edge. During the Cretaceous and into the Cenozoic, a thick sequence of fluvial, deltaic, and shelf sediments prograded seaward to form a well defined slope and rise. The result is an eastward-thickening wedge of primarily unconsolidated sediments, about 2.4 km thick in the Delmarva area, thickening to 9 km in the Baltimore Canyon Trough (Folger et al., 1979) (day 8.)

TECTONICS AND SEDIMENTATION OF THE UPPER PALEOZOIC FORELAND BASIN IN THE CENTRAL APPALACHIANS

Rudy Slingerland and Christopher Beaumont

INTRODUCTION

Foreland basins are sedimentary basins lying cratonward of major compressional zones. They are formed during continent-continent collisions as a result of outboard crustal loading, or by a combination of loading and subduction of oceanic lithosphere. Those due primarily to outboard loading are especially interesting because the creation of the basin and the source terrain both arise from the same cause --- thickening of the crust by overthrusting. In these basins we expect to see a pattern of evolution that reflects adjustments to the size and rate of application of the overthrust load, variations in time and space of the lithospheric rheology, and feedback between sedimentation in the basin and rates of erosion of the thrust stack.

Our intention here is to illustrate just such an interplay between tectonics and sedimentation in a particularly revealing example, the Appalachian foreland basin of the Appalachian Orogenic Belt. Our method is to first describe some concepts of basin creation using models of flexural response of the lithosphere and then to describe and interpret the character of two orogenies---the Acadian and Alleghanian---and the foreland clastic wedges that resulted from them. The treatment is general; details of the geodynamic modelling can be found in Quinlan and Beaumont (1984), Stockmal et al. (1986), Beaumont et al. (1987), Beaumont et al. (1988), and Jamieson and Beaumont (1988). More in-depth discussions of the field relationships and tectonic evolution can be found in Fisher et al. (1970), Williams and Hatcher (1982), Donaldson and Shumaker (1981), Tankard (1986), Rodgers (1987), and Van der Voo (1988).

FLEXURAL MODELS: CONCEPTS AND BASIC RESULTS

The best starting point for a discussion of the models is a review of the flexural response of the lithosphere to supracrustal loading. The lithosphere's flexural properties determine the form of the foreland basin produced by a given overthrust load as shown diagrammatically in the cross section cartoon of Figure 1. A load emplaced on the surface of an originally flat lithosphere deforms the plate into the profile indicated by curve 1. If the lithosphere's response is effectively elastic, then it will maintain

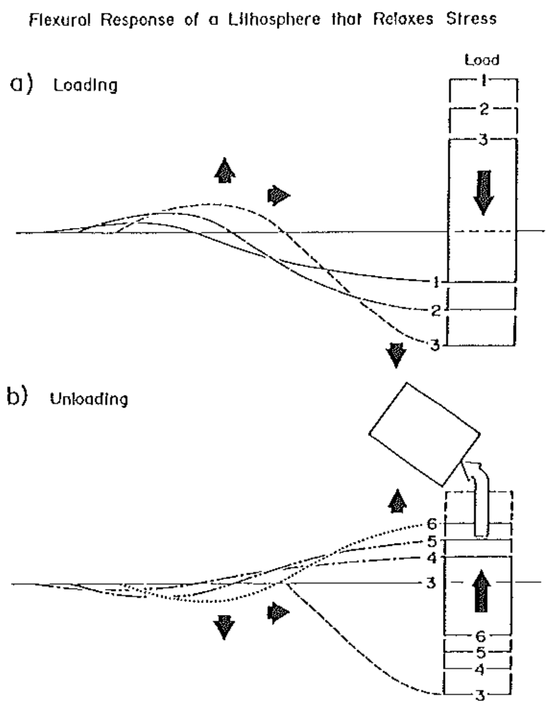


FIGURE 1 Qualitative representation of the loading and unloading response of a model lithosphere that releases stress by some form of thermally controlled creep mechanism. See text for discussion.

this flexural shape while the surface load changes. If, however, the lithosphere can relax the bending stresses set up by the surface load by creep, then its flexural profile will evolve through time to assume the shapes indicated by curves 2 and 3, even though the magnitude of the load remains constant. The timescale over which stress relaxation occurs depends on the mechanism by which stress is relaxed. If viscoelasticity provides a valid model of the relaxation mechanism (e.g. Quinlan and Beaumont, 1984; Beaumont et al., 1988), then it is the viscosity distribution within the lithospheric plate that determines the relaxation timescale. Given that the viscosity of rocks decreases with increasing temperature and that the viscosity of the mantle apparently determines the approximately 10^4-10^5 year relaxation timescale of glacial rebound, relaxation times spanning the range 10^5-10^8 years are expected for the lithosphere. Note that in Figure 1 the peripheral bulge adjacent to the flexurally downwarped region migrates toward the surface load as stress is relaxed and the basin deepens and narrows. This migration may uplift and allow erosion of sediments deposited earlier within the foreland basin. In principle therefore, erosional patterns at the distal edge of the basin can be used to determine whether the lithosphere is able to relax stress and the timescale over which this relaxation occurs. However, there are other mechanisms, such as sea level change, that may also create unconformities, and it is therefore difficult to attribute any particular unconformity unequivocally to lithospheric stress relaxation.

Panel (b) of Figure 1 assumes that part of the orogenic load is removed from a surface made horizontal by erosion of uplifted areas and sedimentary infilling of depressed areas (curve 3). Note that the foreland response to unloading is a mirror image of the response to loading. Uplift first occurs over a broad region (curve 4) and becomes successively concentrated near the unloaded region (curves 5 and 6) if there is stress relaxation. Net reduction of orogenic loading should therefore be recorded in the foreland stratigraphy as an erosional unconformity present over wide areas and having the greatest missing section near the unloaded orogenic region.

Two additional points can be made from these simple concepts. First, each load change applied to the lithosphere evolves through the same sequence of flexural deformation. If the lithospheric response to loading is linear, their superimposed effect in time and space is the sum of the individual effects. Second, an overthrust load that migrates laterally toward the foreland faster than relaxation allows the peripheral bulge to migrate in the opposite direction will create an unconformity as the peripheral bulge is driven across the foreland ahead of the overthrust load (Jacobi, 1981; Quinlan and Beaumont, 1984).

These concepts can be combined to give a first-order explanation of the sequence of events in the development of a multistage foreland basin, like the Appalachian basin (Fig. 2). The first stage shows the development of a basin-wide unconformity as the peripheral bulge migrates ahead of the thrust loads. This phase is followed by subsidence and the formation of a foreland basin. During the quiescent (relaxation) phase, the peripheral bulge is uplifted and migrates toward the thrust load, only to be halted by the next orogeny and loading phase which superimposes the next major sedimentary package of the foreland basin. Thus, as earlier workers recognized in principle, the stratigraphy and sedimentology of the basin fill and the positions of the unconformities in space and time contain important evidence on activity in the adjacent orogen, a point we will return to later.

The question of antecedent conditions and inheritance is important for the style of foreland basins. Although the role of these conditions and details of their effect have yet to be worked out in detail, some aspects have been modelled (Karner and Watts, 1983; Royden and Karner, 1984; Stockmal et al., 1986; Stockmal and Beaumont, 1987). Figure 3 illustrates how Stockmal et al., (1986) incorporated thermal effects and lateral changes in the flexural properties of the lithosphere into models of rifting, passive margin development, plate collision, and overthrusting. Simple elastic plates, the bases of which are defined by a given isotherm were used in the flexural models (Beaumont et al., 1982; Keen and Beaumont, in press).

The significance of these model results (Fig. 4) is that some sense of the geometrical relationship between the overthrusts, their topography, and the flexed crust of the inherited margin is obtained. For example: about 20 km thick loads can overthrust the outboard part of the margin before they need be subaerially exposed (Panel IV); mountain roots beneath orogens of Himalayan proportions may be in excess of 60 km thick (Panel VII); the ultimate preservation of a foreland basin, once the orogen has been eroded to base level, can be attributed to that part of the overthrust load that still remains on or outboard of the antecedent rifted margin (Panel VIII), and; the characteristic Bouguer gravity anomaly common to many compressional orogens (Fig. 5) may be interpreted as the superposition of the anomaly from the inherited rifted margin (the steep gradient above the transitional zone of crustal thinning, Figure 5)

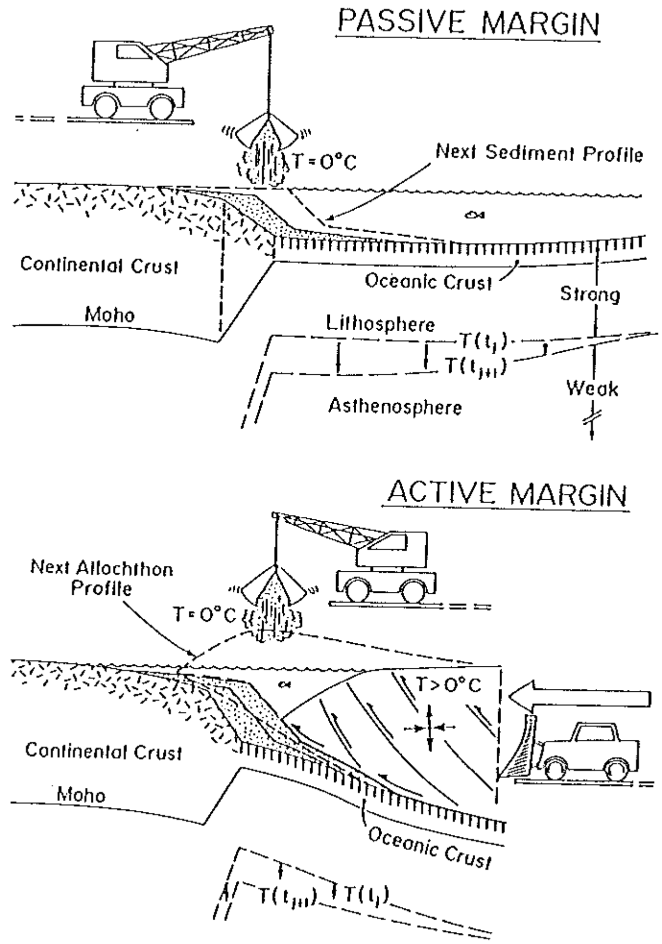


FIGURE 3 Schematic diagram of the quantitative approach to modelling the transition from passive (rifted) to active (convergent) margin used by Stockmal et al., (1986). Stages are constructed in steps following geologically instantaneous rifting; stretched continental crust beneath the margin is located between vertical dashed lines of the upper panel. Steps involving the addition of sediments to a specified bathymetric profile (upper panel) are alternated with thermal time steps during which thermal relocation occurs (shown schematically as a single isotherm T at two times, t_j and t_{j+1}). The flexural response of the lithosphere changes through time because the thermally controlled effective thickness of the lithosphere also changes. The tectonic switch from passive to active margin is modelled by overthrusting loads sequentially onto the passive margin (lower panel). These loads are shoveled into position up to a specified topography (dashed line of lower panel) instead of being pushed in a geologically correct manner. This approximation is reasonable when considering subsidence and sedimentation in the undisturbed part of the foreland.

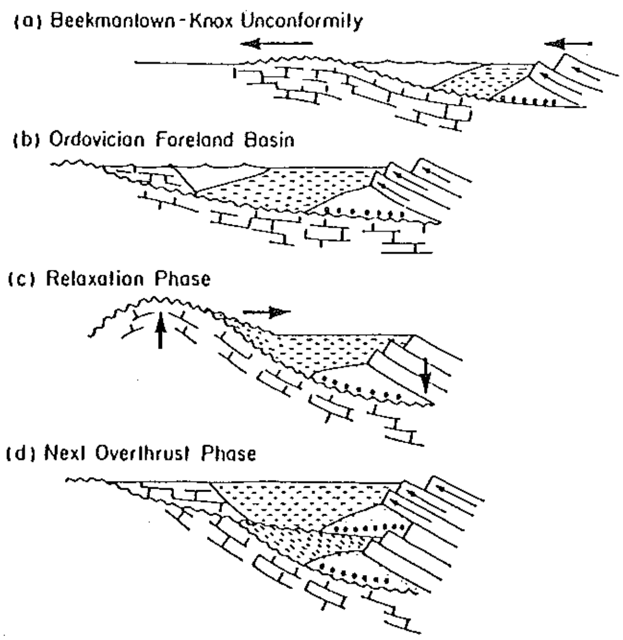
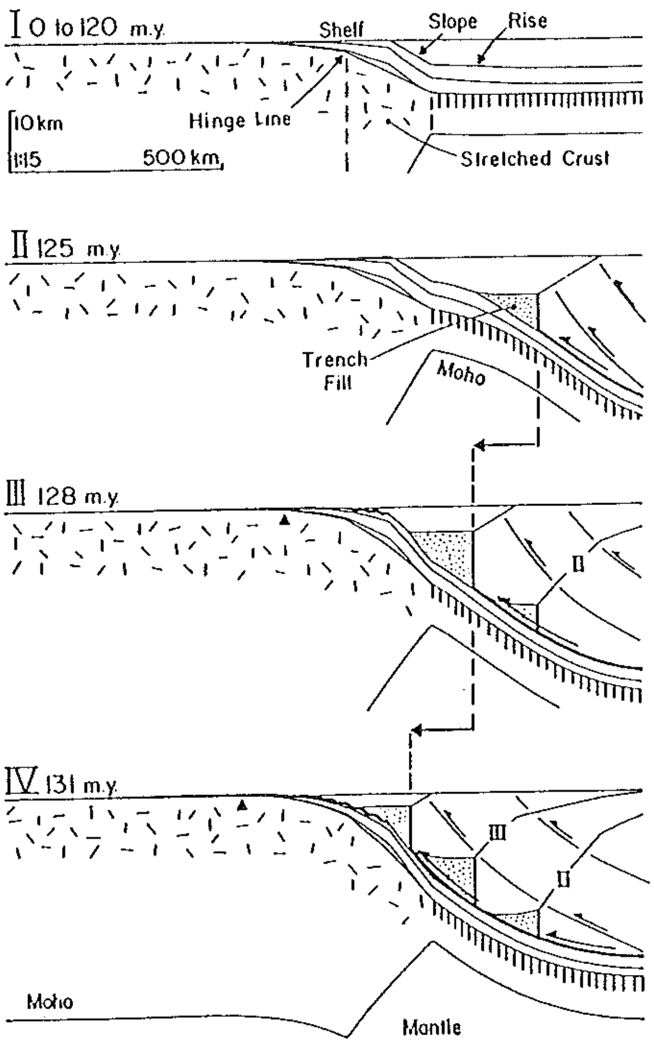


FIGURE 2 Cartoon illustrating the development of a multi-stage foreland basin on a lithosphere that relaxes load-induced stress. The uplift of the peripheral bulge is shown exaggerated by a factor of 10 in (c). Circles represent conglomerates, dots represent sandstone, dashed pattern represents shale, and the brick pattern represents carbonates. Bold arrows show overthrust and peripheral bulge migration. Fine arrows illustrate active overthrusting.

and the longer wavelength flexural component above the foreland. The position of the steep Bouguer gradient may therefore give the approximate location of the inherited rifted margin beneath an orogen (Stockmal and Beaumont, 1987). The change in geometry with increasing amounts of convergence between the overthrusts and the inherited margin can also explain the major change in the associated sedimentary facies from flysch to molasse. Figure 4 (Panels II and IV) also shows that in the early stages of convergence, before the overthrusts have completely mounted the margin, the foreland basin may take the form of a deep asymmetric trough that does not have a characteristic flexural shape.



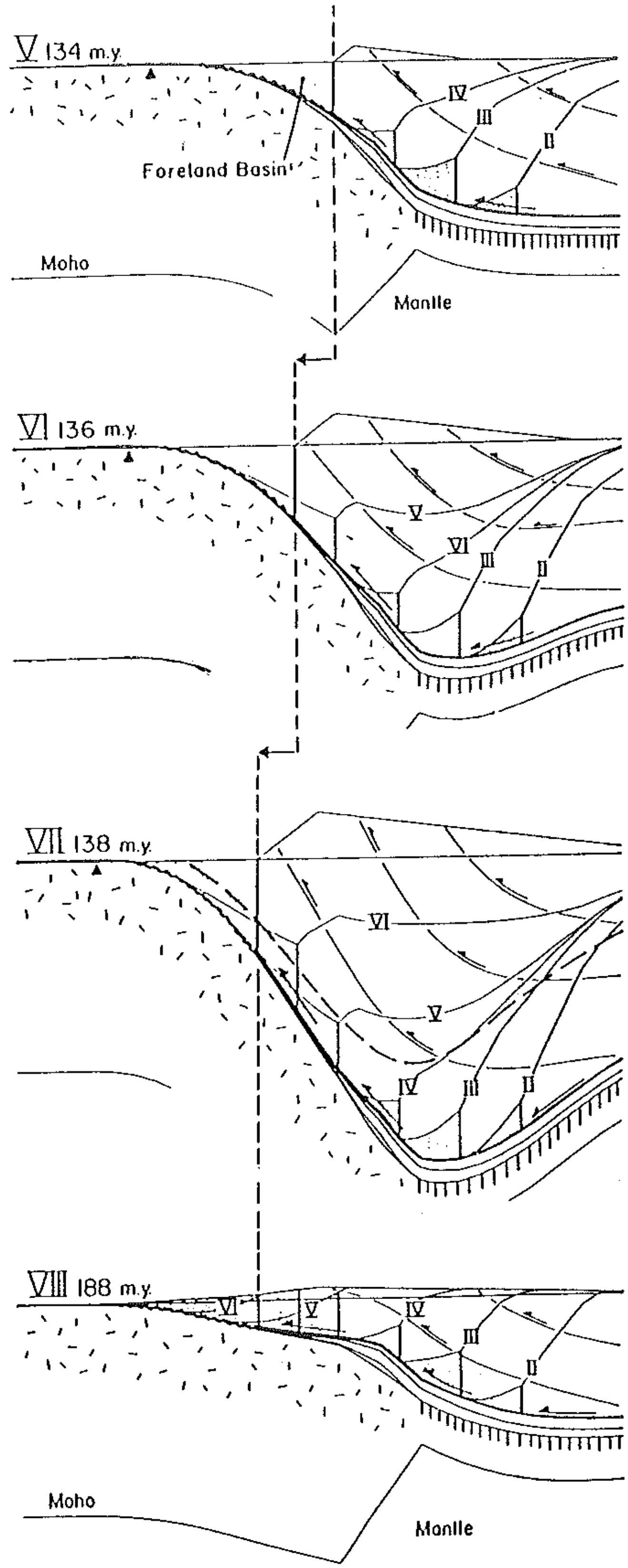


FIGURE 4 Selected steps in the evolution of the model shown in Figure 3 (from Stockmal et al., 1986) in which an orogen is built on a 120 My old rifted margin and then eroded. The vertical exaggeration is 15:1. Random-line pattern is continental and stretched continental crust. Vertical ruled pattern is ocean crust. Bold lines with bold arrows represent the decollement. Bold wavy line represents an unconformity. Stipple pattern marks sedimentary basins. Solid triangles mark the position of the peripheral bulge. Bold dashed line (panel VII) marks the depth within the orogen that is exhumed to the surface during erosion and isostatic rebound between stages VII and VIII.

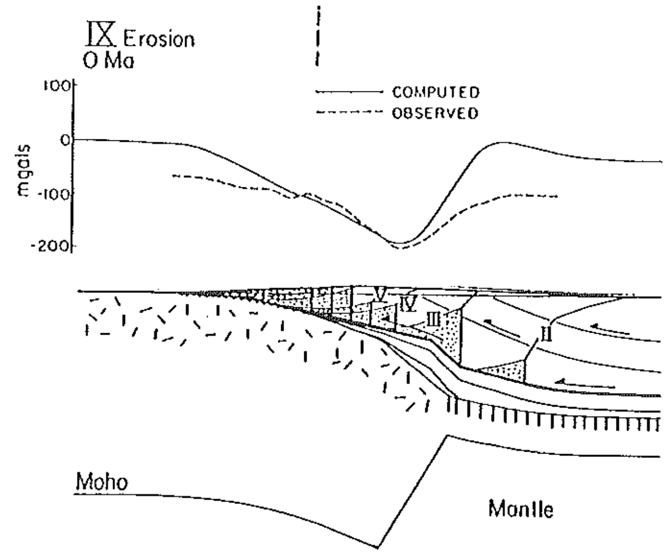


FIGURE 5 Bouguer gravity anomaly (solid line) predicted for a model (lower panel) similar to that of stage VIII in Figure 4. Although larger in amplitude, this anomaly has the same character as that from a typical profile across the western Canada basin and Canadian cordillera (dashed line). The importance of the two parts of the gravity anomaly in relocating the position of the rifted continental margin is explained in the text and in greater detail by Stockmal and Beaumont (1987).

So far in this summary we have concentrated on cross-sectional views, however, the strike variation in loading along an orogen and the interaction of the foreland basin with other sedimentary basins adds considerable variety to the concepts already developed. The cartoon (a) in Figure 6 illustrates the plan view of a foreland basin and peripheral bulge produced by a square load pattern. Figure 6b illustrates the style of coupling between a foreland basin and an intracratonic basin like the Michigan basin. Figure 6c illustrates how superposition of peripheral bulges can generate broad cratonic arches and domes. Our interpretations suggest that all of these features exist within the Eastern Interior region of North America and these figures illustrate the archetypes for the geometrically more complex examples that are presented later. The principles are, however, no more complicated than those illustrated here. The importance of the strike variability in loading is clear when it is remembered that loading in one part of an orogen can cause flexural subsidence and sediment accumulation in the neighboring part of the foreland basin at the same time that it is producing flexural uplift and erosion further along strike.

ANTECEDENT CONDITIONS IN THE APPALACHIANS

The Appalachian foreland basin lies on Grenvillian (1 Ga) North American basement, inboard of the

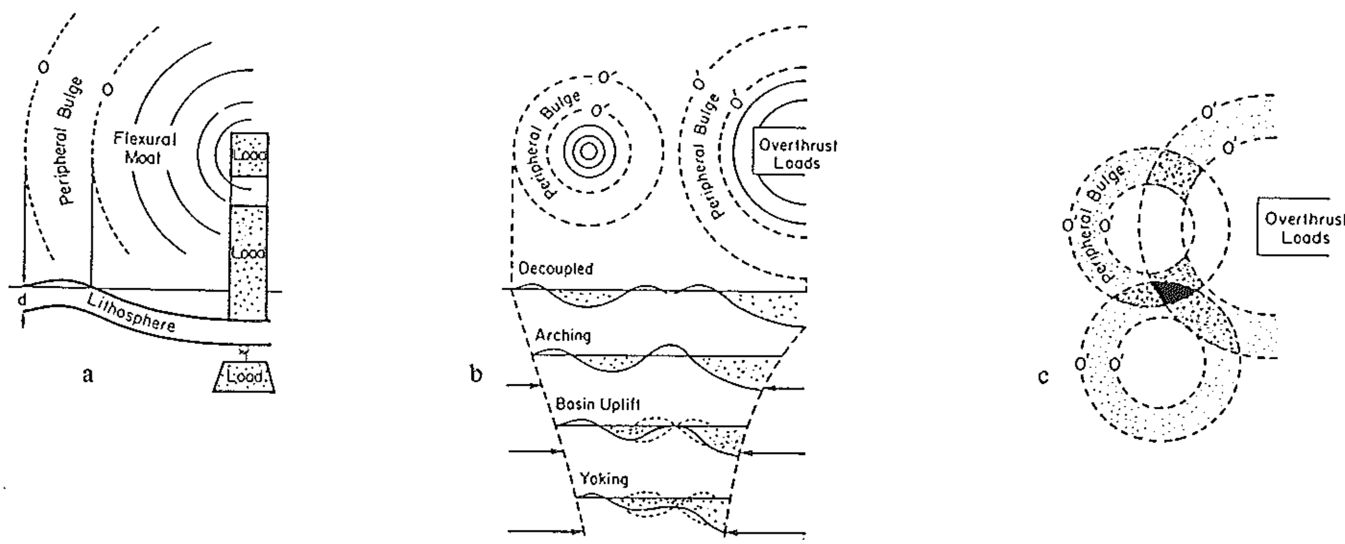


FIGURE 6 a) Cartoon illustrating initial deformation of a uniform lithosphere produced by a square load either deposited on the surface or intruded at depth. The upper panel depicts a plan view (not drawn to scale). b) Flexural interaction between a foreland basin (right) and an intracratonic basin (left). The upper panel shows a plan view of the basins in the decoupled position corresponding to the first cross-section (below). Subsequent cross-sections show the nature of the interaction when basins are closer together. Note that the effect of lithospheric relaxation is not included. This has the tendency to decouple yoked basins with a progression somewhat like that moving upward from the bottom cross-section. c) Flexural interaction between a foreland basin (right) and two intracratonic basins (left). Arched (light stipple) and domed (successively darker stipple) regions are produced by superposition of the peripheral bulges. Note, for the configuration shown, that the deformation yokes the foreland basin with the upper intracratonic basin yet raises an arch between the foreland basin and the lower intracratonic basin because of their greater distance of separation.

crystalline Appalachian Mountains and primarily south of New York State (Fig. 7, Appalachian Basin). It came into existence as early as Middle Ordovician time during the Taconian orogeny, was more-or-less continuously active through the Acadian and Alleghanian orogenies, and ceased receiving sediments sometime in the Permian (see the Introduction for a more complete account of Appalachian orogenesis).

At the start of the Early Devonian and immediately prior to the Acadian orogeny, the Appalachian orogen consisted of an inboard marine foreland basin filled with a maximum 7.3 km of Eocambrian to Late Silurian sediments resting on Grenvillian crystalline basement, and an outboard source terrane consisting of overthrust Taconian island arc, ocean crust, and microcontinent fragments. The source terrane for the most part, rested on attenuated continental crust and ocean lithosphere, and therefore was of low relief.

CHARACTER OF THE OROGENIES AND BASIN FILL

Acadian Orogeny

The Acadian orogeny is characterized by a region of deformation, metamorphism, and plutonism centered in New England and the Maritime Provinces of Canada, but is recognizable as far south as Alabama. In New England the earliest signs of the Acadian orogeny are clastics of late Early Devonian age (Seboomook-Littleton Fms.) overlying carbonates (Rodgers, 1987). By the Middle Devonian, polyphase deformation and metamorphism involved rocks as young as early Middle Devonian. Metamorphism in New England was regional, in places reaching sillimanite grade, and coeval with the emplacement of gneiss domes and intrusion of the voluminous New Hampshire plutonic series. Although deformation continued into the Carboniferous, its style changed to dextral strike slip and normal faulting (Bradley, 1982; Ferrill and Thomas, 1988) with very low grade metamorphism, indicating that the Acadian orogeny, sensu stricto, ended in New England in the Late Devonian (Faill, 1985).

Acadian features can be traced southward from New England where they disappear underneath Long Island Sound and the Coastal Plain deposits. They reappear in central Virginia (Drake, 1980). Surprisingly, the central Appalachians contain no definitive unconformities or intrabasin deformation (Faill, 1985), and no plutonism or widespread metamorphism in the exposed portions. In fact, cooling dates for biotite in the central Piedmont suggest that during the Devonian this terrane mainly experienced westward movement and slow exhumation (Dallemeyer, 1988; Jamieson and Beaumont, 1988). Yet it is here that the largest clastic wedge is preserved, the 3.5 km thick Middle Devonian to Lower Mississippian Catskill-Pocono wedge.

In the southern Appalachians evidence for Acadian orogenesis consists largely of greenshist metamorphism (Hatcher, 1978; Jamieson and Beaumont, 1988), ash fall deposits (Tioga metabentonite) from an early Middle Devonian volcanic center in Virginia (Dennison and Textoris, 1970), and a thick Early to Middle Devonian clastic succession preserved in a thrust slice (Talladega slate belt) (Ferrill and Thomas, 1988).

The explanation of the Acadian orogeny by the over-all plate tectonic movements of the major continents and displaced terranes is controversial. The most recent reconstructions by Van der Voo (1988) (Fig. 8) attribute the Acadian orogeny to the Late Silurian-Early Devonian collision between the Appalachian margin of Laurentia and Gondwana's margin in northwest Africa (with the Avalonian and Armorican accreted terranes caught in between). During Middle and Late Devonian time a newly opened ocean was forming between Laurentia (with its newly accreted Avalonian and Amorican terranes) and Gondwana. This would be consistent with Early and Middle Devonian clastic wedges and associated deformation in New England (Rodgers, 1987) and Alabama (Ferrill and Thomas, 1988), but difficult to reconcile with the Late Devonian clastic wedge of the central Appalachians. Other paleogeographic reconstructions attribute the Acadian orogeny to

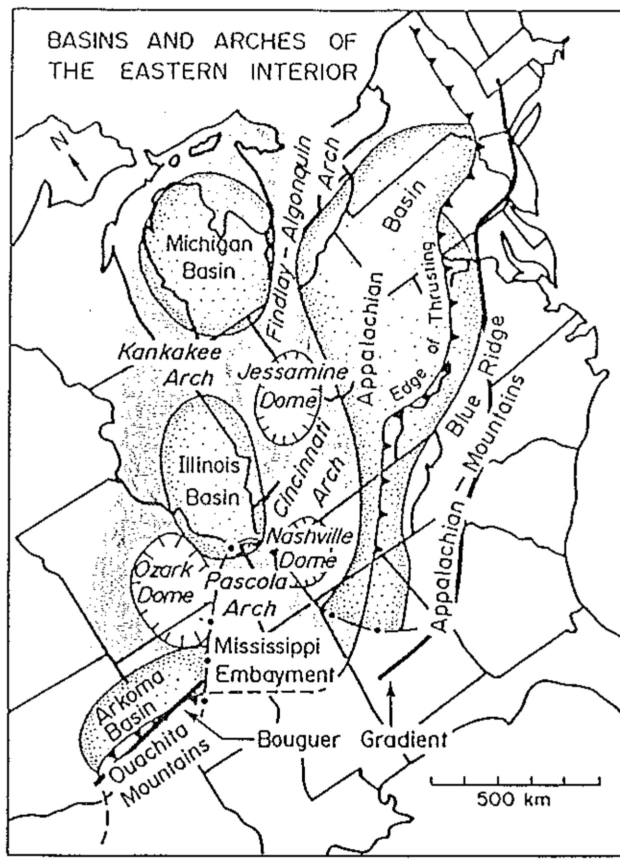


FIGURE 7 General basin configuration of eastern U.S. showing the Appalachian foreland basin (labelled Appalachian Basin), the western extent of Alleghanian thrusting, and the Bouguer gravity gradient thought to reflect the location of the inherited rifted margin (from Beaumont et al., 1988).

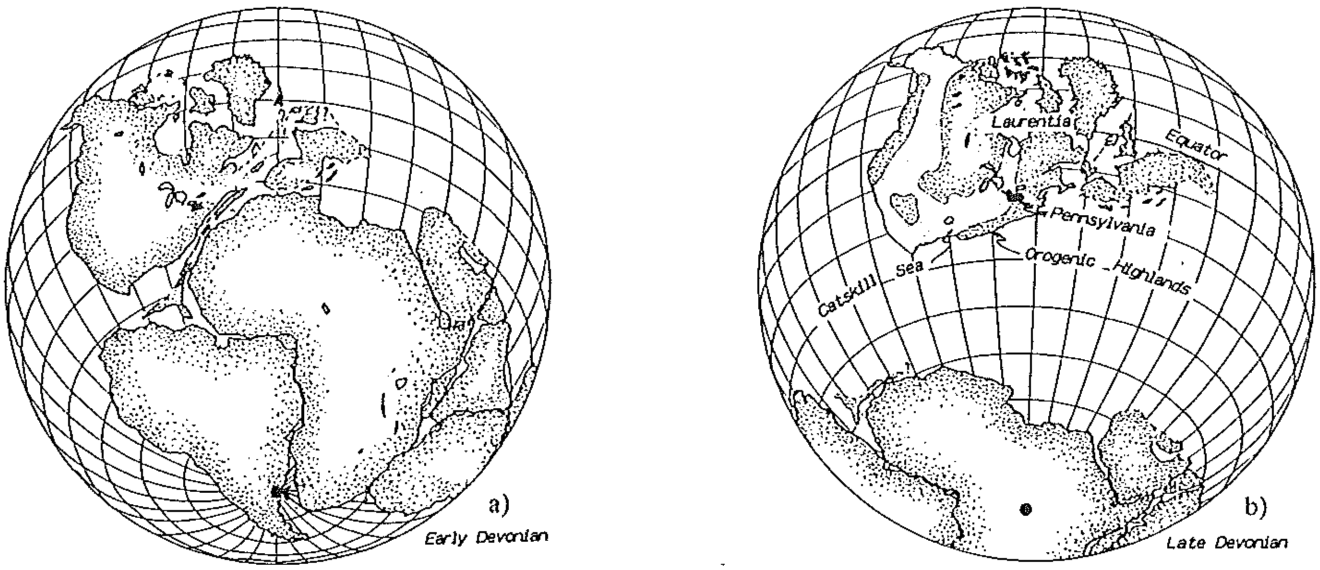


FIGURE 8 Devonian paleogeographic reconstructions using paleomagnetic paleolatitudes and biogeographical and paleoclimatological indicators. The extent of the Catskill epeiric sea is indicated for the Late Devonian (modified from Van der Voo, 1988).

oblique convergence or major transcurrent movement along a sinistral strike-slip zone separating Laurentia and the Avalon terrane during the mid-Paleozoic (Williams and Hatcher, 1982; Ettensohn, 1985) or, more recently, oblique convergence or transcurrent movement along a dextral strike-slip zone separating Laurentia and an unspecified plate during the whole of the Devonian (Ferrill and Thomas, 1988). The southward migration of orogeny, the dextral wrench-fault systems in New England and Alabama, and the discrete location of clastic wedges are attributed to collision of promontories along the irregularly-shaped

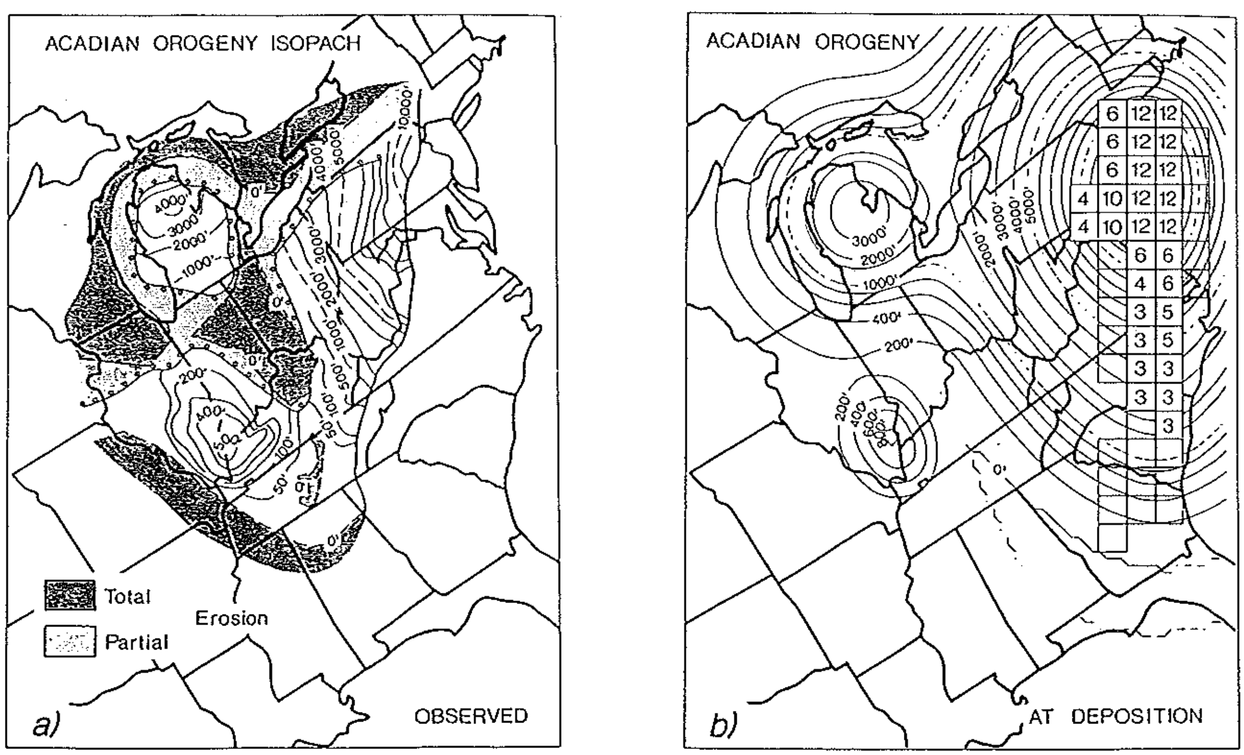


FIGURE 9 Predicted evolution of the total sedimentary isopach associated with the model Acadian orogeny showing its configuration at the end of the Acadian orogeny (panel b) and at present (panel c). Panel b should be compared with panel a, which shows the observed isopach. Shading shows areas of partial and total erosion. All contours are in feet and the numbered grids in panel b are the thicknesses (km) of the overthrust loads necessary to produce the model subsidence in the Appalachian basin.

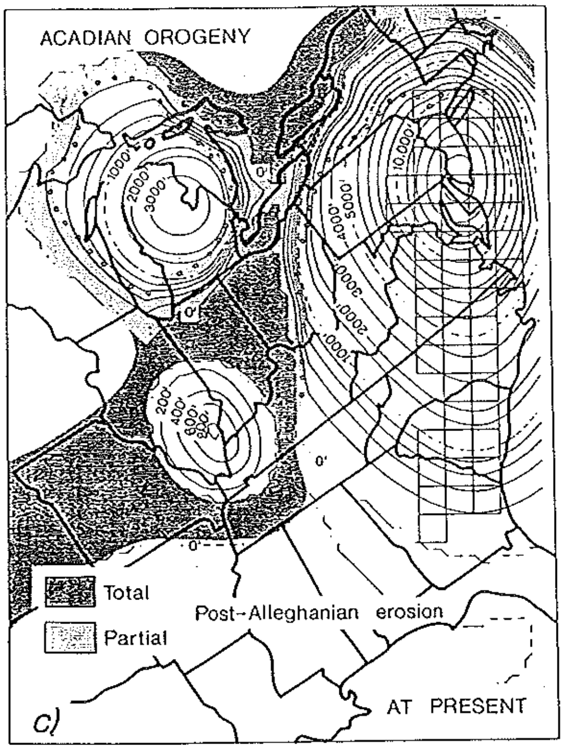


FIGURE 9 (cont.)

plate margins. The modelling presented below suggests compression must have continued into the Late Devonian to produce the basin for the Catskill-Pocono clastic wedge, and therefore we favor this latter view.

Acadian Basin Fill in the Central Appalachians

The most notable manifestation of Acadian orogenesis in the central Appalachians is a pulse of clastic sediments, commonly called the Catskill-Pocono clastic wedge. For the purposes of this discussion, the base of the wedge is placed at the base of the Middle Devonian (in central Pennsylvania, the Needmore Shale) and the top is placed at the base of the Lower Mississippian Loyalhanna Fm. in central Pennsylvania (see Fig. 2 of the Introduction to this volume).

The wedge obtains its thickest expression in eastern Pennsylvania where up to 3500 m (11,400 ft) of predominately alluvial deposits are preserved (Fig. 9a). This accumulation can be explained by the combined effects of the load distribution (Fig. 9b) and the tectonic subsidence of the Michigan and Illinois intracratonic basins by about 830 m and 210 m, respectively. Although the reason for the subsidence of the intracratonic basins is not properly understood, the regional isopach distributions cannot be explained

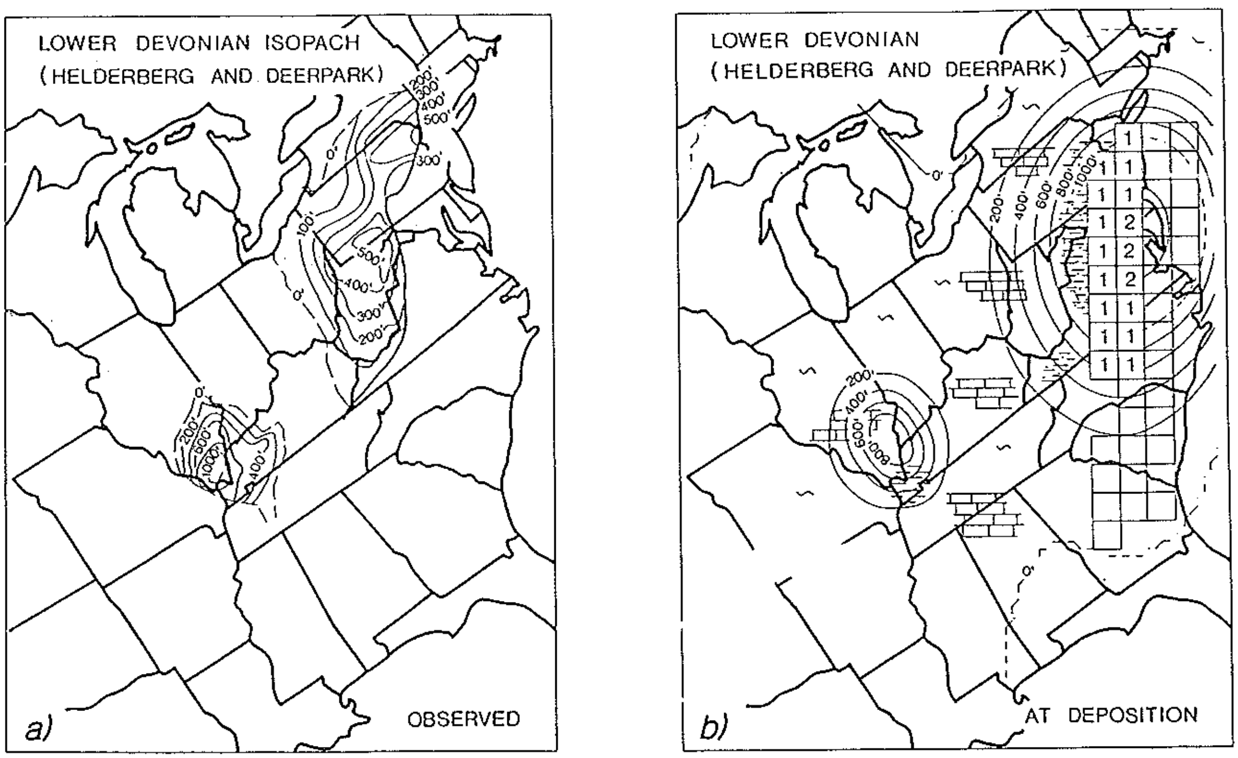
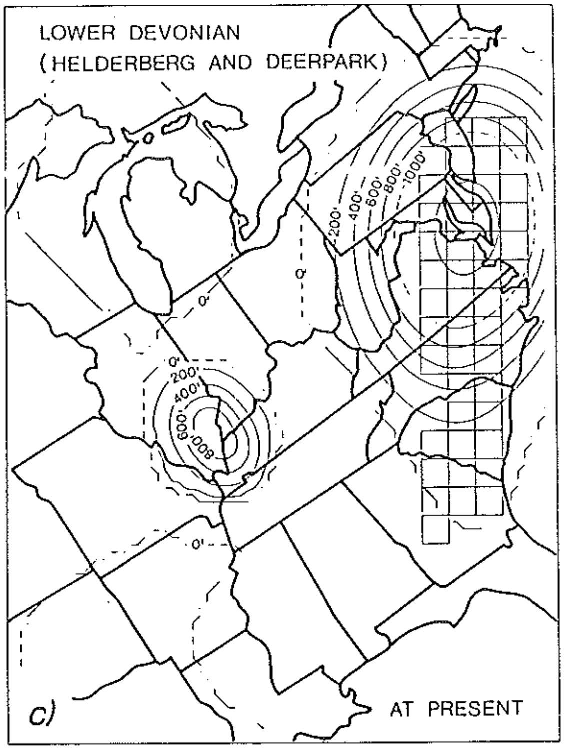


FIGURE 10 Isopach maps of observed (panel a) and predicted (panels b and c) sediment distribution for the Lower Devonian (contours in feet). Panel (d) shows the sediment accumulation rate for Pennsylvania and adjacent regions. The numbered grids in panel (b) are the thicknesses (km) of the overthrust loads necessary to produce the model subsidence in the Appalachian basin. Brick and tilda patterns denote observed and restored chemical sedimentation and marine conditions. Dash and dot patterns denote observed shale and sandstone sediments.



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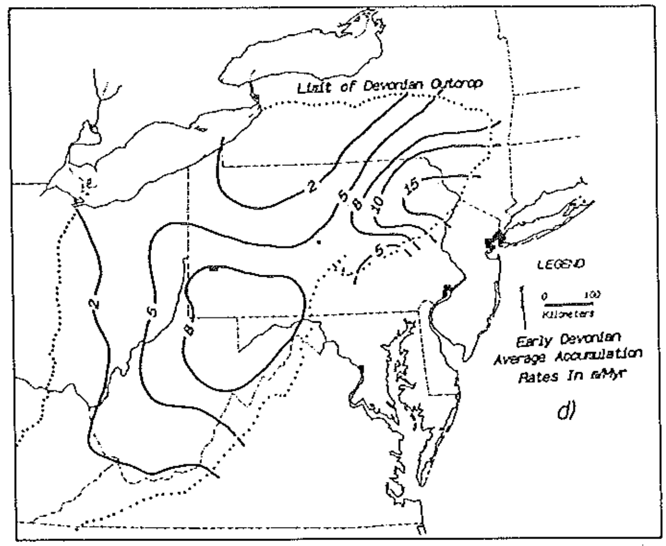
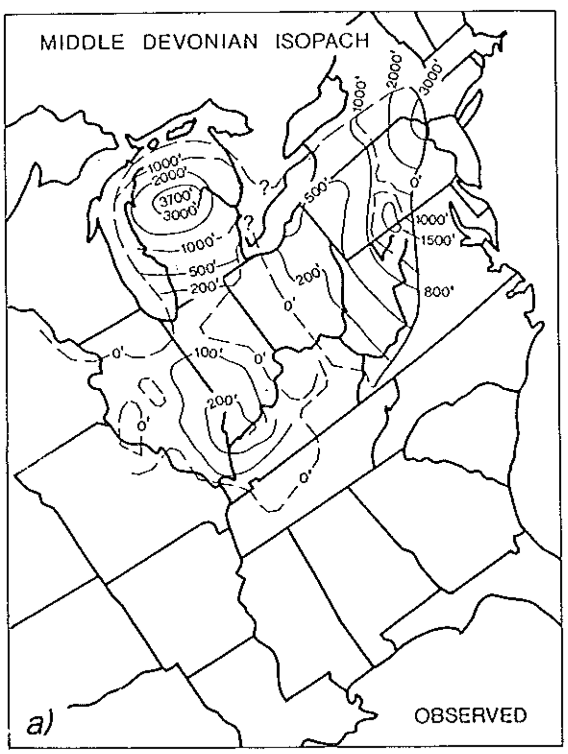


FIGURE 10 (cont.)

without including them. Figure 9b shows the model isopach distribution at the end of deposition, whereas Figure 9c shows the predictions of the present distribution (after uplift and erosion), which agrees quite closely with the observations (Fig. 9a). The model predicts that some erosion occurred before the Alleghanian orogeny but that the majority occurred



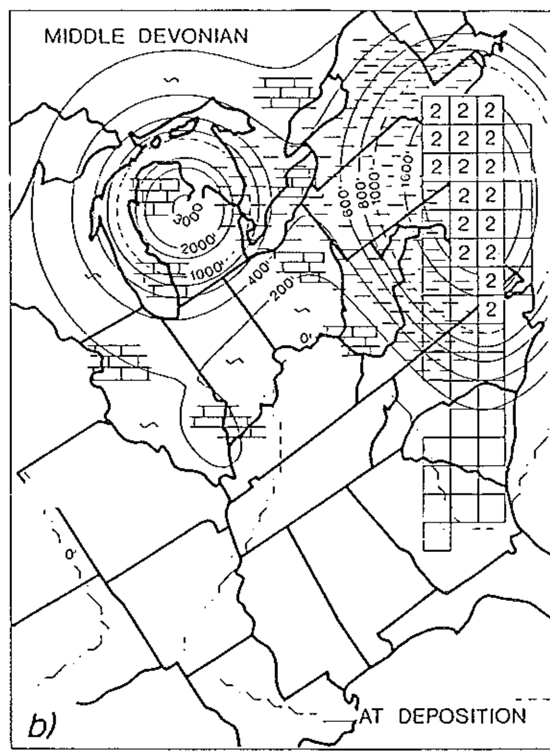


FIGURE 11 Isopach maps of observed (panel a) and predicted (panels b and c) sediment distribution for the Middle Devonian (coutours in feet). Panel (d) shows the sediment accumulation rate for Pennsylvania and adjacent regions. The numbered grids in panel (b) are the thicknesses (km) of the overthrust loads necessary to produce the model subsidence in the Appalachian basin. Brick and tilda patterns denote observed and restored chemical sedimentation and marine conditions. Dash and dot patterns denote observed shale and sandstone sediments.

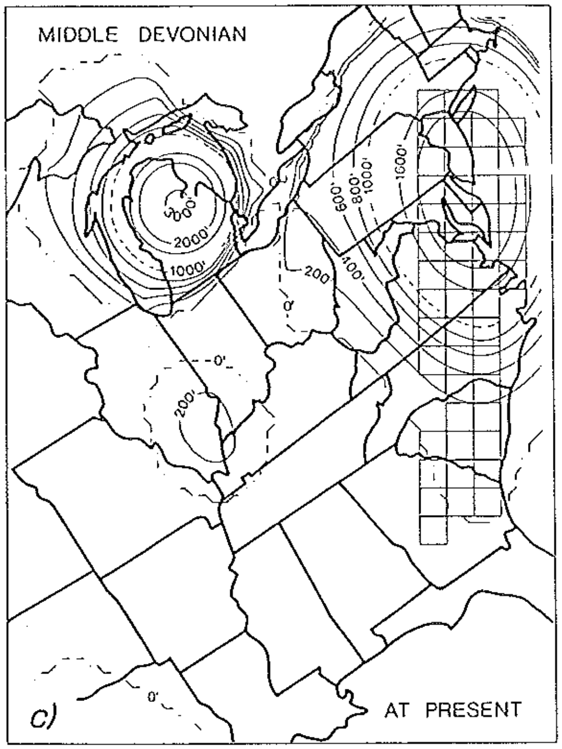


FIGURE 11 (cont.)

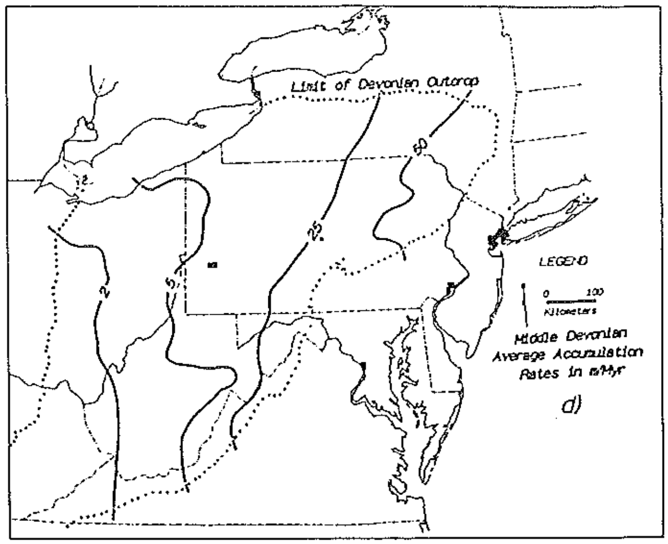


FIGURE 11 (cont.)

between the Permian and present (Beaumont et al., 1988). Erosion predominates on the arches and domes (Figs. 7 and 6c), reflecting the process of stress relaxation and uplift of those regions due to the superposition of the peripheral bulges. It is important to note that a purely elastic model of the lithosphere cannot correctly reproduce this pattern of erosion.

Sediment accumulation rates (as indicated by the thickness remaining per unit time) varied dramatically over the interval of the Acadian orogeny (Figs. 10d, 11d, and 12d). In the Early Devonian two accumulation centers existed in the central Appalachians, with the northern receiving carbonate sediment at a rate of 15 m/Myr (Fig. 10d). This pattern (Oliver et al., 1967), which was originally established in the Upper Silurian (Colton, 1970), cannot be explained by the flexural model if sedimentation completely filled the foreland basin. It is in cases like this that geodynamic models can point to problems requiring a solution. That the flexural model is so successful for other intervals, when there was a large clastic influx into the basin, lends credence to our faith in the model for this Early Devonian interval, yet two closely separated depocenters (Fig. 10a) should be flexurally connected along strike (Figs. 10b and 10c). The obvious explanation is that in central Pennsylvania the basin remained underfilled with paleobathymetry as large as 50 m at the end of the interval. This can be substantiated in part because at the end of Deerpark Age a sizable sea level drop occurred producing the Wallbridge Discontinuity (Dennison, 1985). This discontinuity is absent in western Maryland, northern West Virginia, and southwestern and eastern Pennsylvania, suggesting that the sea was deep there at the end of Deerpark time.

By the Middle Devonian, accumulation rates had increased to a maximum 50 m/Myr along the southwestern border of the basin in response to Acadian overthrusting, whereas on the west side of the basin in Ohio rates remained constant (Fig. 11d). The model results (Figs. 11b and 11c) that best match the observed thicknesses (Fig. 11a) indicate that there was no great increase in the rate of loading between the Early and Middle Devonian, although the load distribution may have migrated somewhat to the north. The increase in sedimentation is most likely a response to the initiation of Acadian mountains outboard of the central Appalachians which provided a good source of detrital sediments to this part of the basin. That these sediments most probably filled the basin completely is reflected in the flexural shape of the preserved isopach (Fig. 11c).

In the Late Devonian, accumulation rates increased by almost fourfold in the east and an order of magnitude in the west (Fig. 12d). As expanded upon below, these rates overwhelmed subsidence rates and a subaerial alluvial plain was created that prograded westward. The preserved sediment distribution is one of the most convincing pieces of evidence in favor of a flexural model of the Appalachian foreland basin in which loads up to 10 km thick were overthrust in the vicinity of what is now southern New York, New Jersey, and Maryland. That the preserved isopach is asymmetric along strike with respect to this depocentre suggests that there was also loading further south within the orogen (compare Figs. 12a and 12c).

There is little doubt that the clastic sedimentation covered the whole of the Eastern Interior as far south as Tennessee (Fig. 12b). Preserved clastic sediments from this time within the intracratonic basins is further evidence that the arches and domes were flexurally depressed. If this interpretation is correct, the initial, or loading, flexural wavelength of

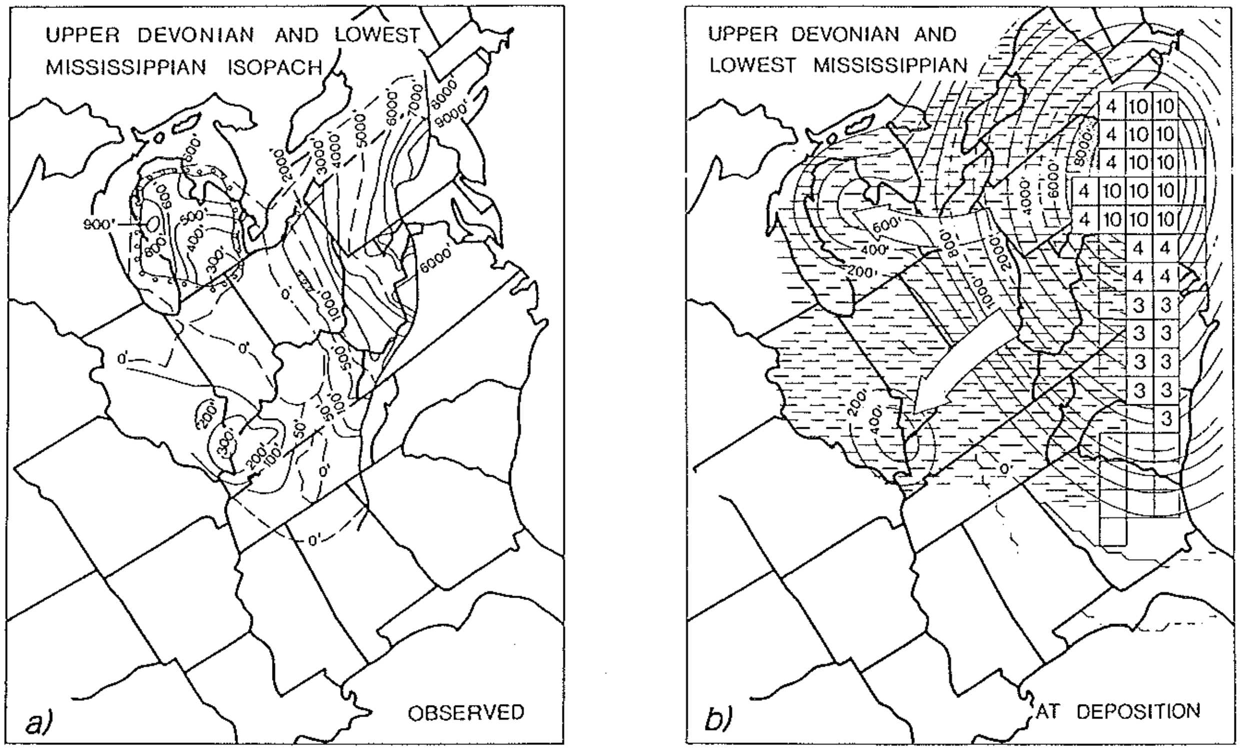


FIGURE 12 Isopach maps of observed (panel a) and predicted (panels b and c) sediment distribution for the Upper Devonian (coutours in feet). Panel (d) shows the sediment accumulation rate for Pennsylvania and adjacent regions. The numbered grids in panel (b) are the thicknesses (km) of the overthrust loads necessary to produce the model subsidence in the Appalachian basin. Dash and dot patterns denote observed shale and sandstone sediments.

the lithosphere under the Eastern Interior must have been sufficiently large to couple the intracratonic basins into the Appalachian foreland basin as shown in Figure 12b.

Middle Devonian Depositional History. Immediately following deposition of the Tioga metabentonite, organic rich, black and grey shales (Marcellus Shale) spread westward through the epeiric sea into Ohio at the same time that 300 m of siltstones and sandstones

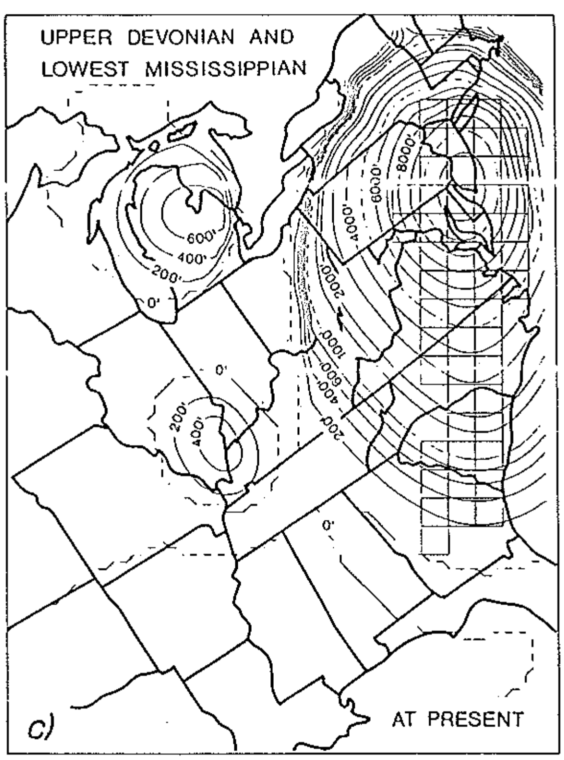


FIGURE 12 (cont.)

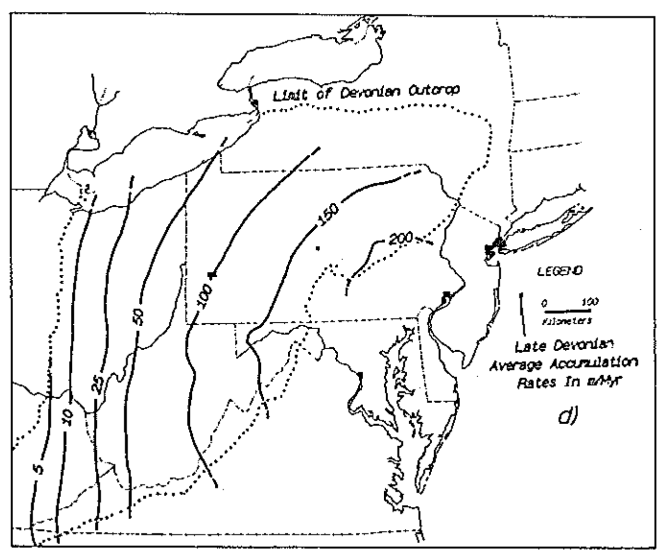


FIGURE 12 (cont.)

of the Mahantango Formation were deposited in eastern Pennsylvania (Fig. 13). These units are interpreted by Kaiser (1972) to be the result of a delta complex that prograded northwestward from Maryland into eastern Pennsylvania during upper Middle Devonian time. The shoreline at the time of maximum progradation in the Givetian Stage is given in Figure 14 as number 3. This first phase of shoreline progradation, was terminated by a eustatic (?) sea level rise, the Taghanic onlap, which transgressed the shoreline to position 4 (Fig. 14) and deposited the Tully Limestone Mbr. (Fig. 13), a deep water micrite interbedded with black shale.

Late Devonian Depositional History. By the Late Devonian, the Appalachian orogen was in the subtropics (Fig. 8) where southeasterly trade winds created a tropical climate with alternating wet and dry seasons restricting plants to the fringes of rivers, lakes, and the shoreline, and promoting redbed formation. A large epeiric sea, the Catskill Sea of Woodrow and Sevon (1985), covered the eastern interior. Increasingly higher rates of clastic sediment flux to the basin quickly prograded the shoreline of this sea back to the west (Fig. 14, position 5), producing the famous Catskill regressive sequence (Figs. 13, 14, and 15). At most stratigraphic sections in Pennsylvania (Fig. 15), the sequence starts with deposits of distal-basin dark shales, passes upwards into grey turbidites (eg., Brallier Fm. of Day 3, Site 2, Outcrop 1) of the shelf slope rise or clinoform (Woodrow, 1985), that in turn give way to upper slope and storm-dominated shelf facies (eg., Loch Haven and "Chemung" Fms. of Day 3, Site 2, Outcrop 2)(Slingerland and Loule, in press). Lying above the shelf facies are marginal marine deposits (eg., Irish Valley Mbr., Catskill Fm. of Day 3, Site 2, Outcrop 3) that, in Pennsylvania (Fig. 15), consist of two tide-dominated deltaic depocenters separated by the extensive tidal flat facies of a muddy shoreline (Rahmanian, 1979; Williams, 1985; Warne, 1986; Slingerland and Loule, in press). Petrographic differences among the depocenters to the south (Kirchgessner, 1973) indicate variations in the source

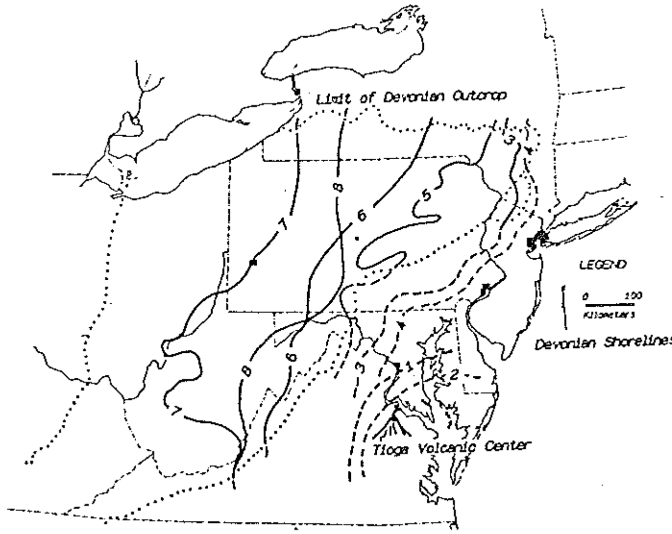


FIGURE 14 Devonian shorelines in the central Atlantic states. Dotted line encompasses the preserved Devonian strata; dashed lines are inferred from clastic wedges preserved further into the basin. Variation in age along any one shoreline can be millions of years: 1 = early Onesquethawan (377 Myr); 2 = late Onesquethawan; 3 = Tioughniogan; 4 = Taghanican; 5 = Finger Lakesian; 6 = Cohocton; 7 = early Bradfordian; 8 = late Bradfordian (346 Myr) (modified from Dennison, 1985).

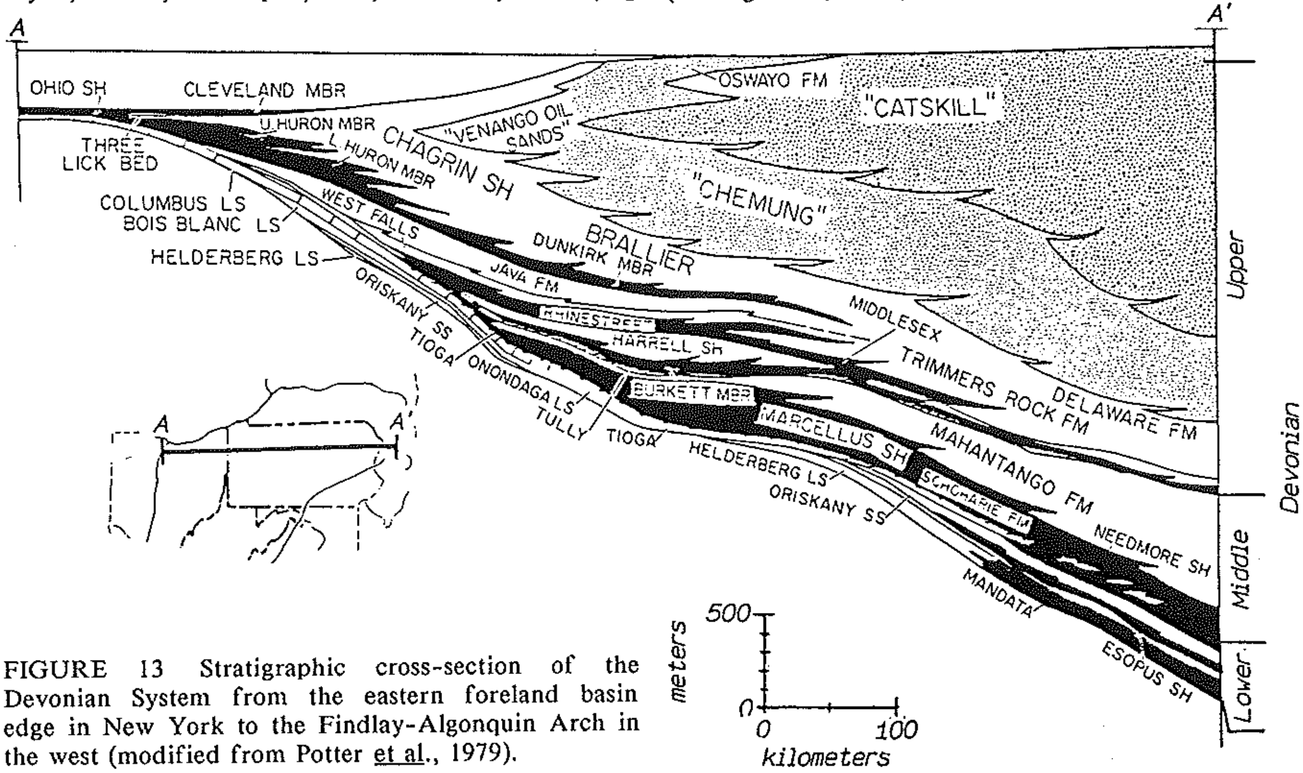


FIGURE 13 Stratigraphic cross-section of the Devonian System from the eastern foreland basin edge in New York to the Findlay-Algonquin Arch in the west (modified from Potter et al., 1979).

terrain from a greenschist facies provenance to the south to a higher grade or more igneous provenance to the north. The shoreline deposits are overlain by fluvial deposits of a vast alluvial plain that extended east to the Acadian Highlands. Low on the plain the rivers meandered (Bridge and Gordon, 1985) whereas higher on the plain the streams were low sinuosity meandering or braided (eg., Day 3, Site 2, Outcrop 4 and Day 5, Site 1 of Duncannon Mbr., Catskill Fm.)(Sevon, 1985; but see Bridge and Nickelsen, 1986 for an alternative view). The locations of the major streams across Pennsylvania were relatively fixed (Williams, 1985; Slingerland and Loule, in press; Sevon, 1985), probably by topography in the source region or basement tectonics.

During early and middle Famennian time the Catskill alluvial plain prograded an additional 167 km (100 miles) across the central Atlantic states (Fig. 14, shoreline 6), reaching its maximum westward position (shoreline 7) in late middle Famennian time. The world's first commercial oil well was drilled by Col. Edwin Drake in offshore shelf sandstones of this age (Fig. 13, "Venango Oil Sands"). Subsequently, a widespread and rather abrupt marine transgression overran the alluvial plain for 80-160 km (50-100 miles) (shoreline 8), depositing the Riceville Shale and Oswayo Fm. The time-equivalent alluvial rocks (eg. lower two-thirds of the Pocono [Rockwell] Fm. in outcrop 4 of Stop 3) have lost their red color but otherwise show little evidence of this change in base level. In fact, most interpretations of depositional environments in this interval (Rahmanian, 1979; Berg, 1981; Williams, 1985) imply that westward progradation of the steeper, upper alluvial plain continued uninterrupted, suggesting the transgression was primarily eustatic in origin. This is substantiated by its effects as far away as the Canadian Rockies (Dennison, 1985).

Early Mississippian Depositional History. The last phase of Acadian deposition is represented by rocks of the Pocono Fm. of Pelletier (1958) (Fig. 16). The braided alluvial plain (eg., Burgoon Ss. of Day 3, Site 2, Outcrop 5) depicted in Figure 16 in Kinderhookian time, prograded westward again, displacing shallow marine facies (eg., Shenango Fm. and Berea Ss.). Average accumulation rates across southern Pennsylvania were more similar to the Middle than Upper Devonian however, being only 49 m/Myr in the east and 15 in the west (Pelletier, 1958). This decrease in accumulation rate defines the end of the Acadian orogeny and its effects.

Alleghanian Orogeny

The Permo-Carboniferous Alleghanian orogeny is characterized by a molasse sequence in the foreland, thrusting and folding of the whole orogen but especially the foreland in the southern and central Appalachians, and regional metamorphism and plutonism along the entire eastern margin of the Appalachians. Its earliest effect in the central Atlantic region was a warping of the foreland basin in Meramecian time with uplift and erosion inboard, renewed sediment influx outboard, and development of a marine embayment in between. Deposition of shales in Arkansas also signals the onset of thrusting in the Ouachita part of the orogen. As discussed below, deformation outboard of the fold and thrust belt must have taken place continuously into the Permian to provide the necessary loads for the foreland. Numerous S-type granitic plutons were emplaced in the eastern Piedmont from mid-Carboniferous to Permian time (Hatcher, 1987; Jamieson and Beaumont, 1988) and most seem to be post-tectonic (Rodgers,

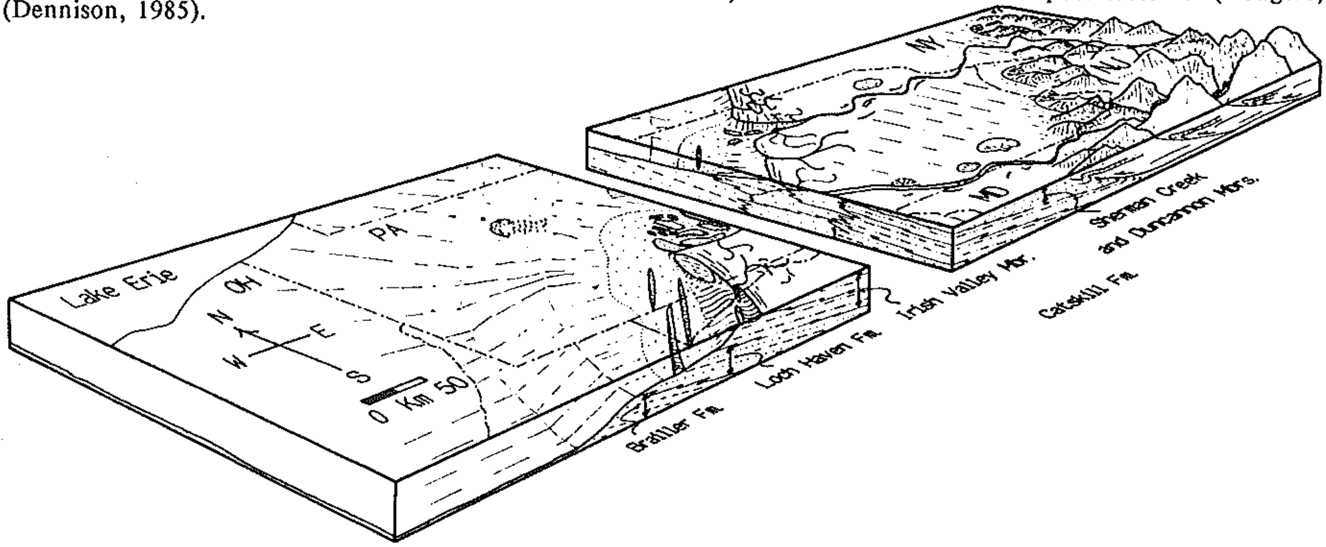


FIGURE 15 One-point perspective sketch of Devonian shoreline 6 (Fig. 14) showing the paleogeography, sedimentary paleoenvironments, and deposits across Pennsylvania. Two major meandering river systems are inferred to have drained the Acadian Highlands (interpreted as thrust sheets), and debouched into the Catskill Sea through trumpet-shaped, tidally influenced estuaries. Offshore, wind-driven geostrophic flows transported sediment plumes to the southwest, forming shelf sand sheets with ridges on an otherwise muddy shelf. Dilute silty turbidity currents carried sediments onto the basin floor.

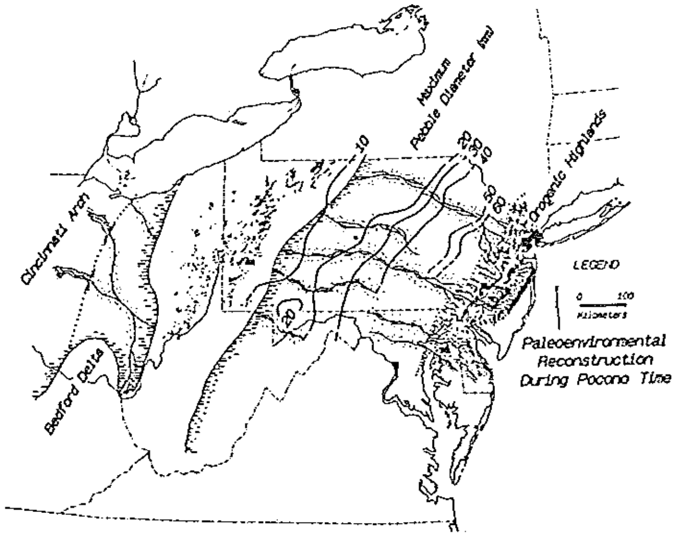


FIGURE 16. Paleogeography of middle Atlantic states during the Kinderhookian (Early Mississippian). Acadian Highlands to the east fed braided streams draining westward across Pennsylvania, producing the Pocono (Rockwell and Burgoon) Fm. The Cincinnati (Findlay-Algonquin) Arch, uplifted by lithospheric relaxation, fed a delta system which prograded south-southeast. Hydrocarbon reservoirs (shown in black) were formed in the narrow seaway (modified from Pelletier, 1958, and Donaldson and Shumaker, 1981).

1987). Folding in the preserved part of the fold and thrust belt of Pennsylvania did not occur prior to early Permian however, as evidenced by the fact that early Permian strata in western Pennsylvania are concordantly folded. These facts suggest that a wave of deformation and heating moved cratonward over the interval from mid-Mississippian to late Early Permian. The orogeny in the vicinity of Pennsylvania was completed by the end of Early Permian time because a remnant magnetization on fold limbs in central Pennsylvania is independent of bedding attitude and follows the cratonic polar wander path after that time (Van der Voo, 1988). Elsewhere in the Appalachians the Alleghanian orogeny probably ended by the end of the Permian because the plutons are no younger than 260 Myr (Jamieson and Beaumont, 1988).

Alleghanian deformation in the foreland is characterized by thin-skinned thrusting towards the continent (Rodgers, 1983, 1987; Hatcher, 1981; Mitra, 1986)(Fig. 17). All along the central and southern portions of the orogen, portions of the western Piedmont and Blue Ridge crystalline rocks moved westward, acting as a plunger to deform the sedimentary rocks of the foreland (Fig. 18). There seems little doubt that the ultimate cause was the final collision of Laurentia and Gondwana. The result in the foreland is the classic fold and thrust belt we see today, exhumed by up to 12,000 m (40,000 ft) of Mesozoic and Cenozoic erosion (Fig. 19).

Alleghanian Basin Fill

We consider the Alleghanian basin fill to commence with deposition of the Loyalhanna Fm. or Mauch Chunk equivalent and end sometime in the Permian. Because the Permian section is partially eroded, the total thickness of Alleghanian molasse is unknown. In eastern Pennsylvania at least 5-7 km of additional overburden are required to account for the level of organic metamorphism of the anthracite (Levine, 1983; 1986), sediment bulk densities and porosities (Paxton, 1983), fluid inclusion paleopressures (Orkan and Voight, 1985), and fission track thermochronometry (Beaumont, et al., 1987). The bulk of this may have been tectonically emplaced however. In western Pennsylvania where overthrusts are not a factor, the moisture content of the coals indicates an additional 2400 m (8000 ft) of Permian strata (Beaumont, et al., 1987). An estimated 2500 m maximum of preserved Alleghanian fill in the Southern Anthracite Field of Pennsylvania plus 2400 m of Early Permian strata (now eroded), yields an accumulation rate of 196 m/Myr, similar to the highest rates of the Late Devonian.

The model reconstruction of total Alleghanian loading and molasse deposition in the Appalachian and Arkoma basins (Fig. 20) is in agreement with data from preserved sediments and reconstruction of eroded section based on coal moisture content (Beaumont et

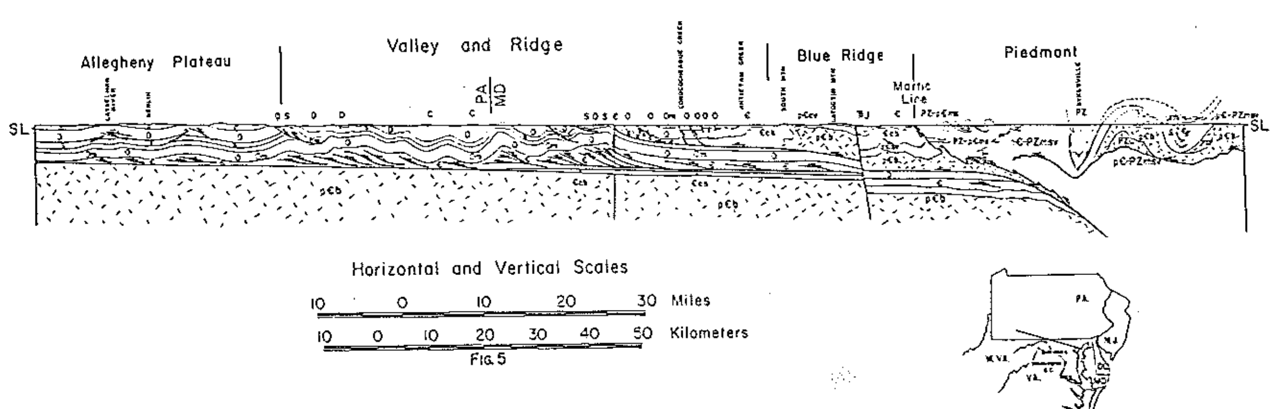


FIGURE 17 Schematic cross-section based on surface geology and seismic data (where available). The northwestward-directed folds and thrusts of the foreland (Valley and Ridge and Allegheny Plateau regions) are due to Alleghanian orogenesis (modified from Hatcher, 1981).

al., 1987). This reconstruction shows the scale of the foreland basin at the end of the orogeny and the magnitude of thrust loads which by this time were concentrated in the southern Appalachians and the region within and to the south of the Ouachita mountains. The model assumes minimum tectonic subsidence of the Michigan and Illinois basins of 160 m and 1115 m, respectively. Cumulative thicknesses of overthrust loads required by the model to reproduce the total foreland basin subsidence are shown in Figure 18.

Late Mississippian Depositional History. The foreland basin in the central Atlantic region responded to Alleghanian orogenesis in early Late Mississippian (Meramecian) time by subsidence along a northeast-southwest axis across Pennsylvania, creating a trough in which transgressive marine carbonates (eg., Greenbrier Ls. of West Virginia and Loyalhanna Fm. of Pennsylvania) were deposited (Fig. 21). Simultaneously, the region to the northwest experienced uplift and became a source for the

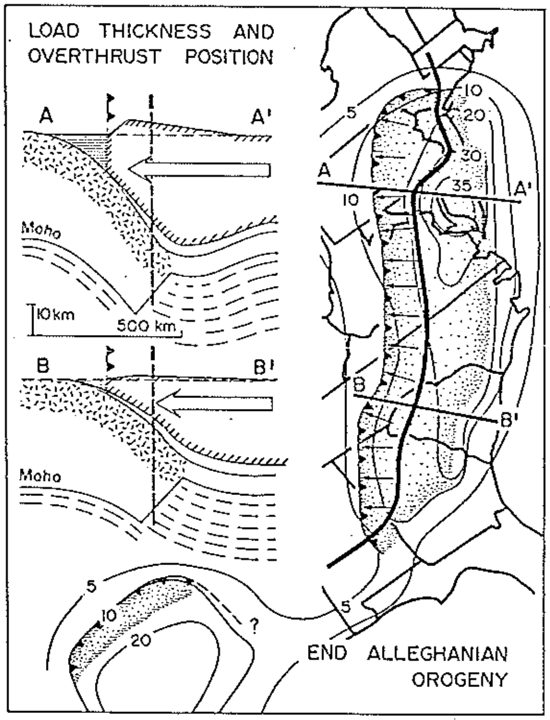


FIGURE 18 Interpretation of the cumulative model loads (fine line contours in km) in terms of overthrusting and thickening in the Appalachian and Ouachita orogens by the end of the Alleghanian orogeny. The bold lines show the location of the Bouguer gravity gradient in relation to the overthrust loads. Note that the thickest loads are to the east and south of this gradient. Barbed lines show the edge of basement involved thrust sheets and stippled areas are the inferred mountainous regions. Fine arrows illustrate the inferred advance of the thrust front during the Alleghanian orogeny in the Appalachian part of the orogen. The cross sections are based on results from Stockmal et al. (1986) and Stockmal and Beaumont (1987) and should be compared with Figures 4 and 5.

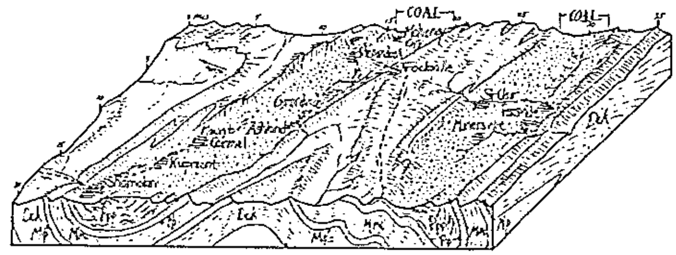


FIGURE 19 Schematic relationships among differentially resistant foreland strata, Alleghanian folds, present geomorphology, and coal in the Southern Anthracite Field. Days 5 and 6 will be spent in Ashland and Pottsville.

trough. To the southeast, the orogenic source created earlier, continued to supply sediments, and a delta complex (Mauch Chunk Fm., Day 6, Site 1) built northwestward. By the end of the Chesterian Stage the Loyalhanna embayment was completely filled. Continued uplift to the northwest allowed streams to erode the Mauch Chunk margin soon after deposition and transport the sediment along the basin axis to the Kentucky region (Fig. 22). On the southeastern margin of the basin where Mauch Chunk sedimentation was continuous, approximately 1140 m (3800 ft) of alluvial redbeds are preserved (Arkle, 1974), yielding a maximum accumulation rate over the interval of 38 m/Myr.

The Mississippian sediment distribution can be explained approximately by the two model timesteps

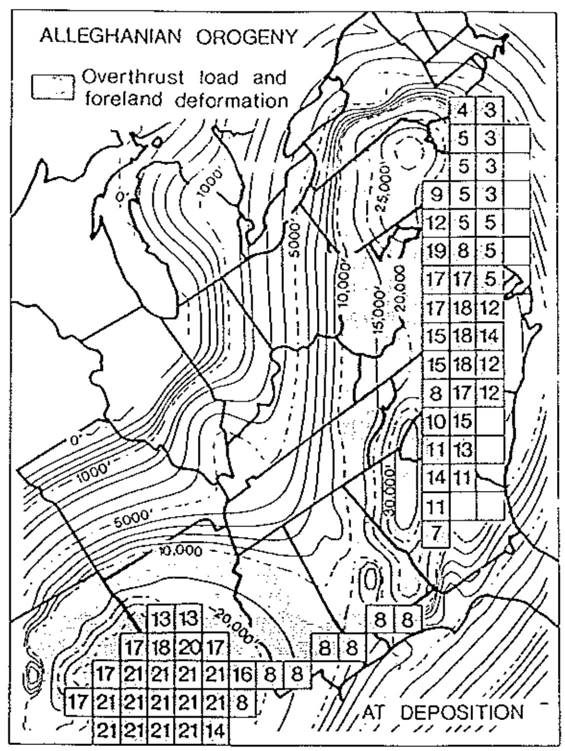


FIGURE 20 Total Alleghanian orogeny isopach map. Model prediction at deposition showing the cumulative load thickness for the Pennsylvanian and Permian (km). Contours of sediment isopach are in feet.

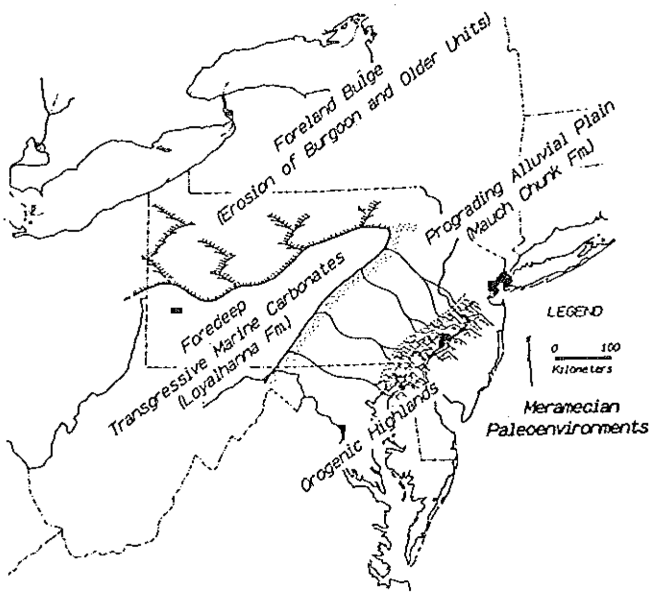


FIGURE 21 Paleogeography of the middle Atlantic states during the Meramecian (early Late Mississippian). Continued relaxation of the forebulge allowed transgression of the Loyalhanna sea into Pennsylvania as the Acadian orogenic highlands continued to downwaste, depositing the Mauch Chunk Fm. (modified from Edmunds, et al, 1979).

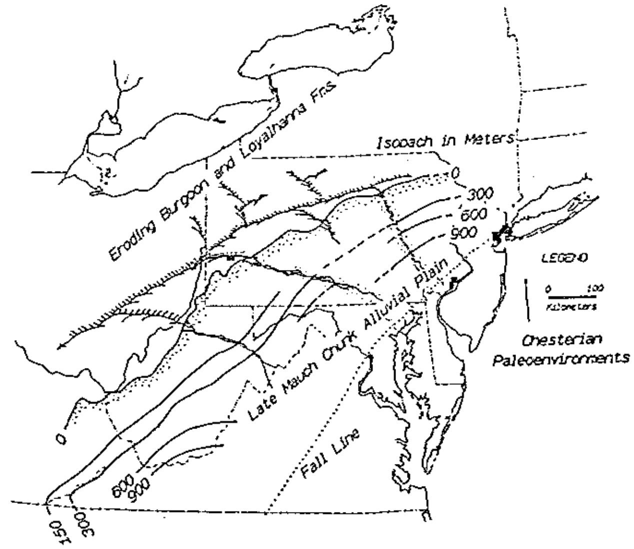
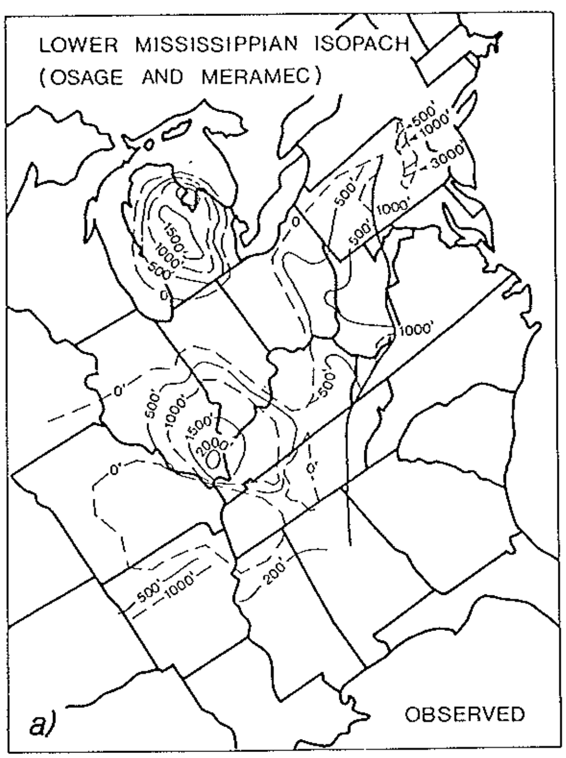


FIGURE 22 Paleogeography of the middle Atlantic states during the Chesterian (middle Late Mississippian). Increased loading to the south (Fig. 24) renewed uplift of the forebulge region, causing erosion of previously deposited formations to the northwest as Mauch Chunk alluvium continued to accumulate to the southeast (modified from Edmunds et al., 1979).



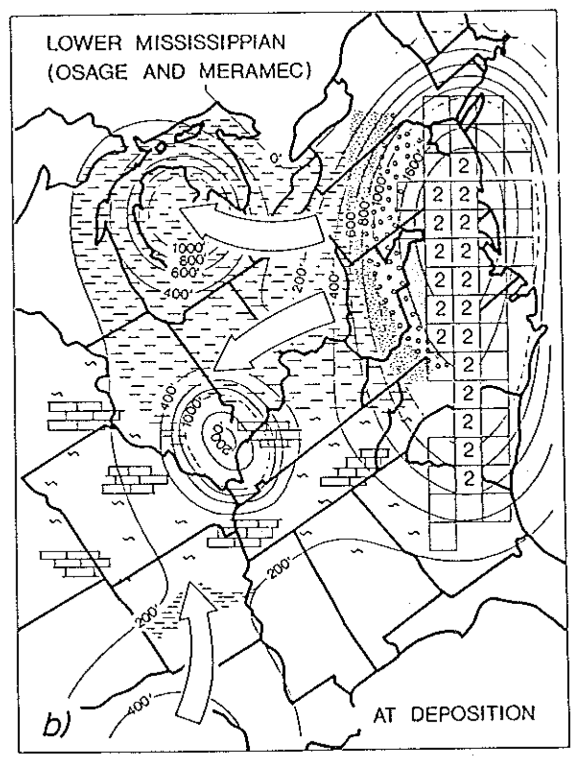


FIGURE 23 Isopach maps of observed (panel a) and predicted (panels b and c) sediment distribution for the Lower Mississippian (contours in feet). The numbered grids in panel (b) are the thicknesses (km) of the overthrust loads necessary to produce the model subsidence in the Appalachian basin. Brick and tilda patterns denote observed and restored chemical sedimentation and marine conditions. Dash and dot patterns denote observed shale, and fine and coarse sandstone sediments. Large arrows show the inferred major sediment dispersal directions.

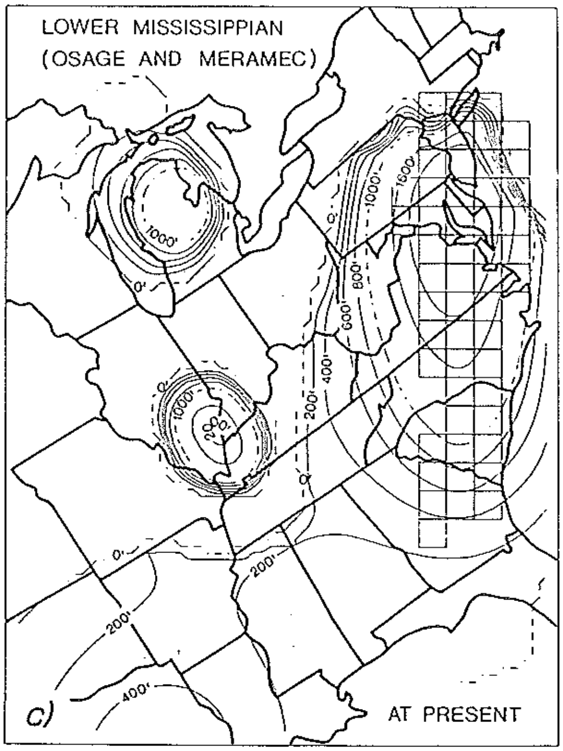


FIGURE 23 (cont.)

given in Figures 23 and 24. Although more subdivisions would be necessary to capture the dynamics of evolution during this period, the change appears to be the consequence of a southward migration of the load distribution and uplift of and erosion from the Findlay-Algonquin arch during an interval of lithospheric stress relaxation.

The southward load migration from the Late Devonian into the Mississippian meant that the areas of western New York and Pennsylvania became progressively closer to the edge of the foreland basin and, therefore, were more influenced by sediment influx from the uplifted arches (Figs. 12b, 23b, and 24b). The Bedford Delta (Fig. 16) is the first evidence of reworked cratonic sediments, presumably from a source that could have been as proximal as the vicinity of southern Ontario by the Kinderhookian to Osagean transition. During the Meramecian and Chesterian the edge of the basin retreated into Pennsylvania (Figs. 24b and 21), and older Mississippian sediments were uplifted, exposed, eroded, and reworked into the southeasterly retreating basin.

Pennsylvanian Depositional History. Commencing in latest Mississippian time in eastern Pennsylvania and continuing through the Middle Pennsylvanian, a wedge of braided stream gravels (eg., Pottsville Fm., Day 6, Site 1) flooded northwestward over the Mauch Chunk delta complex from a south-southeastern source terrane (Fig. 25). Although this has traditionally been

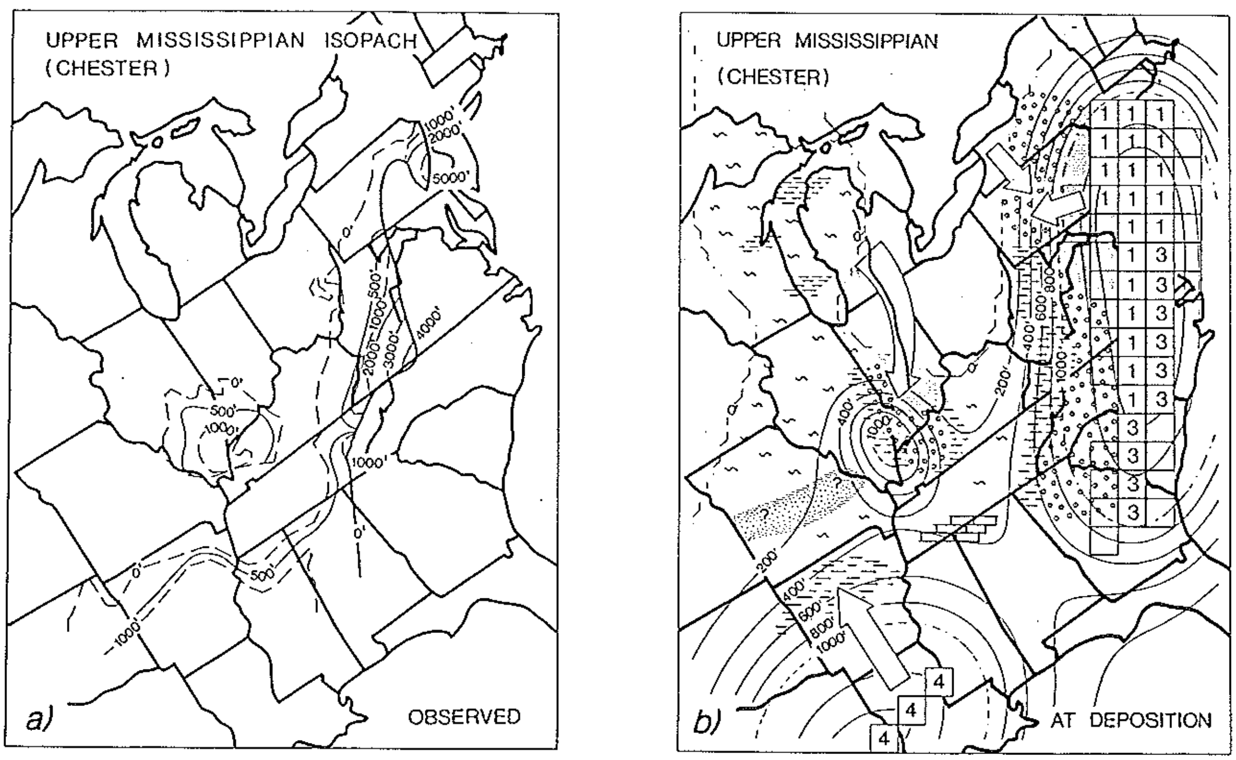


FIGURE 24 Isopach maps of observed (panel a) and predicted (panels b and c) sediment distribution for the Upper Mississippian (contours in feet). The numbered grids in panel (b) are the thicknesses (km) of the overthrust loads necessary to produce the model subsidence in the Appalachian and Arkoma basins. Brick and tilda patterns denote observed and restored chemical sedimentation and marine conditions. Dash and dot patterns denote observed shale, and fine and coarse sandstone sediments. Large arrows show the inferred major sediment dispersal directions.

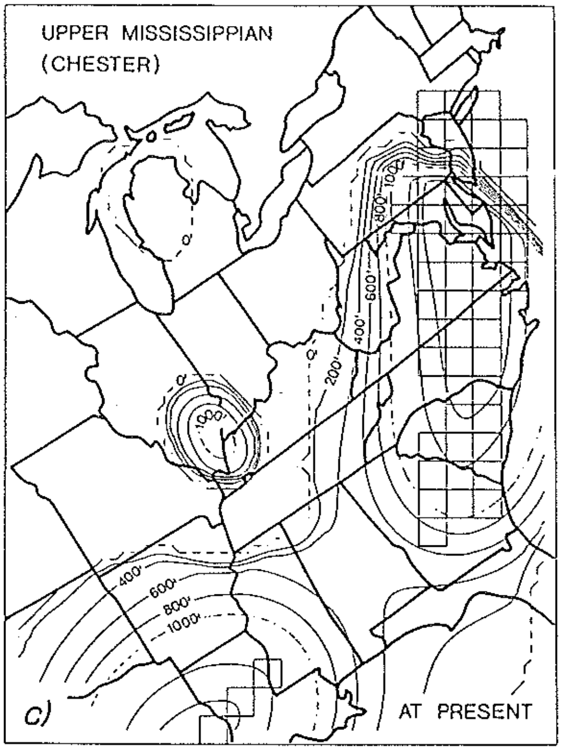


FIGURE 24 (cont.)

interpreted as evidence of dramatic orogenesis immediately to the southeast (cf. Meckel, 1967), the maximum accumulation rate at Pottsville, PA was only 19 m/Myr, and clast and rock fragment lithologies suggest the source terrain was composed primarily of sedimentary and low grade metamorphic rocks (Meckel, 1967; Houseknecht, 1979). The flexural modelling presented below (Fig. 27), suggests only modest additional thrust loads were present outboard of Pennsylvania during the Early Pennsylvanian. As described at Site 1 on Day 6, an alternative explanation for these gravels is a change to a wetter climate and higher discharge, perennial streams draining the Virginia orogenic belt.

At approximately the same time, gravels derived from older Paleozoic sedimentary rocks to the north filled incised stream channels along the northern tier of Pennsylvania (Meckel, 1967). This was apparently in response to subsidence below base level caused by crustal loading in Virginia and further south and is the first evidence of the type of transition in loading between that shown in Figures 27 and 28.

By early Desmoinesian (Middle Pennsylvanian) time, subsidence and eustatic sea level rise (Heckel, 1986) were sufficient to flood western Pennsylvania (Figs. 26 and 28), creating broad delta plains conducive for the formation of coal swamps (eg., Allegheny Group, Day 4). At the same time, alluvial plain slopes

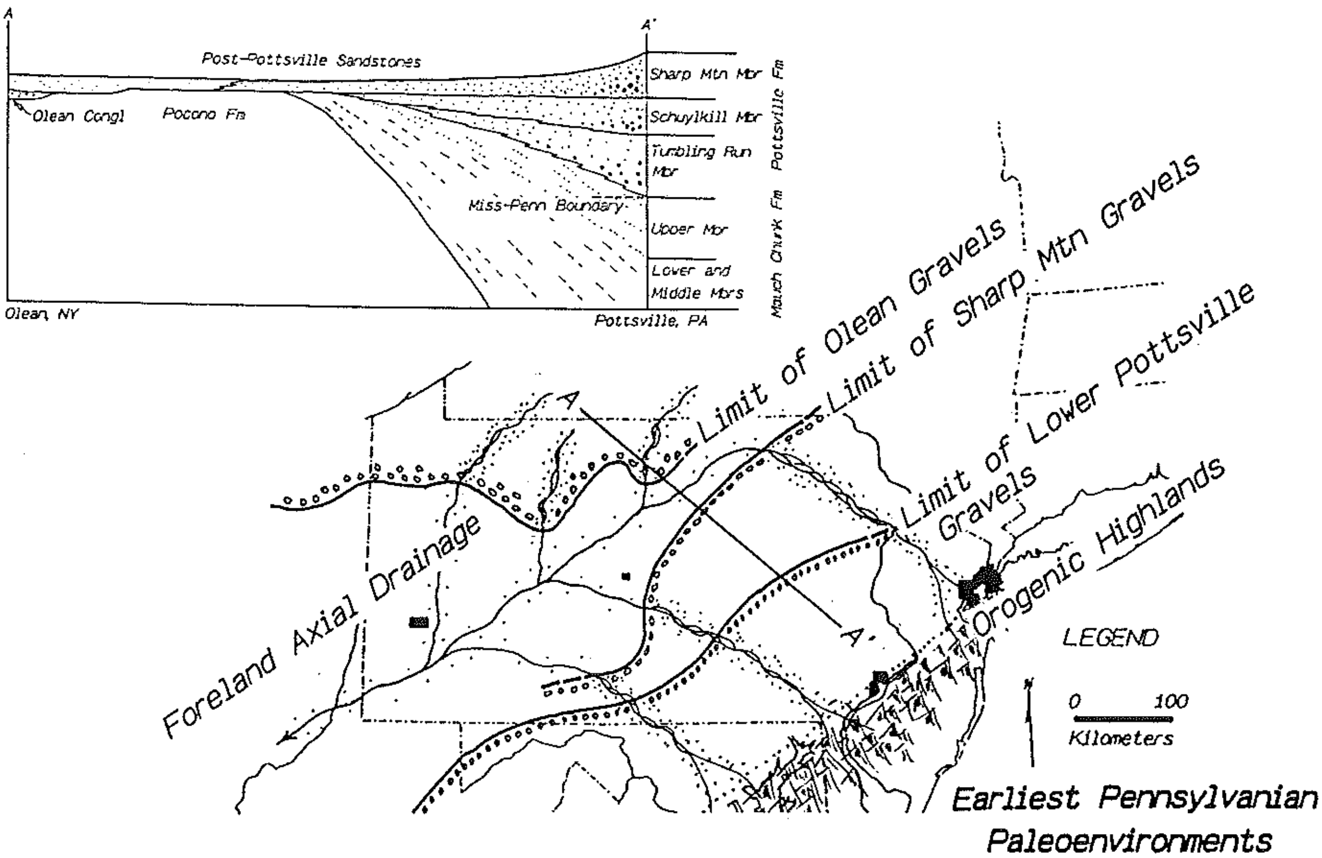


FIGURE 25. Paleogeography of the middle Atlantic states during the earliest Pennsylvanian. Major loads to the south-southeast (Fig. 27) depressed Pennsylvania and created differential relief such that a flood of gravels (Pottsville, Olean, Sharon Fms.) swept over the region from the north as well as the southeast (modified from Meckel, 1967).

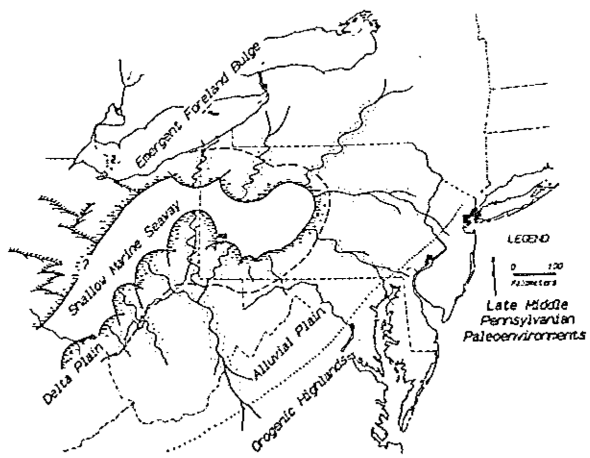


FIGURE 26 Paleogeography of the middle Atlantic states during the late Middle Pennsylvanian. As the overthrust loads migrated northward (cf. Figs. 27 and 28), a narrow seaway returned to the area and the alluvial plain spread further northwestward from the orogenic highlands (modified from Donaldson and Shumaker, 1981).

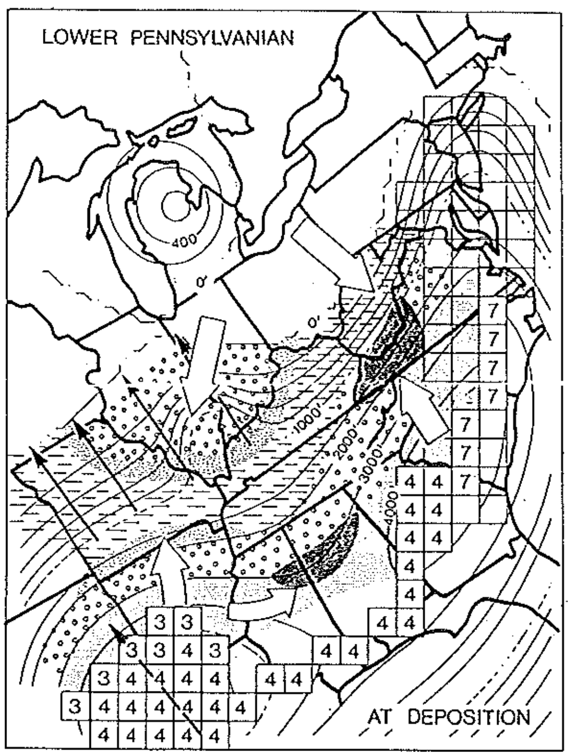


FIGURE 27 Lower Pennsylvanian isopach map (contours in feet). Model prediction at time of deposition showing the load thicknesses (km) necessary to produce the model subsidence in the Appalachian and Arkoma basins. Dash and dot patterns denote observed and restored shale, and fine and coarse sandstones. Light and dark shading represent coastal plain and coal swamp environments. Large arrows show the inferred major sediment dispersal directions. Fine arrows show the migration of the peripheral bulge during the advance of the overthrust loads in the Ouachita orogen.

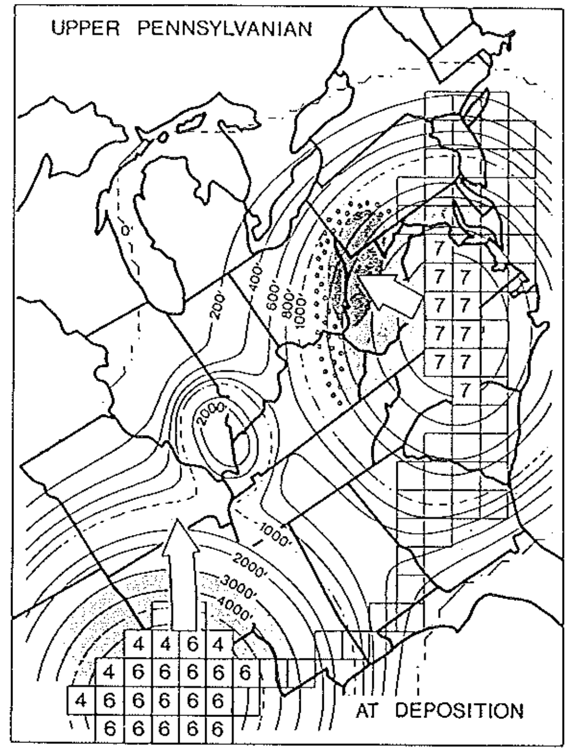
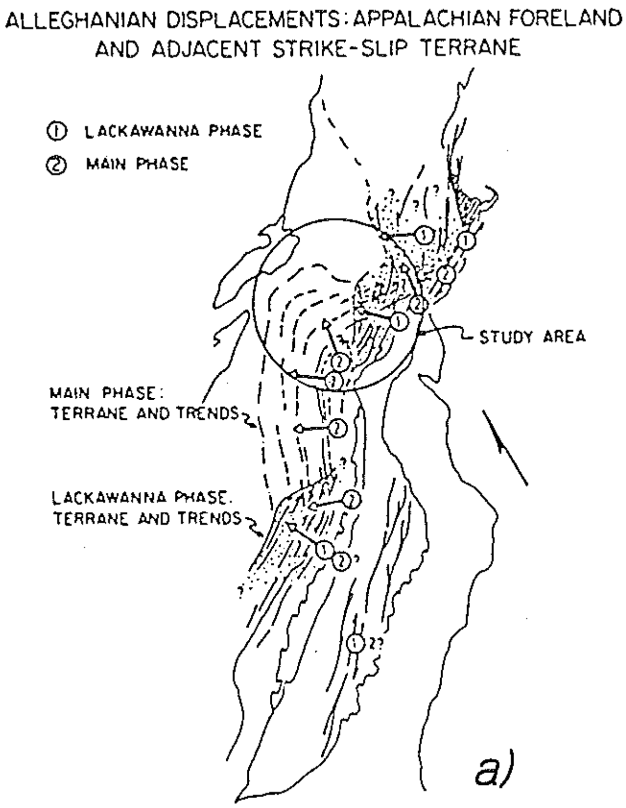


FIGURE 28 Upper Pennsylvanian ispoach map (contours in feet). Model prediction at time of deposition showing the load thicknesses (km) necessary to produce the model subsidence in the Appalachian and Arkoma basins. Dot pattern denotes sandstones. Light and dark shading represent coastal plain and coal swamp environments. Large arrows show the inferred major sediment dispersal directions.

declined, probably due to increased loads and subsidence near the source, and thick peat swamps developed as close to the source terrane as Pottsville, PA (eg., Llewellyn Fm., Day 5, Site 2, and Day 6, Site 1).

Throughout the Upper Pennsylvanian into the Permian, base level oscillated, producing the famous cyclothems of the coal measures (Busch and Rollins, 1984; Heckel, 1986). Accumulation rates increased, presumably in response to increased thrust loading to the east. By the time the youngest preserved strata were deposited, accumulation had so outstripped subsidence, that the northeastern end of the Appalachian foreland basin was a subaerial alluvial plain stretching from its southeastern source terrane to the Cincinnati Platform.

The regional setting of Pennsylvanian subsidence and sedimentation can be explained in terms of the flexural model results (Figs. 27 and 28). Like the Mississippian, more subdivisions would be necessary to represent details, but the pattern is apparently explained by the initiation of substantial overthrusts along the southern rim of North America and the progressive migration of overthrusting northward into Virginia during the middle and later part of the Pennsylvanian. The shift in the locus of loading from



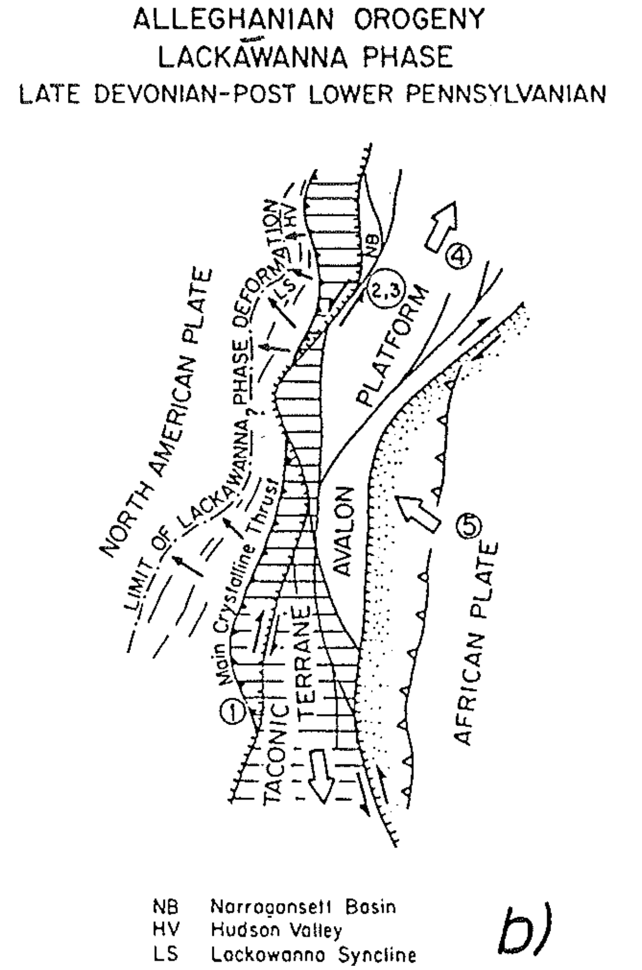


FIGURE 29. a) Directions of layer parallel shortening during the Lackawanna (Pennsylvanian?) and main (Permian?) phases of the Alleghanian orogeny. b) and c) Tectonic interpretation of deformation within the Appalachian orogen during these phases (modified from Geiser and Engelder, 1983). The inferred directions of thrusting should be compared with the positions of overthrust loads shown in Figures 24, 27, 28, and 30.

that of the Late Mississippian causes uplift and exposure of sediments on a broad east-west peripheral bulge (Fig. 27), which explains the increasing importance of this region as a source of reworked sediments and the development of an unconformity over this area.

The northward sweep in the development of the unconformity can be attributed to the migration of the peripheral bulge at the time of rapid convergence between the loads and the continental margin (Beaumont et al, 1988; Ettensohn and Chestnut, sub.). The development of the unconformity in this southern region therefore parallels that of the Ordovician post-Beekmantown Knox unconformity in the central Appalachian foreland north of Alabama. During the early stages of convergence of overthrusts on a rifted margin, these loads migrate large distances laterally as subduction continues. The peripheral bulge therefore sweeps across a much larger area than in subsequent orogenies at the same margin where the loads tend to grow in situ by shortening and thickening of terranes that have already accreted.

Erosion in central and northwestern Pennsylvania in the Early Pennsylvanian is therefore attributed to uplift of the Findlay-Algonquin arch during stress relaxation that continued from the Mississippian compounded with the superimposed peripheral bulge from the newly arrived southern loads. Although eustatic sea level changes cannot be entirely dismissed as the cause of the Pennsylvanian unconformity, the tectonic-flexural model provides an internally consistent explanation.

Upper Pennsylvanian sedimentation in the central Appalachian basin is attributed to the completion of the Lackawanna phase of the Alleghanian orogeny (Fig. 28). Deformation caused by thrusting from load emplacement in central Virginia also agrees with the explanation of layer parallel shorting during the Lackawanna phase (Geiser and Engelder, 1983)(Fig. 29) which requires northwesterly directed compression in Pennsylvania. This phase of layer parallel shortening should be contrasted with the later, "main" phase related to Permian compression and loading (Figs. 29 and 30) when the compression was directed towards the west. Although all of the loading of this later phase cannot be unequivocally termed Permian, structural, sedimentological, coal metamorphic and moisture, and fission track data (summarized in Beaumont et al., 1987) require a further stage of loading which to first order is satisfied by the model (Fig. 30).

The main phase of Early Permian compression deformed the foreland itself, thus ending its life as a

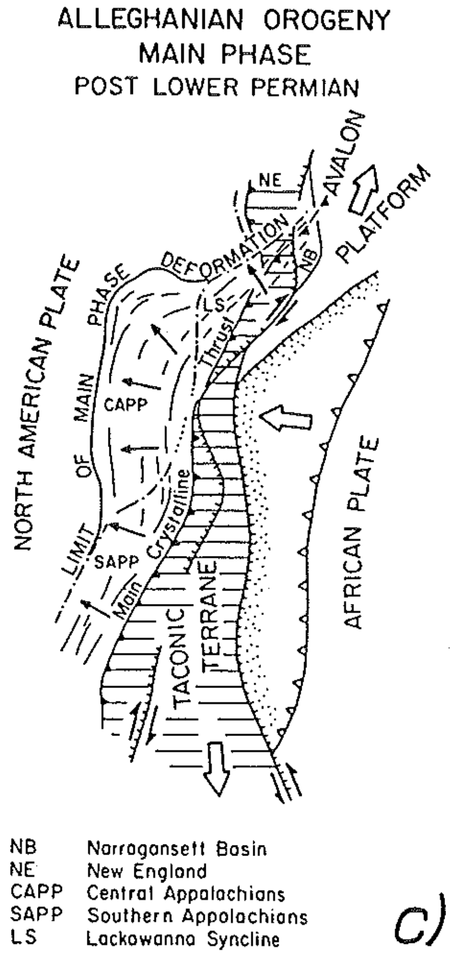


FIGURE 29 (cont.)

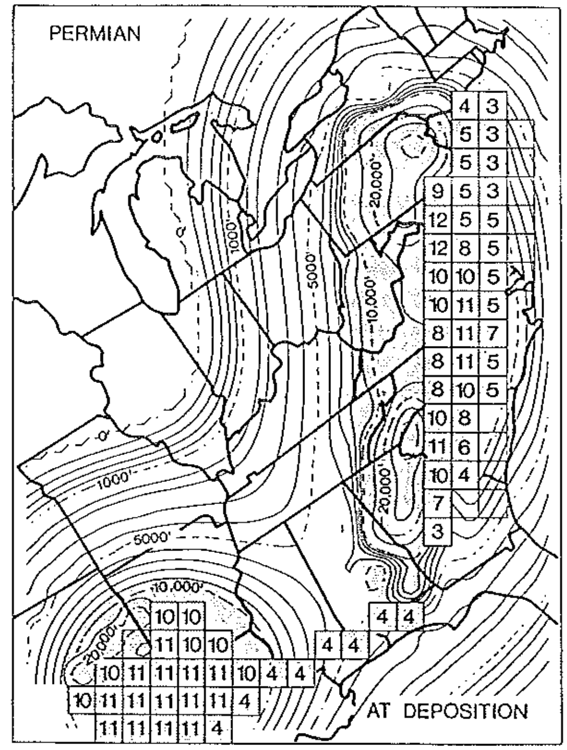


FIGURE 30 Permian isopach map (contours in feet). Model prediction at time of deposition showing the load thicknesses (km) necessary to produce the model subsidence in the Appalachian and Arkoma basins. Shading represents regions deformed by thrusting in which the extra thickness of sediment may, in part, be occupied by older sediments that were shrotened and thickened during thrusting.

sedimentary basin. Significant erosion at a rate of 200 m/Myr started almost immediately as evidenced by the fact that Late Triassic rift basin sediments overlie Cambro-Ordovician carbonates in southeastern Pennsylvania (see Manspeizer and Huntoon, this volume, for more details). By comparison with denudation rates as functions of relief for present-day mountain belts in similar latitudinal (and therefore climatic) settings, and given the thickness of the loads (Fig. 30), the ancestral Appalachians must have had considerable relief.

EARLY MESOZOIC RIFT BASINS OF EASTERN NORTH AMERICA: ORIGIN AND EVOLUTION

Warren Manspeizer and Jacqueline Huntoon

INTRODUCTION

Continental rift basins form in response to divergent stresses that may eventually result in the separation of continental plates and formation of new oceanic crust (Rosendahl, 1987). Because of their unique geographic position, these basins potentially contain much information about the geodynamic processes responsible for plate motions on earth. In geologic examples, however, basin analysis is complicated by the fact that the strictly tectonic signature is obscured by the superimposed effects of weathering, erosion, and sedimentation. The most complete source of data reflecting the complex interaction of tectonic activity and climatically controlled processes is the sedimentary package contained within a subsiding rift basin. By viewing sedimentary rocks as a record of the dynamic behavior of the earth system through time, this data set can be used to distinguish among the effects of the various mechanisms controlling basin evolution.

Early Mesozoic tectonic activity in the Appalachian orogen was a consequence of the breakup of Pangaea and the opening of the Atlantic Ocean. These events embrace a major tectonic cycle marked by Late Triassic-Early Jurassic rifting and Middle Jurassic to Recent drifting. The rift stage involved heating and stretching of the crust accompanied by uplift, faulting, basaltic igneous activity and rapid sedimentation in deep elongate fault troughs. The following drift stage, which involved slow cooling of the lithosphere over a broad region, was accompanied by thermal subsidence with concomitant marine transgression of the newly formed plate margin. The changeover from rifting to drifting records a fundamental change in the tectonic history of the Appalachians that corresponds to the onset of sea-floor spreading in the Middle Jurassic.

This paper and the accompanying outcrop descriptions address several aspects of Early Mesozoic rifting and the evolution of the Atlantic passive margin. First, we describe the structure and tectonic setting of Mesozoic rift basins within the Appalachian orogen. Second, we focus on the stratigraphy, facies, and inferred depositional systems of the basin fill. Third, we examine the timing of tectonic events, and the interaction between tectonism, paleoclimates, and sedimentation. And fourth, we discuss thermal-mechanical models that describe continental rifting and passive margin development.

Special emphasis is given throughout this paper to the Newark Basin, which is the largest, and perhaps best known, of these rift basins and the object of our field trip.

TECTONIC SETTING

Basin Structure

Approximately 40 to 50 northeast-trending elongate basins are found within the Variscan-Alleghanian orogen and on the bordering cratons of Africa and North America (Fig. 1). These elongate, northeast trending basins typically follow the fabric of the orogen. About 25 onshore and offshore rift basins have been identified on the American Plate. Almost all of the exposed basins, and presumably the offshore basins, appear to have developed along re-activated Paleozoic low-angle thrust or dextral strike-slip faults. The Fundy Basin, for example, occurs along the Avalon and Meguma suture, which according to Brown (1986) is a Variscan thrust fault (the Fundy Decollement) that formed as a compressional component of the Cobequid-Chedabucto transform.

Most of the rift basins are asymmetric half-grabens, bounded on one side by a system of major high-angle normal faults that are en echelon, curved, or otherwise offset, and on the other side by a gently sloping basement with sedimentary overlap and/or by secondary normal faults. The exposed basins are aligned with right-stepping offset in the Northern Appalachians and a modified left-stepping offset in the Central and Southern Appalachians (Fig. 1). They thus appear to be linked to each other by strike-slip faults that have been identified in other rift basins as transform segments (Bally, 1981), transfer faults (Gibbs, 1984), and accommodation zones (Burgess et al., in press). Whereas none of the major onshore basins conform to a classical graben structure (e.g. the Red Sea), paired and symmetrical basins on the Long Island Platform and those astride the Piedmont Gravity High may have had an early graben history. For these basins, asymmetry may have resulted from subsequent uplift along the graben axis (this Broad Terrane Hypothesis is discussed below).

In the Newark Basin, the western border fault lies along an older, gently dipping stack of imbricate thrust slices (Ratcliffe and Burton, 1985). The narrow corridor or neck, connecting the Newark and Gettysburg Basins, occurs along a major east-west lineament (Fig. 1 and 4), that includes the: (a) N40°-Kelvin transform (Van Houten, 1977; Manspeizer, 1980); (b) Transylvania continental fracture zone (Root and Hoskins, 1977); (c) sinistral Chalfont fault (Sanders, 1963), and the (d) prominent east-west deflection of the basement hinge-zone (Hutchinson and Klitgord, in press, b). Structurally paired basins with basinward-dipping listric faults and outward-dipping strata of the Long Island Platform (e.g. the

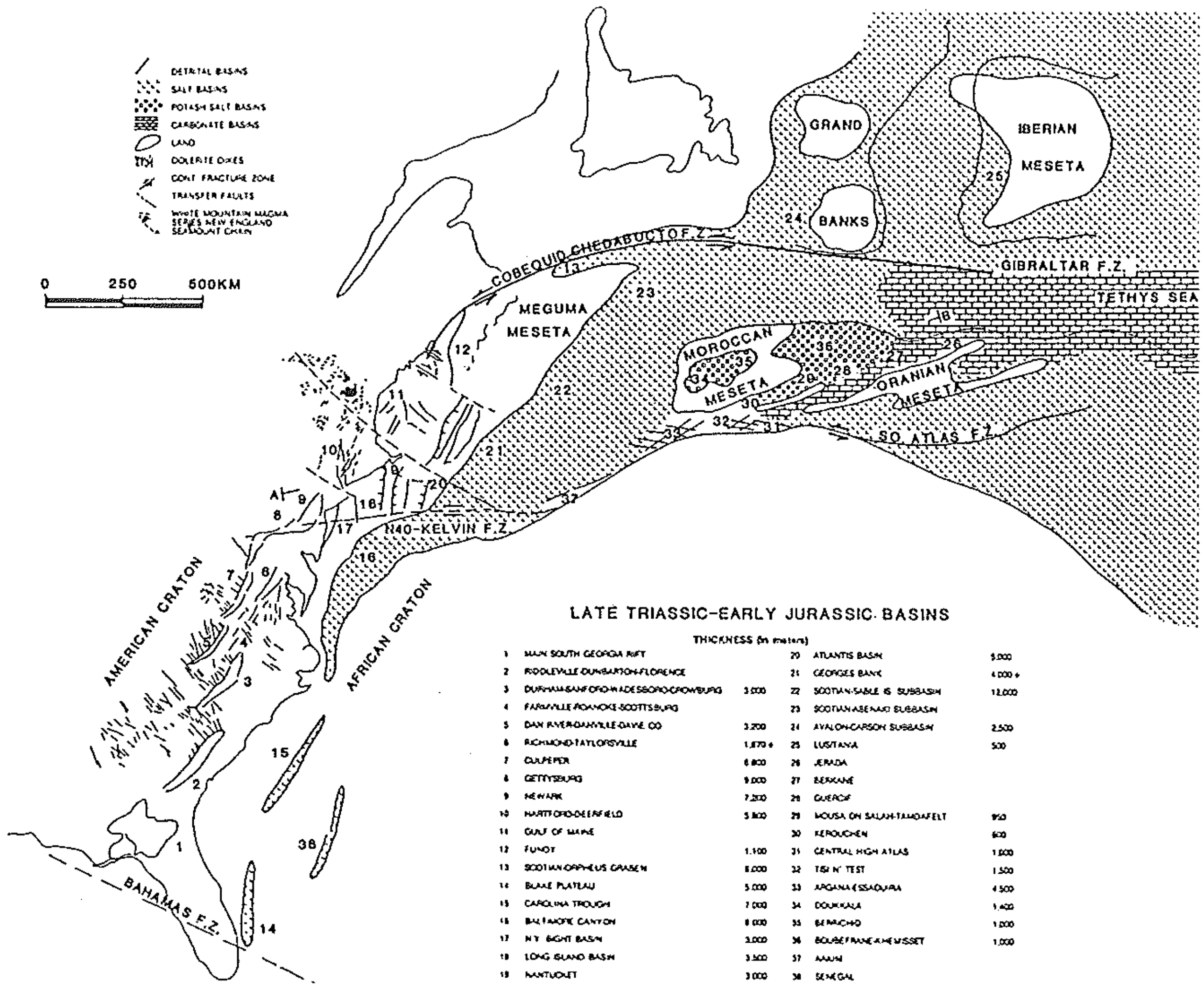


Figure 1 Lithofacies map of the North American and African continental margins at the time of initial rifting (Modified from Manspeizer, 1981).

Newark-New York Bight, Nantucket-Atlantis and Long Island basin also occur along the lineament, suggesting that they may have formed as extensional or transtensional basins along east-west trending transforms.

Strata within each basin typically dip about 10^{o}-15^{o} toward the border fault, where they are commonly bent into broad synforms and antiforms (Fig. 2), termed warps by Wheeler (1939), and/or into more tightly compressed en echelon folds (Davis, 1898). Wheeler (1939) related the warps in the Newark and Hartford Basins to differential dip-slip along salients and reentrants of the border fault. Later studies by Sanders (1963), Faill (1973), Manspeizer (1980) and Ratcliffe (1980) related folds within these basins to horizontal stress in a compressional setting. En echelon foreland-type folds with axial plane spaced cleavage (which is the subject of one of the fieldstops) have been mapped in the Jacksonwald Syncline near the narrow neck, connecting the Newark and Gettysburg Basins (Lucas et al., in press); these structures formed under regional compression, perhaps related to transpression and basin formation.

Strata within these Mesozoic basins are also cut by major oblique-trending cross faults. Sanders (1963) reports that in the Newark basin some faults have as much as 20 km of horizontal displacement and 3 km of vertical displacement, and Van Houten (1969) reports that some northeast trending strike-slip faults may be part of a strike-slip system involving the Ramapo border fault (Fig. 1 and 2). Published geophysical and subsurface data from the Newark-Gettysburg Basin (Sumner, 1977; Cloos and Pettijohn, 1973), Hartford Basin (Wenk, 1984), the Durham Basin (Bain and Harvey, 1977), and the Sanford Basin (Deep River; Randazzo et al., 1970) show that the basement is cut by cross faults creating intrabasin grabens and horsts that may have formed as growth faults (see Sumner, 1977; and Cloos and Pettijohn, 1973). However, field data (e.g. Faill, 1973; Lucas et al., in press) also show that some rift-related deformation of the onshore basins postdates the youngest rift strata (Early Jurassic), and may have continued into the Middle Jurassic, when sea-floor spreading and drifting began

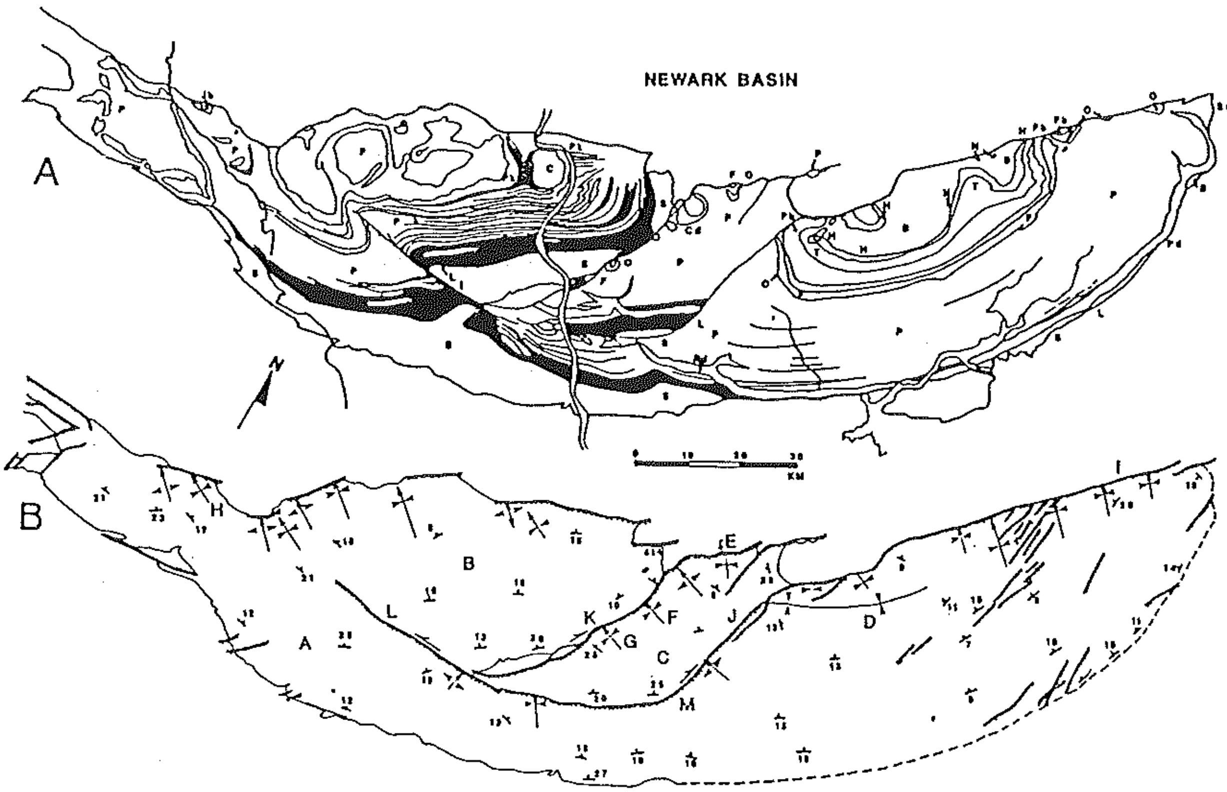


Figure 2 Lithologic map (A) and structural map (B) of the Newark Basin. The Ramapo Fault forms the northwestern margin of the basin (Modified from Manspeizer and Olsen, 1981).

offshore. Seismic profiles across the passive margin show that, except for intrusions by salt and igneous rocks, the younger drift strata appear essentially undisturbed (Fig. 3).

Lithospheric Structure and Geophysical Characteristics

The deformational history of the Appalachian orogen (discussed in the Introduction to this field guidebook) determined the large-scale lithospheric structure of eastern North America at the onset of Mesozoic rifting. The earliest events, associated with Late Proterozoic rifting, involved substantial heating and consequent weakening of the lithosphere (Bodine et al., 1981). This thermal perturbation had decayed for more than 100 m.y., and therefore an equilibrium thermal structure was re-established by the time the region was first over-thrusted during the Ordovician

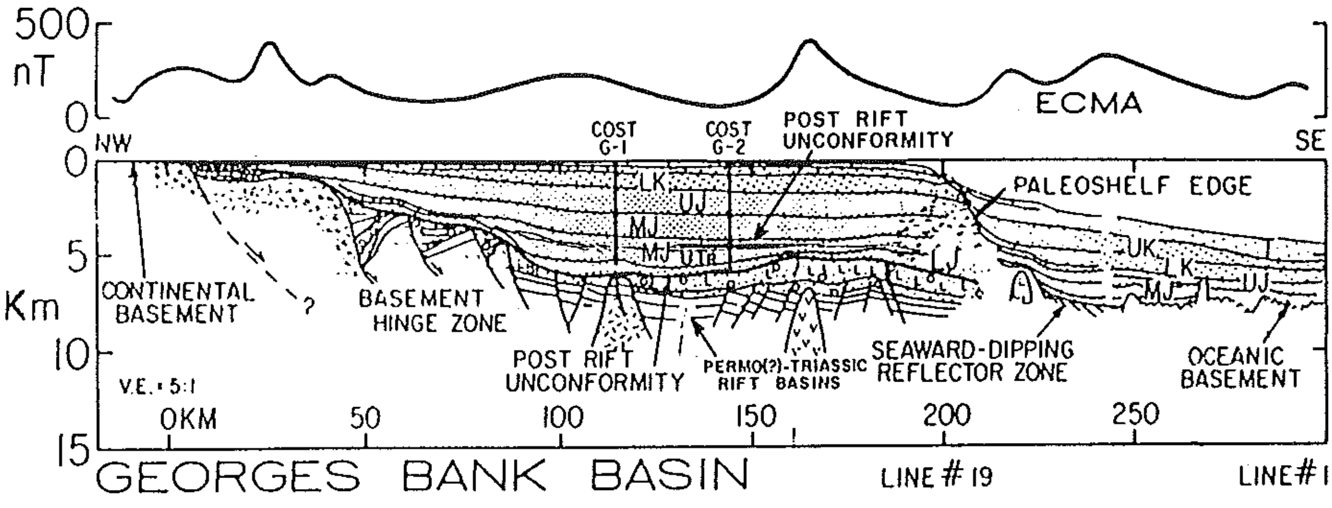


Figure 3 Interpreted seismic section of the Georges Bank Basin area of the Atlantic Margin of North America. Rift phase sediments are found within the several half-grabens shown. Drift phase strata are continuous across the margin, and mantle the rift phase basins (Modified from Klitgord and Hutchinson, 1985).

(Huntoon and Furlong, 1987; Jarvis and McKenzie, 1980; Sleep, 1971). The thermal effects of Paleozoic thrusting, plutonism, and volcanism were then superimposed on the background equilibrium thermal structure. By the time Mesozoic rifting began in the Late Triassic, however, steady state thermal conditions were once again reached. The high levels of organic thermal maturity observed in some rocks from the Mesozoic basins (Braghetta, 1985; Hatcher and Romankiw, 1985; Pratt et al., 1985), the linear relationship between the subsidence of the passive Atlantic margin of North America and $t^{1/2}$ (where t is time since rifting) that is characteristic of cooling lithosphere (Parsons and Sclater, 1977; Sleep, 1971), and the common observation of high heat flow in active continental rifts (eg. Blackwell, 1978; Chapman and Pollack, 1975) all suggest that Mesozoic continental extension and rifting resulted in a third major tectono-thermal event affecting eastern North America.

Bouguer gravity across the Appalachian orogen is presently complex, reflecting superposition of several phases of deformation. The onshore Mesozoic basins are distributed about the axis of the Piedmont (Appalachian) gravity high (Longwell, 1943) (Fig. 4). This feature is a prominent, long wavelength anomaly attributed to density variations at depth (Kane, 1983). A deep source for the anomaly is also suggested by its continuity along the length of the Appalachian orogen despite changes in surface geology. Near-surface lithologic units do not significantly affect the gravity signature across most of the orogen because of their limited vertical extent. Rocks exposed at the surface were laterally displaced up to a maximum of several hundred kilometers during Paleozoic thrusting (Cook et al., 1983; Cook et al., 1979). The Appalachian gravity high is interpreted to reflect an area of crustal thinning (Cook and Oliver, 1981), possibly corresponding to the Proterozoic hingeline of the Laurentian continental margin (Cook, 1984) (Fig. 5). Other interpretations have also been offered however (eg. Nelson et al., 1986). A change in Moho depth is likely responsible for the Appalachian gravity high because of the large density contrasts at the crust-mantle boundary (Kane, 1983). The spatial coincidence of the onshore Mesozoic basins and the Proterozoic hingeline suggests the presence of lithospheric-scale discontinuities in that region.

Major continental thinning associated with Mesozoic rifting begins at the Atlantic margin hingeline (Fig. 4) which is buried beneath several kilometers of sediments (Watts and Thorne, 1984; Sheridan, 1976), and is located within 100 km of the present shoreline in the central Appalachians. Bouguer

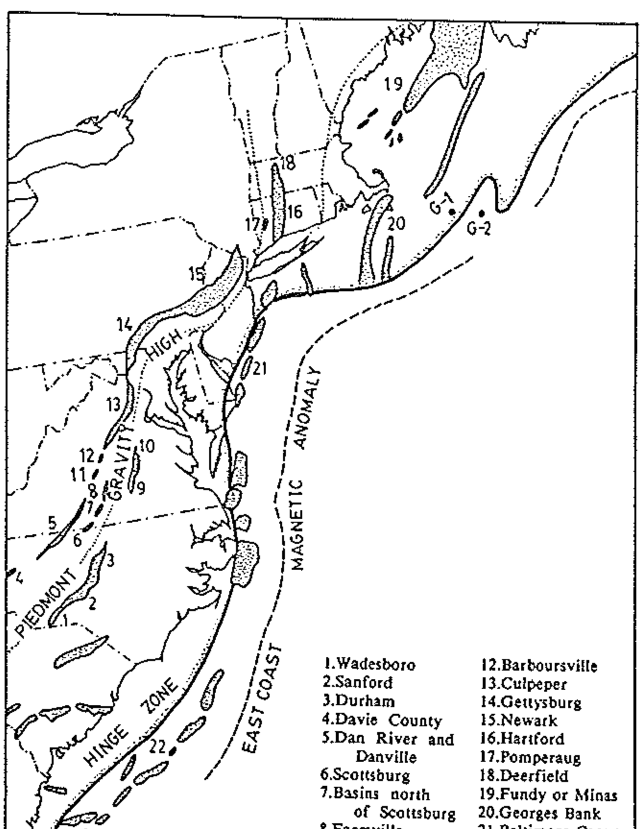


Figure 4 Location of the Mesozoic basins along the U.S. Atlantic margin. Locations of the Appalachian (Piedmont) gravity high (dotted line), Mesozoic hingezone (heavy solid line), East Coast Magnetic Anomaly (dashed line), and COST G-1 and G-2 wells are also shown (Modified from Manspeizer and Cousminer, 1988).

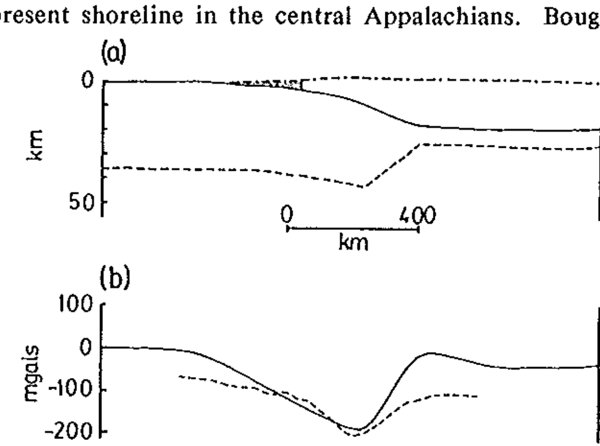


Figure 5 Lithosphere geometry and gravity anomaly of the Canadian Cordillera, an ancient passive margin that has been overthrusted during a collisional event. This example demonstrates that Moho geometry established during an ancient rifting event may persist throughout and after fold and thrust belt development. (a) Modelled configuration of Moho (dashed line), ancient passive margin (solid line), and land surface after approximately 35 m.y. of erosion of the fold and thrust belt. Passive margin sediments (dot pattern) are shown schematically. (b) Observed Bouguer gravity across the Cordillera (dashed line), and computed gravity anomaly based on the model in (a) (solid line) (Modified from Stockmal and Beaumont, 1987).

gravity values become increasingly positive across the hingeline region (Haworth et al., 1980; Grow et al., 1979a). This gradual increase in Bouguer gravity is consistent with recent seismic observations that indicate there is no abrupt change in Moho depth in the hingeline area (Hutchinson et al., 1986).

The magnetic signature of the orogen at the onset of Mesozoic rifting reflected the superposition of allochthonous thrust sheets and micro-plates upon the deep structure related to Proterozoic rifting. There is great variability in the magnetic mineral content of crustal rocks, and magnetic anomalies are usually the result of variations in near-surface lithology (Telford et al., 1978). In the Appalachian orogen, shallow changes in rock types across fault planes may mask the effects of regional changes in lithosphere structure at depth. The Mesozoic basins themselves are generally associated with a negative magnetic anomaly (Klitgord and Hutchinson, 1985) reflecting the relatively non-magnetic character of the basin fill sediments and minor volcanics. Magnetic and gravity highs flank the basins indicating the presence of denser, more mafic material surrounding the basins (Bell et al., 1988; Klitgord and Hutchinson, 1985). Magnetic anomalies associated with large basins are segmented into smaller sections, probably reflecting the position of initial offsets in the continental crust (Klitgord and Behrendt, 1979).

The East Coast Magnetic Anomaly (ECMA) is the most significant magnetic feature observed along the length of the Appalachian orogen today. It has been interpreted as an indication of highly deformed transitional crust (Schlee et al., 1976; Mattick et al., 1974), or the boundary between thinned continental crust and oceanic crust (Grow et al., 1979a,b; Klitgord and Behrendt, 1979). The latter interpretation is preferred because of the relative position of the ECMA seaward of the Mesozoic hingeline which, based on gravity evidence, probably represents the transition from normal to attenuated continental crust. The age of the ECMA is approximately 175-190 Ma (Klitgord and Schouten, 1986, p. 364), and it is probably associated with the first oceanic crust produced as a result of Mesozoic rifting of North America and Africa. This anomaly possesses a distinctly segmented pattern similar to that observed where transform faults offset the Mid-Atlantic Ridge (Schouten et al., 1985). It is unclear whether this basic fragmentation is a reflection of discontinuities in the continental lithosphere at the time of rifting, or the basic nature of sea-floor spreading.

Seismic studies of the continental margin in the central Appalachians led to the development of three models of rift basin evolution (Hutchinson and Klitgord, in press, a) (Fig. 6). In all three cases, brittle failure in the upper crust is substantiated by prominent reflections on several seismic lines, but the mode of extension in the lower crust is not clear. Observations suggest that different mechanisms, determined by the large-scale structure of the margin prior to rifting, controlled the evolution of the rift basins. For example, formation of basins along the axis of the Appalachian gravity high (including the onshore basins)

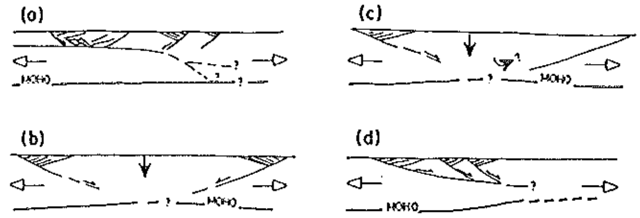


Figure 6 Models of Mesozoic rift basin development based on seismic data. See text for discussion (Modified from Hutchinson and Klitgord, in press, a).

appears to be associated with a discontinuity in the trend of a low-angle detachment surface at depth (Hutchinson and Klitgord, in press, a) (Fig. 6a). These half-grabens formed in pairs and the border faults of opposing basins dip toward one another. Border faults in these basins are planar as well as listric, and probably extend only to the level of the detachment (Behrendt, 1986; Hutchinson and Klitgord, in press, a). Crustal thinning may occur near the basin, or may occur kilometers to the east (Hutchinson and Klitgord, in press, a) if the basal detachment was active during extension and continues eastward as suggested by Ando et al. (1984) and Cook et al. (1981). The second type of basins formed between pairs of facing planar or listric faults that extend to deep crustal levels (Fig. 6b&c). Localized crustal thinning is observed between the faults (Hutchinson and Klitgord, in press, a). The third type of rift basin occurs along a low-angle detachment surface that continues to mid or lower crustal levels (Fig. 6d). In this type, adjacent basins form along sub-parallel border faults. Crustal thinning may occur adjacent to, or far removed from these basins. In general, offshore basins that formed near the hingeline are associated with extensive nearby crustal thinning because of the change from normal continental crustal thicknesses of 35-40 km to extended crust 10-25 km thick in this region (Hutchinson and Klitgord, in press, a; Watts, 1982; Grow et al., 1979a).

Basin Origin

After almost 150 years of study and debate, the origin of Mesozoic rift basins along the Atlantic continental margin remains controversial. Early workers, including Barrell (1915), considered the Triassic basins to be rather simple asymmetric fault grabens, produced by extension more or less orthogonal to the basin margins. The Broad Terrane Hypothesis, formulated by Russell (1922), and later modified by Sanders (1963), speculates that the Newark and Hartford Basins formed initially as a single large graben within an extensional regime; it was then arched, deformed by several episodes of horizontal shear, and subsequently eroded, producing two asymmetric half-grabens. Hutchinson and Klitgord (in press, b) suggest that the Newark-New York Bight Basin may have formed initially as a graben along the western edge of the Appalachian detachment surface. Late rifting in their model, resulting from regional uplift, caused tilting and erosion of synrift strata landward of the basement hinge-zone. The age of uplift and erosion is dated as Lias (Early Jurassic) by the postrift unconformity in the COST G-2 cores (Manspeizer and Cousminer, 1988).

Other studies have focused on the role of Late Triassic transforms as the possible mechanism through which wrench-induced, pull-apart basins and grabens are formed. Manspeizer (1981) interprets the Newark Basin as a strike-slip basin, which evolves through sinistral shear along the east-west trending N40°-Kelvin Lineament that includes the Narrow Neck between the Newark and Gettysburg Basin. Ballard and Uchupi (1975) also employed a similarly oriented, left-lateral shear couple to explain the origin of Triassic basins in the Gulf of Maine. Their conclusions are consonant with data from the Fundy Basin, showing Late Triassic-Early Jurassic sinistral slip of 75 km (Keppie, 1982) along the Cobequid-Chedabucto Fracture Zone. Ratcliffe and Burton (1985), invoking a constant stress field for the Newark Basin, however, note that extension across complex curvilinear thrust-ramp structures may produce the fault geometry and strike-slip characteristics of that basin. Lucas et al. (in press) also demonstrate that while the Narrow Neck may have originated in a wrench zone with sinistral shear, the Newark and Gettysburg Basins formed in sinistral transtension. But these same workers further conclude that no model employing a single stress field can explain all the features of these basins. Similar conclusions are drawn by deBoer and Clifford (in press), who use a three-phase system of Late Triassic extension, Early Jurassic strike-slip, and Middle Jurassic extension to explain the structures of the Hartford Basin. Venkatakrishnan and Lutz (in press), also noting that extension orthogonal to the marginal faults can not explain the structures of the Richmond Basin, postulate a changing stress field with both dip-slip and strike-slip deformation to explain the basin. While there is no unanimity on this issue, it is evident that a simple extensional model can not explain the origin of all these basins.

BASIN FILL

Newark Supergroup

Lithosomes. Rocks within the Mesozoic basins comprise the Newark Supergroup (Fig. 7). Typically they form wedge-shaped lithosomes that become thicker and coarser grained towards the bounding border faults, and finer grained towards the opposing hinged margins (Fig. 8). From the Culpeper Basin north to the Fundy Basin, lithosomes are dominated by Late Triassic-Early Jurassic fluvial-lacustrine strata that are interbedded, near the systemic boundary, with tholeiitic lavas. The lavas, dated as Hettangian or Early Liassic, were emplaced about 20 million years after synrift

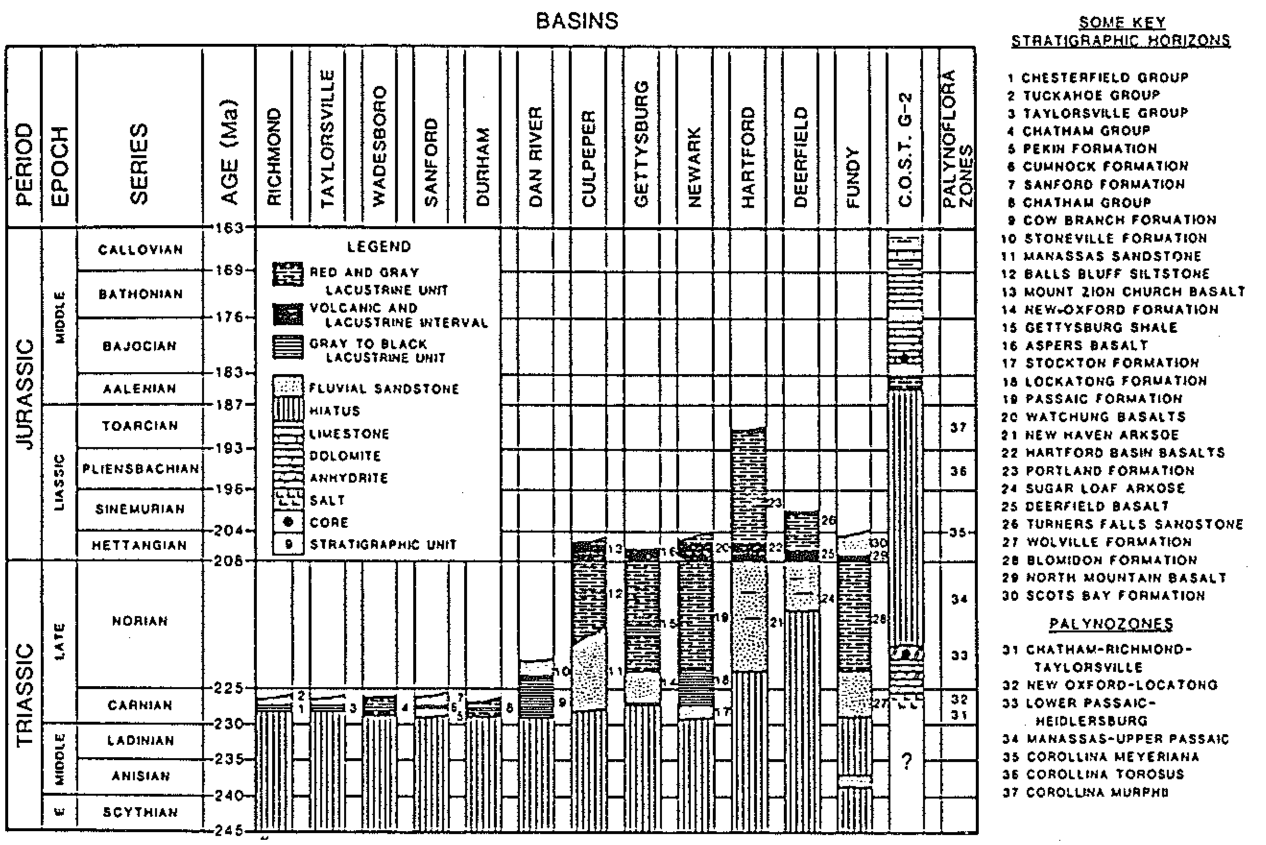


Figure 7 Time-correlation chart: interbasinal correlation of Newark strata based on palynofloral zones and/or extrusive horizons, data principally from Olsen, written communication, 1985. Correlation is also made with lower Mesozoic strata of the COST G-2 cores (From Manspeizer, 1981).

sedimentation began and after 2-8 km of clastics had accumulated in these basins. In the exposed southern basins, where the youngest synrift strata are dated as Late Triassic (Carnian or Norian), only the basal part of the lithosome is preserved. The occurrence of tholeiitic lavas in the concealed basins of South Carolina, Alabama, Florida and Georgia (Daniels et al., 1983), however, suggests that synrift sedimentation and volcanism also may have persisted into the Early Jurassic in the Southern Appalachians. Both the synrift sequence and the adjacent basement are cut by hypabyssal sills and dikes that are now thought, by some workers, to be essentially contemporaneous with volcanism. If this is correct, then all igneous activity within these basins occurred within a relatively brief interval of about 500,000 years in the Hettangian (Olsen, in press, a).

In contrast to the onshore detrital basins, the offshore synrift basins are largely evaporitic, filled with massive halite, dolomite, and anhydrite (Holser et al., in press; see Fig. 1, and discussion under evaporite facies).

Synrift sedimentation appears to have been slightly diachronous, beginning in the Carnian in all onshore basins, except for the Fundy Basin, where sedimentation may have begun as early as Anisian (Middle Triassic; Fig. 7). Although the data base is sparse for determining the age of the offshore basins, Carnian palynomorphs extracted from the COST G-2 well of Georges Bank indicate that rifting was essentially contemporaneous across the orogen (Manspeizer and Cousminer, 1988). Late Triassic sedimentation in most onshore basins consists of a lower dominantly fluvial interval, a middle 'deep water' gray or black lacustrine sequence, and an upper mostly lacustrine interval which is usually red (Olsen, in press, a). Within the context of Newark Basin stratigraphy and this field trip, these units are equivalent to the Stockton, Lockatong and Passaic Formations respectively (Fig. 8). Whereas as Olsen (in press, a) claims that these stratigraphic horizons are vertically isolated facies representing basin-wide paleoenvironments, other workers, e.g. Turner-Peterson and Smoot (1985) believe that they may be viewed as large-scale time-transgressive facies within formations that are largely isochronous across the basin (Fig. 9).

Dominance of one facies, fluvial or lacustrine, within each formation appears to reflect the expansion and contraction of lakes along the axis of these basins, while asymmetry of facies within each formation appears to record the dominance of different processes acting across the steep and gentle slopes of these basins (see discussion below). Gravity models of the Newark Basin suggest that sedimentary infill is less than 5 km thick (Sumner, 1977).

Lacustrine facies. Volumetrically, the majority of these beds are finely-laminated gray-black siltstones that record the cyclical expansion and contraction of a lake located near the center of the basin (Van Houten, 1969; Olsen, 1980; and Olsen, in press, b). Each lake cycle, according to Olsen (in press, b) consists of a sequence of three lithologically identifiable units (see

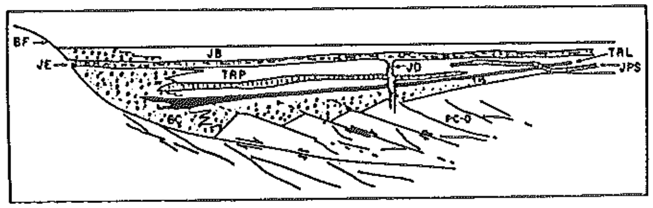


Figure 8 Hypothetical northeast-southwest cross-section of the Newark Basin along the Delaware River during depostion of the Early Jurassic Boonton Formation (JB). Abbreviations as follows: BF, border fault; JD, Early Jurassic diabase dikes; JE, Early Jurassic tholeiitic extrusives with interbeds of sedimentary strata; JPS, Early Jurassic Palisades sill and realted intrusives; BC, border conglomerate; TrP, Passaic Formation; TrL, Lockatong Formation; TrS, Stockton Formation; PC-O, Precambrian and Cambro-Ordovician rocks of the Taconic and Allegheny thrust sheets. Small arrows indicate direction of Taconic and Allegheny thrust movement during the Paleozoic; large arrows indicate movement of re-activated faults during the Mesozoic (From Manspeizer and Cousminer, 1988).

Fig. 7.1.6 in the field guidebook section). Division 1 records the transgression of the lake, and consists of platy to massive mudstones to conglomerates with some desiccation cracks, frequent ripple-to-large-scale crossbeds, root zones, and less frequently stromatolites, oolites, reptile footprints, and mollusc-rich zones. Division 2, recording the high-stand of the lake, is a platy to microlaminated, sometimes organic carbon-rich calcareous claystone, mudstone, or limestone rarely with desiccation cracks, but at times with a rich assemblage of ostracodes, clams, shrimp, insects, fish and reptiles. Division 3, recording the regressive component and low stand of the lake, consists of platy to massive calcareous mudstones to conglomerates with frequent desiccation cracks, fenestral fabrics filled with carbonates or zeolites, root and burrow zones, ripple-to-large-scale crossbeds, and reptile footprints. We will examine this sequence in the field.

The largest Lockatong Lake had an areal extent greater than 7000 m^2, and a depth in excess of 100 m (Manspeizer and Olsen, 1981). Although some lakes were deep, most were shallow and ephemeral, displaying mudstone fabrics that record extensive, and perhaps complete, desiccation (Smoot and Olsen, in press). The distribution of related lacustrine facies, e.g. alluvial fans and fluvial-deltas, is closely tied to the asymmetry of these basins, with alluvial fans prograding from the active fault margins, and fluvial-deltaic systems advancing from the gently sloping hinged and axial margins.

Alluvial Fan Facies. Alluvial fans, that prograded across the basin axis, away from the faulted or uplifted basin margin, are well-documented in the literature. A particularly enlightening example is offered by LeTourneau (1985), who has interpreted the Lower Jurassic Portland Formation of the Hartford Basin as a sequence of fault-margin fans that prograded onto a fluvial basin-floor during dry spells,

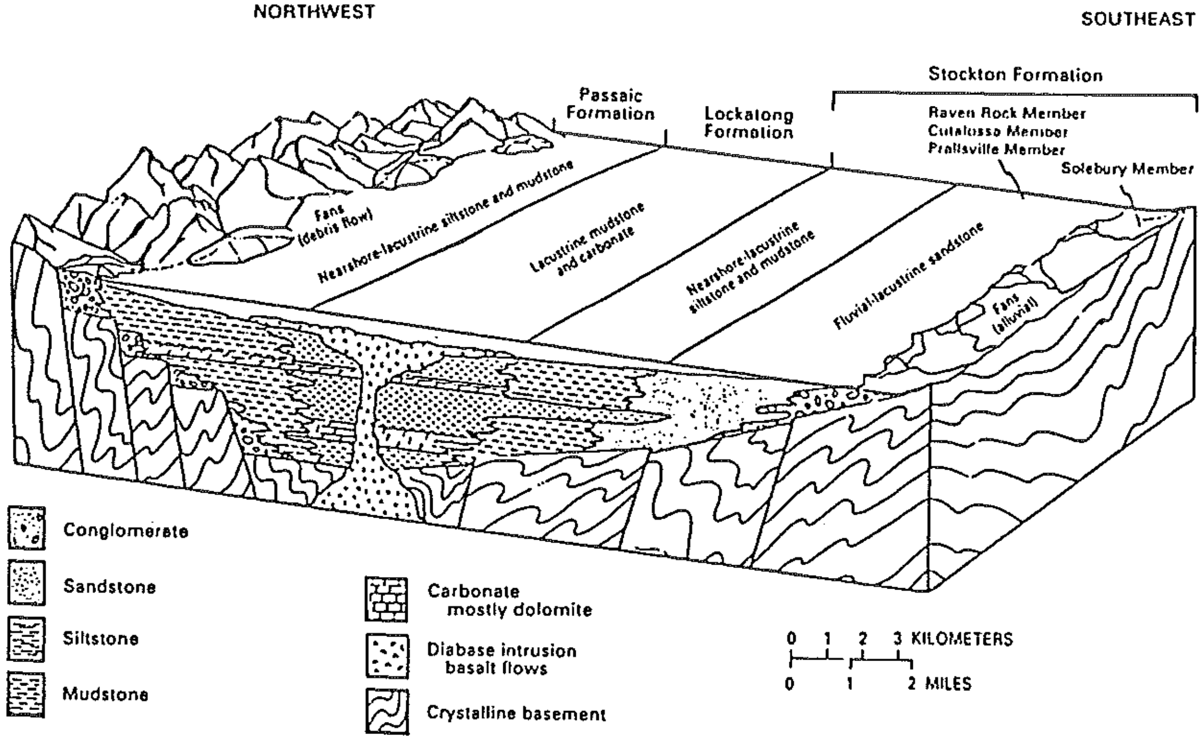


Figure 9 Primary distribution of facies in the Newark basin, inferred from the cross-section in Figure 8 (Modified from Turner-Peterson, 1980).

and into lacustrine basins during wet periods (Fig. 10). Conglomerates within these fans were laid down as upper to mid fan deposits by debris flows and by ephemeral (to possibly) perennial braided streams, whereas the finer sands and silts were deposited on the fan toe and valley floor by sheetflow and by ephemeral braided streams. A compatible interpretation of fan development in the Newark Basin (based on bedding types, grain size, sorting and fabric) was made by Arguden and Rodolfo (1986), who identified: debris flow deposits (matrix-supported conglomerates), streamflood deposits (clast-supported conglomerates), braided stream strata (coarse pebbly sandstones), sheetflood strata (medium-to-fine-grained sandstones), and a waning-flood facies (thin mudstones) that in places is overlain by caliches, documenting a pause in deposition. Our field trip will examine many of the same features described by these workers.

Fluvial-Lacustrine Shoreline Facies. Although fluvial strata comprise a major lithosome in most basins, interpretation of these deposits vary widely, a condition often related to the fact that synrift fluvial systems are large and complex, while exposures are small and scattered. Further complications arise because within closed-basin systems, stream channels are incised into older lacustrine strata as the lake, or depositional baselevel, drops, and are flooded as the lake level rises. In the Dead Sea, for example, where sealevel drops about 0.5 m per year, Holocene streams have been incised over 200 m into older Pleistocene lacustrine deposits, thereby laterally juxtaposing Recent sands against strata dated at 60,000 years (Manspeizer, 1985). A case in point follows.

Poorly sorted pebbly sandstones, with paleocurrent trends aligned parallel to the basin axis, make up an important part of the basin fill. Hubert et al. (1978a,b) and Weddle and Hubert (1983), for example, mapped a meander belt in the New Haven Arkose (Hartford Basin) that was fed by a braided stream complex draining alluvial fans to the east. Sandstone bodies, up to 3 m thick, composed of sandstone layers separated by scour surfaces and composed of plane beds, planar crossbed sets and ripples with apparent random orientation, were interpreted as linguloid bars in a braided channel. On the other hand, fining-upward sequences with scoured basal contacts and lag gravels that are overlain by trough crossbed sets, and followed upward by plane beds and rippled sandstones that give way upward into siltstones, were identified as meandering deposits by these same workers (see Fig. 7.1.3 in field guidebook section).

Coarsening-upward sequences, however, may be interpreted as deltaic, or the result of 'seaward' progradation of shoreline features, e.g. beach or barrier island. Turner-Peterson (1980) and Turner-Peterson and Smoot (1985) have interpreted the upper part of the Stockton Formation as a sequence of low-angle deltas, dominated by wave reworking, with only a minor fluvial component (Fig. 9). Allen (1979), on the other hand, has interpreted similar sequence as the product of braided and meandering streams, issuing from alluvial fans to the south. We shall examine several coarsening-upward sections in the field.

Volcanic-Lacustrine Facies. A distinctively different episode of basin filling, starting near the beginning of the Jurassic Period, is marked by

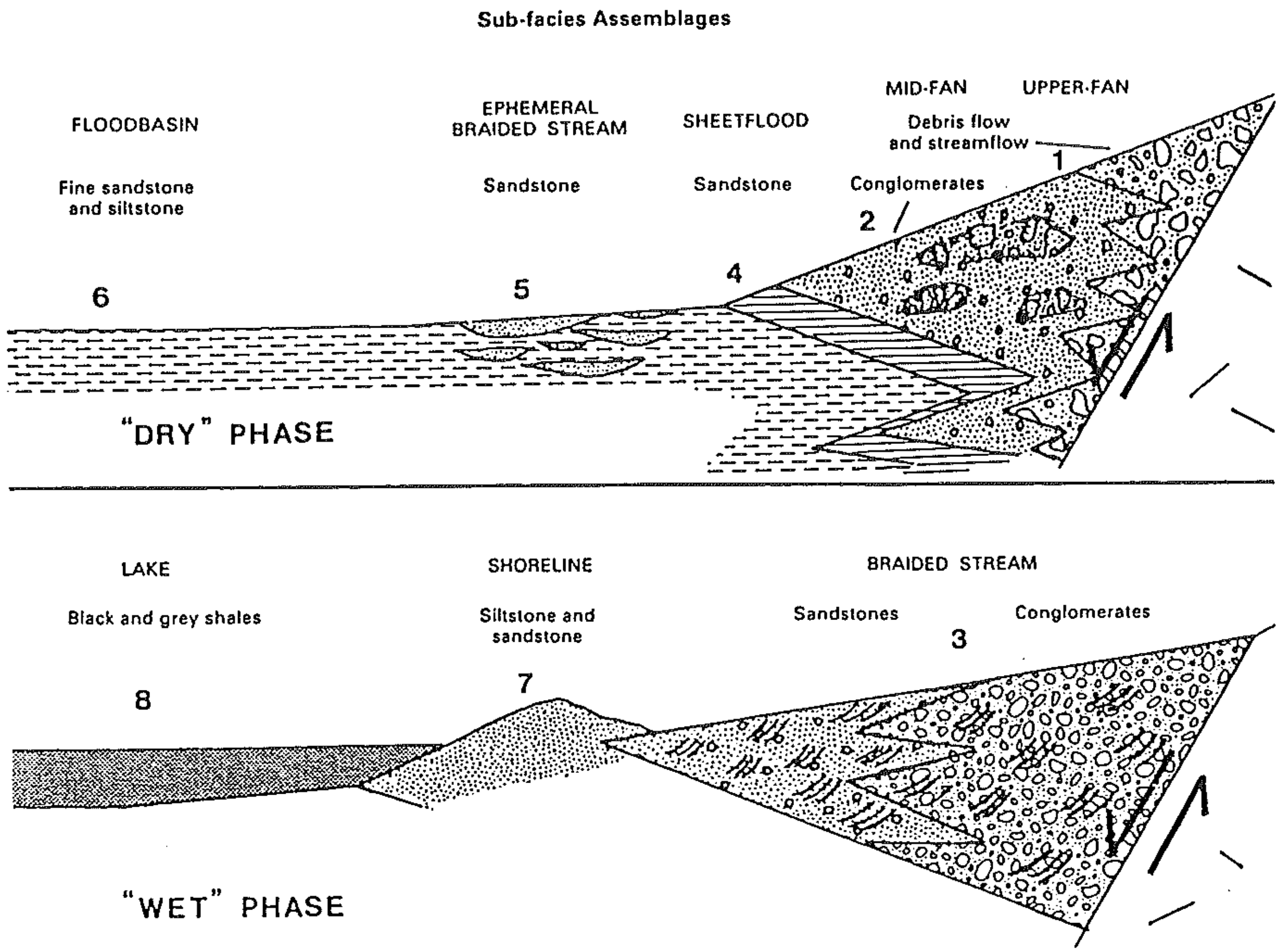


Figure 10 Lateral relations of depositional environments of the two climate types affecting deposition during the development of the Mesozoic rift basins (Modified from LeTourneau, 1985).

interbeds of tholeiitic lavas and lacustrine strata. From the volcanic province of Florida to the Fundy Basin and east to North Africa, plutons intruded the crust, feeding tholeiitic lavas that flowed along the valley floor where they impounded surface runoff forming lacustrine basins (Manspeizer, 1980).

Wherever the Liassic strata are exposed, as in the Gettysburg, Newark and Hartford Basins, they are marked by a sequence of 3-13 tholeiitic lava flows that are intercalated with 25-300 m of lake beds. Volcanism began at least 20 million years after the onset of sedimentation in the Triassic, and lasted for a brief interval of about 500,000 years during the Hettangian Epoch. It is unlikely that the Triassic border faults served as conduits for the egress of the lava, as paleoflow structures in the lavas of the Newark Basin (Manspeizer, 1980) show that the feeder dikes lie to the southwest near the narrow neck and close to the basin axis (Fig. 2). Geologic maps of these basins show that very few lava flows occur along the border faults. Geochemical studies by Puffer et al. (1981) demonstrate that the lavas of the Newark and Hartford Basins are rock and time-rock correlatives, and thus were emplaced by a synchronous crustal event. In addition to the diabase sills and dikes, e.g. the Palisades Sill of the Newark Basin and the intrusives of the Gettysburg Basin, another set of dikes appear to cut across the synrift strata. These are most densely concentrated in the Carolinas, where, on a Bullard reconstruction, they form a radial pattern centered over a possible hotspot hear the Bahamas (May, 1971).

It is curious that while the dynamics of these basins must have changed with the onset of igneous activity in the Early Jurassic, the fundamental geometry for major basin-segments remained largely unaffected. This conclusion is drawn from lithofacies studies (e.g. Manspeizer, 1980; McGowan, 1981; Hubert et al., 1978a,b; LeTourneau, 1985) showing that, within the same basin, Triassic and Jurrasic alluvial fan and fluvial-deltaic sedimentation are time-transgressive across essentially isochronous volcanic horizons.

Evaporite Facies. As clastic sedimentation was occurring in eastern North America and in the High Atlas of Morocco, the western margin of the transgressing Tethys Sea was the site of a complex network of northeast-trending basins that were dominated by carbonates and sulphates in the Alps and southern Spain, and by halite with minor amounts of anhydrite and dolomite in Algeria, Tunisia and the

Aquitane (Jansa et al., 1980; Manspeizer, 1981). Further west, massive deposition of halite-rich evaporites dominated the initial opening of the Atlantic rift from the Scotian-Essaouira Basin to eastern Newfoundland (Fig. 1; Manspeizer, 1981; Holser et al., in press). To the south, extreme aridity resulted in the precipitation of potash minerals in the Doukkala, Berrichid and Khemisset Basins of the Moroccan Meseta, at DSDP site 546, off the coast of Morocco, and in the Orpheus Graben on the Scotian Shelf (Holser et al., in press). Normal marine conditions characterize most of the thick salt deposits, except for the early evaporites, which had a significant contribution from continental sources. Palynologic studies (Cousminer and Steinkraus, in print) also document that these Late Triassic shallow marine seas extended as far south as Georges Bank, off the coast of New England (Manspeizer and Cousminer, 1988).

COST G-2 Well

The COST G-2 well is the deepest and the single most important stratigraphic test well on the U.S. margin (Fig. 4 and 7). It was drilled to a depth of 6,667 m, where it penetrated a thick upper Triassic section of dolomite with limestone and anhydrite, bottoming in Upper Triassic salt. Palynomorphs, in both cores and cuttings (Cousminer, 1983), indicate that the post-rift unconformity occurs within an attenuated Liassic section (less than 330 m thick) of carbonates and evaporites, thereby establishing the time of uplift, erosion, and subsidence in the outboard basins (Manspeizer, in press). Significantly, the well records the occurrence of Triassic marine phytoplankton at 4,441 m, indicating that marine conditions were present on Georges Bank as early as the Late Triassic. A very different interpretation of the COST G-2 well, based on seismic stratigraphy, is offered by Poag (1982).

TECTONICS AND SEDIMENTATION

Timing of Tectonic Events

Determining the age of deformation is important because it bears on various models concerning the origin of these basins. Early rifting with syntectonic deposition and recurrent activity along the border fault may be inferred (for the Newark Basin) from a thick (up to 8 km) succession of time-transgressive border conglomerates, coupled with syndepositional thickening observed on proprietary seismic lines (Katz et al., in press), syndepositional folds and unconformities (Ratcliffe, 1980), and growth faults (Van Houten, 1980, citing Cloos and Pettijohn, 1973).

Another clue to deformation comes from the occurrence of phacolith-like structures in the Newark and Hartford Basins (Manspeizer, in press). There they occur along the hinge of synforms and consist of tholeiitic intrusives, dated at 200 Ma that are cross-cut by an apparently undeformed younger set of dikes that are dated as Early Middle Jurassic, suggesting that folding and intrusion are essentially co-eval. This conclusion is compatible with data from the COST G-2 well, showing that the post-rift or breakup unconformity (marking the end of the rifting stage) occurred within the Lias or Early Jurassic (Manspeizer, 1988 and in press).

Provenance and Sediment Dispersal

The long lives of these basins, with accumulated sediment thicknesses on the order of 6-8 km, provokes the following questions: what is the nature of the source rock(s) and what can we infer about its uplift history? Combined paleocurrent and petrographic studies typically show that rift basins are filled with sediment that is derived locally from marginal highlands. Because many Mesozoic rift basins lie along sutured contacts, wherein contrasting basement terranes have been juxtaposed against each other, hinged and border fault margins typically produce different suites of detrital sediment. In part these clastic suites reflect different weathering domains brought out by differences in elevation, slope, climate, vegetation, etc. along each margin.

The broad west-facing deeply-eroded paleoslope of the Newark Basin, broken by growth faults and the source of much feldspathic detritus, consisted of soda-rich crystalline bedrock of the Piedmont uplands (Van Houten, 1980). On the other hand, unroofing of the crystalline highlands to the northwest supplied, first Paleozoic and later Proterozoic, detritus across a steep paleoslope to the basin on the east. Paleocurrent data for fluvial sandstones in the Newark Basin, together with petrographic modal analyses of 120 Triassic and Jurassic sandstones throughout the basin, show that all the sandstones were locally derived from highlands adjacent to the present margins of the basin (Oshcudlak and Hubert, in press). It is thus unlikely that these strata were connected in a broad terrane rift valley with those of the Hartford and Pomperaug Basins. Based on petrographic analyses of Late Triassic sandstones in the Newark, Hartford, and Deerfield basins, Weddle and Hubert (1983), also concluded that each basin was separated from each other by basement highs and was filled from local sources.

We may speculate from the vast thickness of the synrift deposits, the local nature of their sources, and the recurrent stratigraphic record of conglomerates along the border fault, that the margins were episodically rejuvenated by tectonic activity. Evidence for syntectonic deposition, recorded in the field by localized unconformities and deformed clasts, will be examined in the field.

Tectonic and Climatic Considerations

The stratigraphic record is largely influenced by tectonism and climate. We have seen that in the Triassic, tectonism plays the dominant role because it controls the history of the plate, distribution and shape of the basins, and the relief and uplift history of the margins. Climate, on the other hand, commonly influences the type of weathering products found in the source, the type of detritus carried into the basin, and the chemical and biological processes in the basin. In this section we shall examine briefly the interrelationships of these factors.

The final phases of the Variscan-Alleghanian orogeny produced a broadly convex, elongate landmass that extended from about paleoaltitude $80^{o}S$ to $70^{o}N$. It had an area of about 184 x 106 km^2, a broad central arch standing about 1.7 km high and 720 km across, and an average elevation of more than 1,300 m above the early Mesozoic sea level (Hay et al., 1981).

During the Middle Triassic, the future U.S. and Canadian margin lay between the equator and $20^{o}N$ latitude, thereby encompassing climates ranging from equatorial rain forest to tropical savannah (Manspeizer, 1981). As the plate migrated farther north (Morgan, 1981), trangressing about 10^{o} latitude between the Late Triassic to the Middle Jurassic, the northern rift basins (e.g. those on the Scotian Shelf and Moroccan Meseta) were subjected to increasing aridity as they moved under the subtropical high pressure cell. At the same time, the southern basins (e.g., the Dan River and Richmond) stayed humid, as they remained under the continuing influence of the equatorial low pressure system. These long-term climatic trends are documented in the stratigraphic record; see Hubert and Mertz (1980) for the Fundy Basin, Jansa and Wade (1975) for the Scotian Shelf, and Reinemund (1955) and Olsen et al. (1981) for the southern basins. Savannah-like climates with strongly alternating wet and dry seasons, as first suggested by Krynine (1935), seem to have prevailed over much of the region.

Continents, by virtue of their size, low heat capacity, and topography, strongly modify their climates. Large continents (e.g., Pangaea in the Triassic, and Eurasia today) have a more profound effect on their climates than do smaller landmasses (e.g., North America). Monsoon circulation, the seasonal alternation of wind direction, must have been an important factor in sedimentation. The Pangaean winters, for example, probably were dominated by a subtropical high-pressure cell that carried in warm dry air from aloft to the northwest. On the other hand, Pangaean summers probably were dominated by the Intertropical Convergence Zone (or equatorial low pressure systems) that brought in warm moist air from the Tethys Seaway to the east (Manspeizer, 1981).

As this moist air was uplifted almost 2 km over the mountainous terranes of high-standing Pangaea, it must have cooled adiabatically, owing to a decrease in atmospheric pressure. Condensation followed, yielding rainwater for streams that flowed away from the axis of doming, and for streams filling high-altitude deep water lakes within the rift valley complex. These high altitude Mesozoic lakes were thermally stratified and had a well-developed thermocline separating oxygenated surface water from anoxic bottom waters, as interpreted from the stratigraphic record by Olsen (1980) and Manspeizer and Olsen (1981). However, where air descended into low-altitude rift basins along the proto-Atlantic axis (COST G-2 well), it would have warmed adiabatically due to increasing atmospheric pressure. This would have lowered base level, aiding the formation of evaporite minerals wherever marginal marine embayments or fresh water lakes existed (see Manspeizer, 1981). The occurrence of Late Triassic salt flats in the offshore basins that were synchronous with moderately deep lakes, coal swamps, and zeolite-rich playas most eloquently shows the impact of rift topography and elevation on Early Mesozoic lithofacies (Fig. 1).

The Triassic-Liassic column records pervasive cyclical patterns of wetting and drying, wherein lakes expanded and contracted, with periodicities of 21,000, 42,000, 100,000, and 400,000 years. These intervals agree with the Milankovitch astronomical theory of climates (see Van Houten, 1969; Olsen, 1980; Olsen, in press, a,b).

Basin Geometry: Impact on Lithofacies

The depositional history of these basins is strongly influenced by their half-graben geometry. Where rainfall was excessive, as along the faulted margins of these basins, it fed high-discharge ephemeral streams, flowing down steep inclines feeding coarse-grained, poorly sorted braided streams that prograded alluvial fans and/or fan deltas into moderately deep water basins. On the other hand, perennial streams, because of their much larger drainage basins developed along the more gently sloping axial or flexed margins of the basins, transported large quantities of fine clastics from more deeply weathered and dissected uplands.

Closed-Basin Systems. A closed-basin paleohydrology has been postulated (Smoot, 1985) for all Triassic-Liassic basins during some part of their histories. Such basins have no drainage outlet, so that all surface drainage leaves primarily by evaporation. In this system, the distribution of coarse clastics in the depositional basin is largely influenced by the basin size, the amount of inflow, and the presence or absence of a lake. The closed-basin model has major implications for synrift sedimentation, since much of it was influenced by the frequent rise and fall of lake levels. In this construct, the lake level is the base level for the system.

Smoot (1985) notes, for example, that during flood events, small lakes will rise, rapidly transgressing drowned river mouths further upstream, and thereby inhibiting the development of steep delta forsets and the formation of coarsening-upward bottomset-forset-topset sequences. However, classical Gilbert-type deltas probably will develop where rivers enter deep lakes in large basins. Smoot further notes, that the frequent changes in lake level will cause small delta forsets to develop over the top sets of lower lake level stands, and cause shorelines of high lake level stands to be dissected.

Thermal History

The thermal history of these basins is poorly documented. Oshchudlak and Hubert (in press), cite Robbins (1983), who inferred from vitrinite reflectance values in Lockatong mudstones (not in proximity to intrusives) that the maximum regional paleotemperature was about 80-100^{o}C. Schamel and Hubbard (1985) report that the majority of vitrinite reflectance values are between 1.5-3.0, and that the clay minerals are dominated by well crystallized illite and chlorite. Nevertheless, highly thermally altered rocks and relatively unaltered rocks are present without transition in the basins. This suggests either discrete episodes of exceptionally high heat flow during development of the Mesozoic Basins rather than monotonically decreasing temperature after rifting (Hatcher and Romankiw, 1985; Pratt et al., 1985), or the existence of vigorous hydrothermal circulation within the older units in the basin (Katz, in press). Additionally, differences in thermal maturity of rocks that are the same age, but deposited in different basins, may indicate that thermal events were localized, rather than synchronous along the entire margin (Hatcher and Romankiw, 1985). Modelling studies of the Newark Basin suggest that periods of elevated temperatures and heat flow are correlative with initial rifting and subsequent magmatic activity (Katz et al., in press; Kotra et al., 1985), and that the maximum burial depth of the youngest rocks in the basin was only about 1.5-2.0 km (Pratt et al., 1985).

Synrift models

In this section, we shall examine 3 phases of basin filling: large-scale evaporite deposition, fluvial lacustrine basin filling, and late-phase volcanism.

Triassic-Liassic synrift sedimentation is notable for the paradoxical occurrence of evaporites offshore and moderately deepwater-stratified lacustrine and fluvial deposits onshore; it records a complex balance between climatic, tectonic, and hydrologic conditions. The distribution of these lithofacies (Fig. 1) provides a clue to this environmental puzzle (see discussion under 'environmental considerations'). Within the late Triassic-Liassic lithosome, three major evaporite groups are recognized: (1) anhydrite and dolomite, (2) halite, and (3) K+ and Mg++ salts. Those dominated by anhydrite, reported from Georges Bank (Arthur, 1982), probably formed in a sabkha similar to that of the Persian Gulf. The linear distribution of salt deposits in the Baltimore Canyon Trough and in the Georges Bank and Scotian basins indicate that the evaporites formed in near sea-level rift basins where marine, or perhaps fresh water was restricted (Rona, 1982) and evaporated through adiabatic warming of descending air. The marine evaporite sequence typically stops at the halite stage (Eugster, 1982), but the final concentration products in the early Mesozoic seas were enriched in K+ and Mg++ salts, as recorded in the Triassic-Liassic deposits of Morocco (Salvan, 1972), which bordered the North American Atlantic margin at this time. Because these K+ and Mg++ salts are exceedingly hygroscopic minerals (that can be preserved only under extreme aridity), they probably formed in playas where subsiding brine-filled basins filled with halite (Eugster, 1982).

Almost all of the Triassic-Jurassic lacustrine rocks show a pattern of recurrent lithologies, constituting simple and complex cycles (Van Houten, 1969; Olsen, 1980; and Manspeizer and Olsen, 1981). The Lockatong Formation, the best known of these ancient lake deposits, serves to illustrate Triassic lacustrine conditions. As a huge lacustrine lens of Late Triassic age (Carnian Stage), the formation extended from central New Jersey to Pennsylvania. These strata are about 1,150 m thick and commonly arranged in regressive and transgressive cycles that resulted from the expansion and contraction of this ancient lake. According to Van Houten (1962, 1969), their deposition was controlled by the 21,000-year precession cycle.

Compound cycles have been reported by Olsen (in press, a) with peaks near 42,000, 100,000, and 400,000 years. These large periodic changes are related to cycles in the seasonal variation of sunlight, as prescribed by the Milankovich astronomical theory of climate change (Olsen, in press, a,b).

As climates became more arid toward the close of the late Carnian, Lake Lockatong gradually gave way to well-oxygenated mudflats with fringing alluvial fans and braided streams and playas from which glauberite, gypsum, and caliche were precipitated (Van Houten, 1969). Elsewhere in eastern North America (Hubert et al., 1978a,b), the late Triassic Norian Stage was marked by the most extensive development of red beds. It was a time of semi-arid, low-lying source terranes and broad mudflats that sloped gently eastward to a marine sabkha (COST G-2 well). The sabkha was transgressed by Tethyan waters that drained through the Gibraltar Fracture Zone (see Manspeizer et al., 1978; Jansa et al., 1980) or perhaps by Arctic waters flowing through the North Atlantic rift zone of East Greenland (see Manspeizer, in press).

The final phase of basin filling onshore took place in the Early Jurassic. It is clearly marked by multiple sequences of tholeiitic lava flows and with interbeds of moderately deepwater lacustrine deposits (as in North America and Morocco) or with carbonates and evaporite deposits (as in Morocco). Most importantly, volcanism began almost simultaneously over an area perhaps 2,500 km long and 1,000 km wide, and about 20 m.y. after the onset of Triassic sedimentation. Recurrent movement along transforms and continental fractures substantially altered the tectonic framework of these basins, so that lacustrine fan deltas were deposited during the Early Jurassic in the same basins where fluvial red beds had accumulated in the latest Triassic. Early Jurassic basins were asymmetric and marked by active listric fault margins and high rates of sedimentation. Subsidence in the Early Jurassic appears to have been almost 2 to 3 times greater during the volcanic-lacustrine phase of synrift sedimentation than during the non-volcanic Triassic red bed phase, estimated by Van Houten (1969) at about 0.3

m/1,000 years.

Elsewhere, as in the North Sea, the Early Jurassic was a time of continued rifting, eustatic lowering of sea level, and volcanism. On the newly forming margins of North America and northwest Africa, the final phases of synrift deposition were marked in the offshore by regional uplift, marine regression, tilting, extensive erosion, and perhaps volcanism. In the Middle Jurassic, as the locus of igneous activity moved further offshore to the axis of future spreading, the inner margin cooled and subsided, ushering in the beginning of postrift sedimentation.

THERMAL-MECHANICAL MODELS

Thermal-mechanical models of the evolution of continental rifts and passive margins generally assume that rift basins form in response to tensional stresses oriented normal to the rift zone. Two stages of development characterize the evolution of continental rift systems. The first stage, rifting, involves massive extension of the continental lithosphere, similar to that presently observed in East Africa and the Basin and Range province of western North America. This rifting stage persists for a finite length of time and is often associated with abundant volcanism. The drifting stage follows rifting and is characterized by passive margin development as sea-floor spreading begins and the rifted continental margins drift apart. The sedimentary record contained wihin the onshore Mesozoic basins and along the continental margin of North America suggest that the rift to drift transition is associated with a fundamental change in the thermal structure and mechanical behavior of the lithosphere. For example, seismic studies in the southern Appalachians show that a distinct change in basin infill in the Georgia basin is correlative with the shift from rifting to drifting (McBride et al., 1987). In addition, the post-rift unconformity along the continental margin formed because of regional uplift along the future margins of North America and northwest Africa at the end of the rifting phase. Thermal-mechanical models describing the evolution of rift zones attempt to explain the many features observed in the geologic record. Most models, however, have dealt with either the early rifting phase or the later drifting phase and a single theory that accounts for the entire sequence of events has not yet been developed.

Geodynamic models of rifted margins can be categorized into two main types. Pure shear models are characterized by low-angle listric normal faults that sole into a basal detachment at mid-crustal levels, with extension in the lower crust and upper mantle accomodated by ductile deformation (Fig. 11a). Alternatively, simple shear models assume that the detachment surface penetrates the entire crust and upper mantle and then terminates in a region of lithospheric thinning (Fig. 11b). These two classes of models predict different subsidence rates, heat flow, maximum temperatures, and gravity and magnetic signatures across rifted margins. Because of the strong dependence of lithosphere rheology on temperature, most modelling schemes calculate the decay of an initial thermal perturbation associated with rifting, and then determine the changes in the mechanical behavior of the lithosphere through time as a function of the transient temperature structure. These models have been proposed to describe the structure and evolution of the entire Atlantic margin. Individual rift basins may form by similar processes, but it is more likely that their presence reflects the interaction of large-scale stress systems with the pre-existing regional heterogeneity of the continental lithosphere.

Pure Shear Models

The basic pure shear model attempts to explain passive margin and adjacent rift basin subsidence as a consequence of lithospheric necking. Tensional stresses oriented normal to the trend of the future rift are imposed at a location distant from the site of rifting in this model. McKenzie (1978) quantified this process for a simplified case (Fig. 12) where the continental lithosphere begins with an initial thickness (Ƶ), and the crust has an initial thickness of (z). The lithosphere (crust and upper mantle) is then instantaneously extended by a constant factor (β). At the time of extension, both the crust and lithosphere are thinned from their original thicknesses to new thicknesses of z/β and Ƶ$/\beta$ respectively. The base of the lithosphere is taken as an isotherm (eg. 1333^{o}C), so as extension occurs the vertical temperature gradient (dT/dz; where dT = change in temperature and dz = change in depth) and heat flow ($q=k[dT/dz]$; where q = heat flow and k = thermal conductivity) increase by a factor of β. The asthenosphere, an isothermal, inviscid fluid slightly less dense than the lower lithosphere, passively rises to fill vacated space at the base of the lithosphere. According to McKenzie

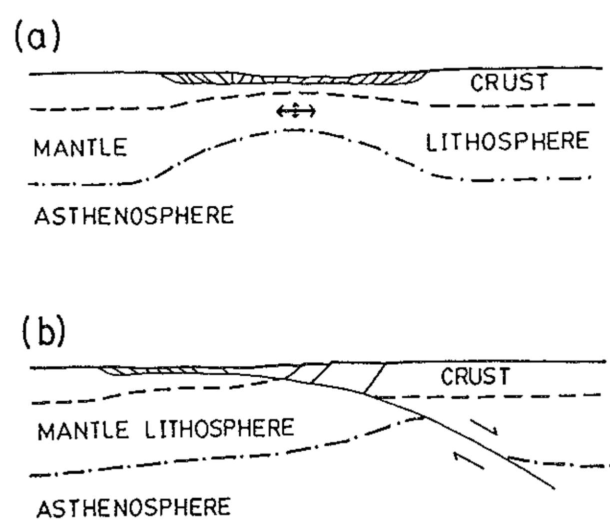


Figure 11 Basic models of lithospheric extension applied to continental rifting. (a) Pure shear model. (b) Simple shear model (Modified from Buck et al., 1988).

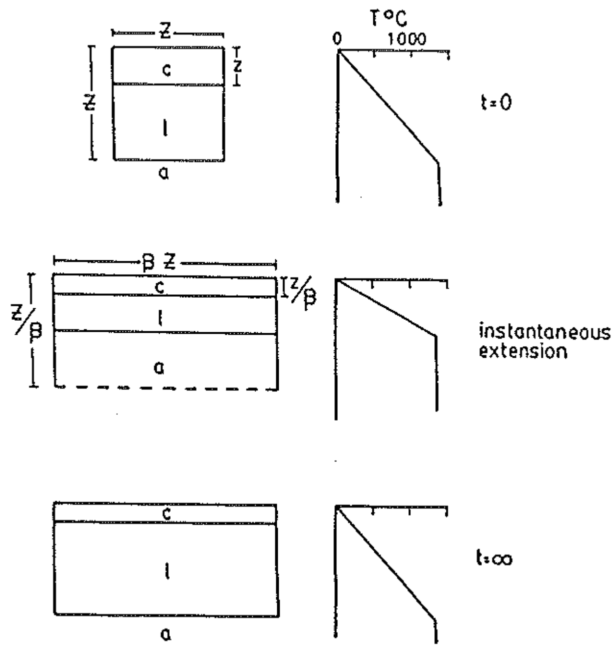


Figure 12 Simplified model of pure shear extension through a necking process. Progression through time from a stable initial condition (t=0), through an instantaneous extensional event, to a final stable condition (t=∞). Lithospheric sections shown on the left, change in geotherm with extension and subsequent cooling are shown on the right (Modified from McKenzie, 1978).

(1978), subsidence following extension occurs in two parts. The initial rapid subsidence results from a redistribution of mass across the extended area. Crustal densities (rho_c=2.8 g/cm^3) are much less than those of both the asthenosphere (rho_a=3.2 g/cm^3) and lithosphere (rho_l=3.3 g/cm^3) so that crustal thinning results in subsidence unless the crust was initially very thin with respect to the lithosphere (McKenzie, 1978). Airy (local) isostasy is assumed, meaning that the rifted margin, or a single basin formed by this mechanism, subsides independently of the surrounding region. As the extended region cools, the lithosphere returns to its initial thickness during the second, thermal subsidence phase. Inclusion of thermal contraction in the necking process accounts for the similarity between passive margin subsidence and subsidence of cooling oceanic lithosphere moving away from mid-ocean ridges (Parsons and Sclater, 1977). This model has been applied to the Atlantic margin of North America with reasonable success. The edge of the zone of lithospheric extension is generally modelled as spatially coincident with the hinge zone (Fig. 4). In this way, the extreme drift phase subsidence outboard of the hinge zone can be modeled as a result of thermal contraction. Each inividual Mesozoic basin has not been shown to overlie an area of thinned mantle lithosphere. Therefore, although it is possible that the pure shear process controlled development of each individual basin, it is also plausible that the entire passive margin formed by this mechanism and the onshore basins are associated, but not located directly above the region of maximum extension.

Structural geologists working on Cenozoic extension features in the Basin and Range Province of western North America provided the initial geologic mechanisms for the pure shear model that were later simplified and quantified by McKenzie (1978). Horst and graben systems observed at the surface were believed to reflect brittle failure in the upper crust, while extension at lower levels was accomplished by plastic deformation (Stewart, 1971). The structural model is supported by studies of the rheology of continental lithosphere in extensional environments that suggest that the crust is composed of an upper brittle layer, underlain by a lower ductile layer that rests on a brittle upper mantle (Smith and Bruhn, 1984). The pure shear model is consistent with these geological observations because it assumes that detachments sole at mid-crustal levels. In addition, the pure shear model accurately predicts the observed heat flow in the Basin and Range Province (Lachenbruch and Sass, 1978). Almost every aspect of the basic pure shear model has been examined and the effect of variation of serveral basic parameters is examined below.

Variations in Extension Factor (β). DeCharpal et al. (1978) suggested that brittle failure and extension in the upper crust may represent much less extension than is experienced by the ductile lower crust and lithosphere. Increasing the amount of extension with depth may result in higher heat flow during rifting than is predicted by a uniform extension model. Synrift uplift that is often observed in continental rift zones (Falvey, 1974) can also be produced by models that vary β with depth (Royden and Keen, 1980). If synrift uplift occurs and the uplifted area experiences significant erosion, that region will undergo greater subsidence upon cooling than is predicted by the basic pure shear model (Sleep, 1971). This mechanism will produce broad shallow basins in regions landward of the continental margin hingeline where only minor crustal attenuation is experienced.

β values in models are often also varied laterally across an extended region to reproduce observed geologic and geophysical characteristics (eg. Barton and Wood, 1984). Such variations in β factor across the Atlantic Margin are reasonable because there is no evidence that an abrupt change in continental crustal thickness exists (Hutchinson et al., 1986).

Finite Rifting Times. Rifting along the margin of North America was not instantaneous, but lasted for approximately 35-55 m.y. Jarvis and McKenzie (1980) considered the effects of rifting over a finite, rather than instaneous, period of time. The results of their modelling show that an instantaneous rifting model accurately predicts the subsidence history of a passive margin when rifting lasts for less than 20 m.y. For rifting over a longer period of time, a significant amount of syn-rift conductive cooling occurs. This increases early, syn-rift subsidence of the lithosphere, and decreases subsidence that occurs during the drifting phase, although the total combined subsidence remains the same. Because a significant portion of the sedimentary record in the onshore Mesozoic basins was deposited during the rifting phase, thermal-mechanical models which examine the interaction of tectonics and sedimentation should account for rifting over a finite period of time. This factor has profound control over the timing of basin subsidence and the resulting creation of space available for sediment accomodation.

Flexural Compensation. Because the lithosphere is not completely decoupled along the margins of the Mesozoic basins, an assumption of Airy isostatic compensation for each individual basin is probably not valid. Initial stages of rifting are, however, associated with extensive faulting and lithospheric heating. An Airy isostatic model is generally believed to be applicable at that stage because of partial decoupling down to the base of the elastic lithosphere, which is often correlated with a specific isotherm (eg. 450°C, Watts, 1982). As the rifted margin cools, its elastic thickness increases and it becomes stronger. Drift phase lithosphere is generally modelled as an elastic plate (Beaumont et al., 1982; Watts, 1982).

Flexural compensation accounts for the internal strength of the lithosphere, so that loads (whether internal, as in the case of McKenzie's model, or external, as in the case of thick sedimentary packages in basins) affect regions adjacent to the site of loading. A particular load will inluence an increasingly large region as the strength of the lithosphere increases. The presence of the Fall Line, the boundary between subsided and uplifted lithosphere, located between the Coastal Plain and Piedmont Province along the Atlantic margin, and its behavior through time is consistent with elastic flexural models (Watts and Ryan, 1976).

Two-Dimensional Heat Flow. Lateral heat conduction results in significantly more rapid cooling and subsidence of rifted margins and basins than does one-dimensional (vertical) heat transport (Huntoon and Furlong, 1987; Cochran, 1983). The inclusion of lateral heat transport in models increases the predicted amount of syn-rift subsidence by about 25% for a 20 m.y. long rifting event (Cochran, 1983). Two-dimensional heat flow also leads to thermal expansion and uplift in regions inboard of attenuated lithosphere (Cochran, 1983). This effect is particularly important to consider for the onshore Mesozoic basins because they are probably located adjacent to, rather than directly above, the zone of major lithopheric thinning associated with Mesozoic rifting. Heat reaching these basins prior to volcanic activity was conducted or advected in from the east. Asthenosphere upwelling across the extended region, particularly to the east of the onshore basins, introduces large lateral temperature gradients in the mantle. This process may induce convection, particularly in narrow rift zones (Mutter et al., 1988; Buck, 1986). that would result in increased uplift adjacent to the rift zones (Buck, 1986). The general lack of volcanics along the Atlantic margin suggests however, that convection and partial melting were not important (Mutter et al., 1988).

Sedimentation. The Mesozoic rift basins contain great volumes of sediment, so that the effects of sediment loading and compaction on subsidence of the lithosphere should be considered. Sediment densities (rho_{sed}=2.5 g/cm^3) are only slightly lower than those of continental crust, and Watts (1982) demonstrated that the post-rift subsidence of passive margins is controlled mainly by sediment loading and thermal contraction. The onshore basins, while receiving a large sediment load, probably overlie an area of relatively thick crust and lithosphere. Based on flexural interpretations of the mechanical behavior of this type of lithosphere, the effects of a large sediment package were probably minimal. This has not yet been fully investigated however.

Backstripping methods may be used to differentiate between tectonic and sediment loading subsidence, and quantitatively determine the effect of sedimentation in a basin. Under the assumption of Airy isostasy this process is fairly straightforward (eg. Steckler and Watts, 1978). Consideration of flexural compensation complicates the process and generally increases estimates of tectonic subsidence in the center of a basin (Huntoon and Furlong, 1987). Sediment compaction has a minor effect on the subsidence of passive margins and basins (Watts, 1982; Keen and Keen, 1973).

Simple Shear Models

Simple shear models were also developed to describe continental extension in the Basin and Range Province. The occurrence of numerous fault blocks, low and high-angle planar or listric normal faults (Wright, 1976; Anderson, 1971) that sole into a subhorizontal detachment surface, and over 100% horizontal extension (Wernicke and Burchfiel, 1982; Wernicke, 1981) are all observed in that region (Fig. 13). Deep data obtained by seismic reflection studies suggest the normal faults sole into a master detachment that penetrates the deepest crust and lower lithosphere. Simple shear models also resolve many of the questions associated with emplacement of metamorphic core complexes (Wernicke 1983, 1985) and the geometry of Basin and Range listric normal faults (Spencer, 1984).

Seismic reflection profiles south of Long Island appear to indicate the presence of low-angle faults penetrating nearly the entire crustal thickness (Phinney, 1986; Hutchinson et al., 1986). While this certainly supports the hypothesis that brittle failure can penetrate the entire crust, the imaged region represents only a small portion of the Atlantic Margin (Fig. 14; note the Block Island Fault is a re-activated Paleozoic thrust). There are perhaps several detachments across the margin, and a single master decollement may be lacking. Pre-existing shear zones, sutures, and thrusts along the Atlantic Margin could all be re-activated during extension to produce many low-angle normal faults.

The hanging walls of the Newark and Gettysburg basins are characterized by a positive gravity anomaly; a situation that is predicted by an asymmetric shear model incorporating the effects of flexural

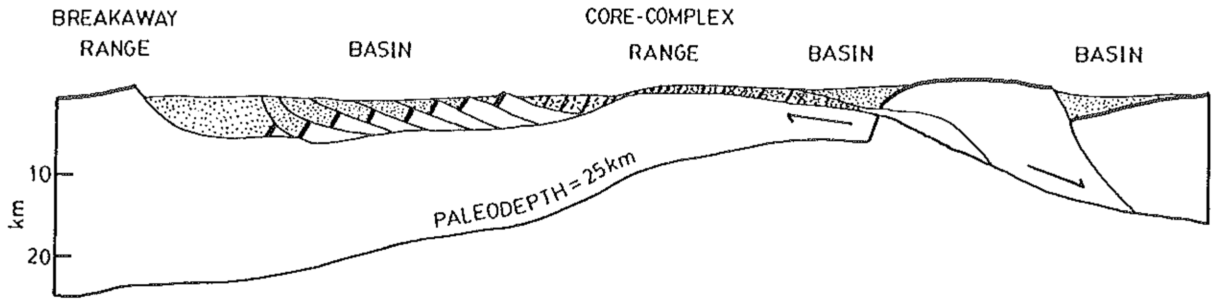


Figure 13 Schematic cross-section demonstrating the compatibility of the simple shear model with geologic features observed in the Basin and Range province (Modified from Wernicke, 1985).

compensation (Bell et al., 1988). The model of Bell et al. (1988) does not, however, entirely account for the positive anomaly adjacent to the basin and the lack of a greater negative anomaly below the basin. They have suggested the presence of a 2 km thick diabase layer emplaced along the basal detachment (reactivated thrust fault) that also forms the northwest margin of the basin. This addition improves the model fit, but it is possible that the inclusion of the effect of Proterozoic crustal thinning below the basin might also be important. The relative positions of the Newark and Gettysburg basins and the highly extended lithosphere below the continental margin are adequately accounted for by a simple shear model (Bell et al., 1988).

One of the major objections to the simple shear model is its inability to account for initially high heat flow in the basins far removed from the region of lithosphere thinning/asthenosphere upwelling, and subsequent volcanic activity in the same area. The simple shear model does, however, allow for nonuniform extension of the crust and lithosphere without space problems.

Studies of the Colorado Plateau/Basin and Range transition indicate that the geophysical response (heat flow, uplift) of the lithosphere to simple shearing is detectable for only about 15 m.y. after extension ceases (Furlong and Londe, 1986). In contrast, effects of pure shearing are felt over a longer period of time and if both mechanisms are in operation, pure shear masks the effect of simple shear over the long term (Furlong and Londe, 1986). This emphasizes the importance of utilizing indicators of paleotemperatures and paleotopography in reconstructing the evolution of the Atlantic margin, and it also emphasizes the importance of study of the Newark and other Mesozoic basin sediments as a source of this type of information.

Hybrid Models

Modelling studies and structural investigation of different basins along the Atlantic margin, of the entire margin, and of other similar tectonic settings worldwide support various aspects of both the simple and pure shear models. A single model is probably not applicable to every basin along the Atlantic margin just as a single model cannot be applied to every margin. The seismic evidence along the central Appalachians margin suggests the operation of both pure shear and simple shear mechansisms. Based on the geophysical observations, and the limitations of the various models presented to replicate this data, neither model is preferred at this time. Kusznir et al. (1987) presented a coupled simple shear/pure shear model of continental extension that may help to resolve several problems. This model utilizes the abundance of low-angle detachments documented from many passive margins to separate regions extending by simple shear (above) from regions extending by pure shear (below). According to their model, the lithosphere experiences brittle failure along low-angle faults that sole within or at the base of the crust (simple shear), and behaves plastically below a detachment level (pure shear) (Fig. 15) Their numerical model assumes instantaneous extension, two-dimensional heat flow, and flexural compensation. Initial deformation results in crustal thinning and perturbation of the geotherm. Basin

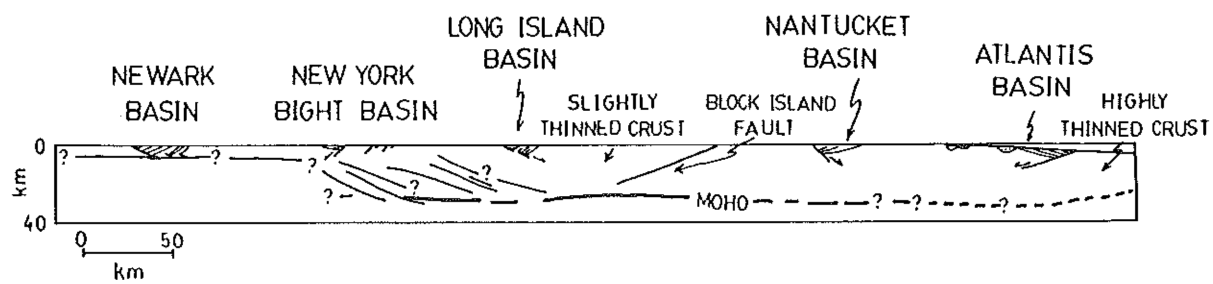


Figure 14 Cross-section across the central Atlantic margin based on seismic reflection data. Evidence suggesting both simple shear and pure shear mechanisms is present (Modified from Hutchinson and Klitgord, in press, a).

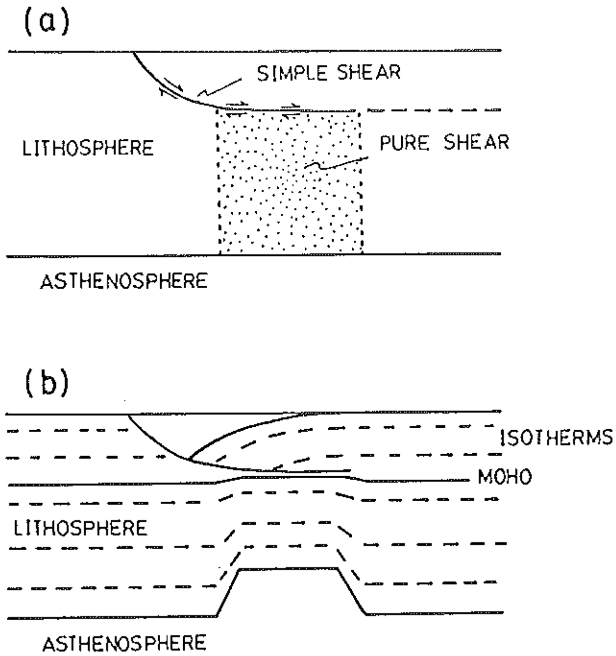


Figure 15 Hybrid simple shear/pure shear model. (a) Extension within the brittle portion of the crust is accomplished by simple shear, while extension in the mantle lithosphere is taken up by pure shear (dotted area). (b) Lithosphere geometry and thermal structure resulting from an extensional event incorporating both simple and pure shear (Modified from Kusznir et al., 1987).

formation is then a two-stage process. First the hanging wall block collapses, and second the redistribution of mass is accounted for flexural-isostatically. Including lithospheric flexure during rifting minimizes deformation of low-angle faults and detachments. This model is potentially useful in describing the evolution of the Mesozoic basins because it explains the apparent superposition of asymmetric rift basins that formed by brittle failure upon a regional thermally subsiding passive margin that can be accurately described by pure shear models.

The behavior of the continental margin is very complex, but the inclusion of several physical parameters may help to explain observed features. For example, inclusion of the effects of non-unifom extension (different extension values at different depths), increasing flexural rigidity through time, lateral heat flow during rifting, secondary convection, and sedimentary loading may all result in uplift along the hingeline during late rifting (Hutchinson and Klitgord, in press, b). This is an important effect because it would result in the creation of the post-rift unconformity, the most obvious sedimentary and seismic marker along the Atlantic margin.

ACTIVE VS. PASSIVE RIFTING

One of the major questions associated with extensional models is the nature of the forces that control extension of the continental lithosphere. In general, continental rifts follow the grain of previously deformed belts (eg. Lake and Karner, 1987; Sykes, 1978), as is the case in the Appalachians. The coincidence of location may indicate that previously deformed regions are structurally weaker than the surrounding areas (King, 1970), or that rifting is a direct consequence of earlier structural deformation (McConnel, 1978). A genetic relationship between mountain building and rifting, if one exists, suggests that the processes controlling rift development actually act at the site of rifting. A particular concept, that of active rifting (Fig. 16), proposes that rifts are tensional features which form at the surface because of regional doming or arching (Burke, 1976). Domal upwarping causes tension in the upper curst which leads to brittle failure and extension (Cloos, 1939). Surface magmatic activity is then facilitated by the fracturing of the crust. The driving force for uplift is either a sustained input of heat from the lower mantle (eg. Burke and Whiteman, 1973), or regional compression. A significant amount of erosion (on the order of kilometers) must occur during uplift to permit regional subsidence below sea level, once the heat source or compressional force is removed (Sleep, 1971).

Alternatively, passive rifting models assume that subsidence preceeds uplift and that plate separation in a horizontal plane causes rifting (Rosendahl, 1987; de Charpal et al., 1978). The force driving plate break-up may be either graitational instability of an orogenic belt located at the site of the future rift, or stresses transferred to the plates from a distant location (Manspeizer and Gates, 1988). Disagreement over the validity and relative merit of the active and passive rifting models is presently widespread, and one model may not be appropriate for all cases of continental rifting. The eastern margin of North America is however an available location for testing the applicability of these hypotheses to at least one case.

Igneous activity, including Mid-Triassic alkaline intrusions in Morroco and Algeria, may be cited as evidence of a pre-rift thermal event, possibly related to hot spot activity (Manspeizer, 1981). These intrusions do not extend over the entire length of the Appalachians however, and the sedimentary record must be consulted elsewhere for evidence of active rifting. Modelling studies of the Appalachian Orogeny

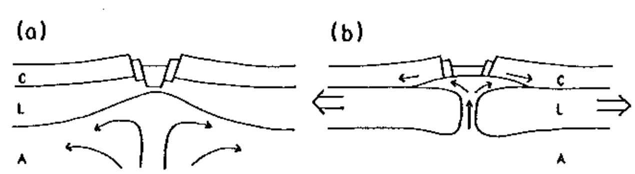


Figure 16 Schematic examples of active and passive rifting. (a) Asthenospheric upwelling causes uplift and extension of the rift zone. (b) Horizontal separation of lithospheric plates causes the lithosphere to thin, allowing the asthenosphere to passively move upward into available space. Uplift in (b) is subsequent to initial extension (Modified from Turcotte and Emerman, 1983).

(Beaumont et al., 1988) suggest that the maximum orogenic loads on the lithosphere were emplaced above and slightly to the east of the present trend of the onshore Mesozoic basins. If doming did occur prior to rifting, as suggested by the active rifting hypothesis, it would have affected a region coincident with the Appalachian orogen because Mesozoic extension was most pronounced in that area. Mesozoic extension was, however preceded by several m.y. of erosion so that the topographic relief produced by Paleozoic compression was somewhat reduced but not completely removed. Because erosion rates vary with topographic relief such that the rate of erosion decreases as the mountain belt is denuded (Ruxton and McDougall, 1967), it is expected that Early to Middle Triassic sediments deposited to the west of the onshore Mesozoic basins in the Appalachian foreland basin would exhibit an upward coarsening, or a decrease in the rate of upward fining if the source area was rejuvenated by doming. Hay et al. (1981) argues for the presence of an uplifted area about 1.7 km high and 720 km across centered over the future continent/ocean crustal boundary. Estimates of the size and position of this arch are based on returning sediments shed during the Triassic (including the abundant red beds of North America) to their source areas. In addition, Triassic basins in New Jersey and Pennsylvania were primarily filled from the east: sediments adjacent to the western border fault were locally derived (VanHouten, 1969). This evidence suggests the presence of a drainage divide to the east and this has been interpreted as evidence of pre-rift doming (Judson, 1975). The Hartford Basin similarly shows a dominant westward transport direction although strata now dip to the east, towards the source area. Interpretation of the sedimentary record in a search for evidence relating to the mechanism responsible for the onset of continental rifting is complicated by consideration of the effects of changes in climate or source lithology which can be responsible for changes in clast size through time, independent of variations in source area relief. The major argument against passive rifting is the lack of early stage volcanics.

Erosion of the Appalachian orogen prior to the onset of Mesozoic rifting is well documented in the geologic record, but details concerning changes in erosion rates through time and timing of drainage pattern reorganization are not well constrained.

Shear or Transtensional Models

Shear or transtensional models are sometimes consiered a subset of the passive rifting group. Extensive data from the East African Rift demonstrate that the long axes of half-grabens and border faults are terminated by strike-slip faults that follow pre-existing structural grain (Rosendahl, 1987). These data are interpreted as evidence that the border faults develop between shear zones in response to regional strike-slip motion. Similarly, there is evidence that Mesozoic rifting in eastern North America and northern

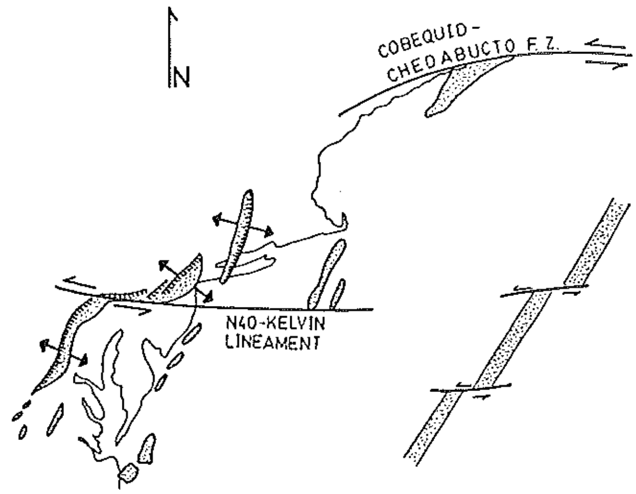


Figure 17 Sinistral shear couple associated with motion on the Cobequid-Chedabucto Fracture Zone and the N40-Kelvin Lineament as a possible driving force for extension of the Mesozoic basins (Modified from Manspeizer and Cousminer, 1988).

Africa began under the influence of an east-west trending, left-lateral shear couple (Manspeizer and Cousminer, 1988). In North America, however, it appears that the border faults follow pre-existing structural trends. The difference in geometry between East Africa and eastern North America is possibly due to differences in the angular relationship between the stress field responsible for transform motion and the orientation of pre-existing structural fabric. In the Appalachians, the sinistral shear stress field was oriented at approximately 45^{o} to the northeast trending axis of the Appalachian orogen (Fig. 17) (Manspeizer and Cousminer, 1988). Under these conditions, grabens may form when left-stepping, left-lateral faults, or right-stepping, right-lateral faults produce pull-apart basins (Aydin and Nur, 1982). After strike-slip motion initiates basin formation, normal faults will surround the entire basin (Freund, 1982). While some basins do appear to have formed by the pull-apart process (eg. Fundy Basin), others, including the Newark Basin, are characterized by normal faulting along re-activated Paleozoic structures with only a slight strike-slip component (Ratcliffe and Burton, 1985). These half-graben structures form in the orientation shown in Figure 17 under the influence of a left-lateral shear couple (Manspeizer and Cousminer, 1988). Extension normal to the orientation of the border faults is the dominant failure pattern observed in the Mesozoic basins, and is adequately explained by regional shearing. The thermal-mechanical models discussed above can be reasonably applied to most of the Mesozoic basins, even if they formed under the influence of regional shear, because relative extension is the dominant failure pattern. Three-dimensional effects become increasingly important to consider in a transtensional environment because of the rotational stresses imposed by shearing.

SITE DESCRIPTIONS

The field sites for this workshop and trip have been chosen to illustrate the interplay between tectonics and sedimentation in three types of Appalachian sedimentary basins: foreland, cratonic rift, and passive margin. Because much of the record of this interplay is contained within the basin fill, it is necessary to characterize the strata in some detail, working from the sedimentary facies through inferred processes to depositional environments and their evolution though time. The outcome of this exercise should be inferences about the tectonic influences on the deposits, such as the magnitudes, rates, and timing of creation of accomodation space, the elevation, rock types, and relief of the source terrain, and the structural evolution of the basin. The latter includes such items as structural lows controlling major fluvial systems, and tectonic oversteepening leading to gravity-driven processes. The outcome also should be inferences about the sedimentologic influence on the tectonics, such as creation of steady-state mountains by erosion, and isostatic compensation. This feedback loop is the lesser explored and therefore the more pregnant with possibilities. For example, it may well be that climate, acting through denudation processes, can control such tectonic processes as thrust advance in fold and thrust belts.

The organization of the field trip is from the oldest and most interior basin, the Paleozoic foreland, to intermediate, the eastern Mesozoic rifts, to youngest and most exterior, the present Atlantic passive margin. This is dictated not by any subtle logic, but by the structure of the Appalachians themselves. Figure 1 of the Introduction illustrates the chronology and path of the traverse.

DAYS 1 AND 2

Day 1 is given over to assembling at the Greater Pittsburgh Airport and driving to Penn State University. If you arrive early, a trip into Pittsburgh, PA is well worth the cost of an airport shuttle bus. It is a picturesque city sitting on incised strath terraces at the confluence of the Allegheny and Monongahela Rivers. Fort Pitt, a historical landmark on the banks of the Monongahela, provides a glimpse of the 18th Century role the region played during the French and Indian War. Excellent museums, such as the Carnegie Museum of Natural History, also are within walking distance from downtown. Day 2 consists of a workshop at Penn State University.

DAY 3

SITE 1: FOLD AND THRUST BELT STRUCTURE, STRATIGRAPHY, AND GEOMORPHOLOGY AT SKYTOP, PA

Rudy Slingerland

LOCATION

This site is 14 km west of State College, PA at Skytop, where US Route 322 climbs over Bald Eagle Mountain (Fig. 3.1.1). The parking area on the northwest side of the road is private property and must be used with discretion.

SIGNIFICANCE

Here can be seen the influence of thrust belt structure and foreland stratigraphy on regional geomorphology. Differential erosion has etched out the less resistant strata to highlight the geomorphic consequences of plunging folds, thrust ramps, and

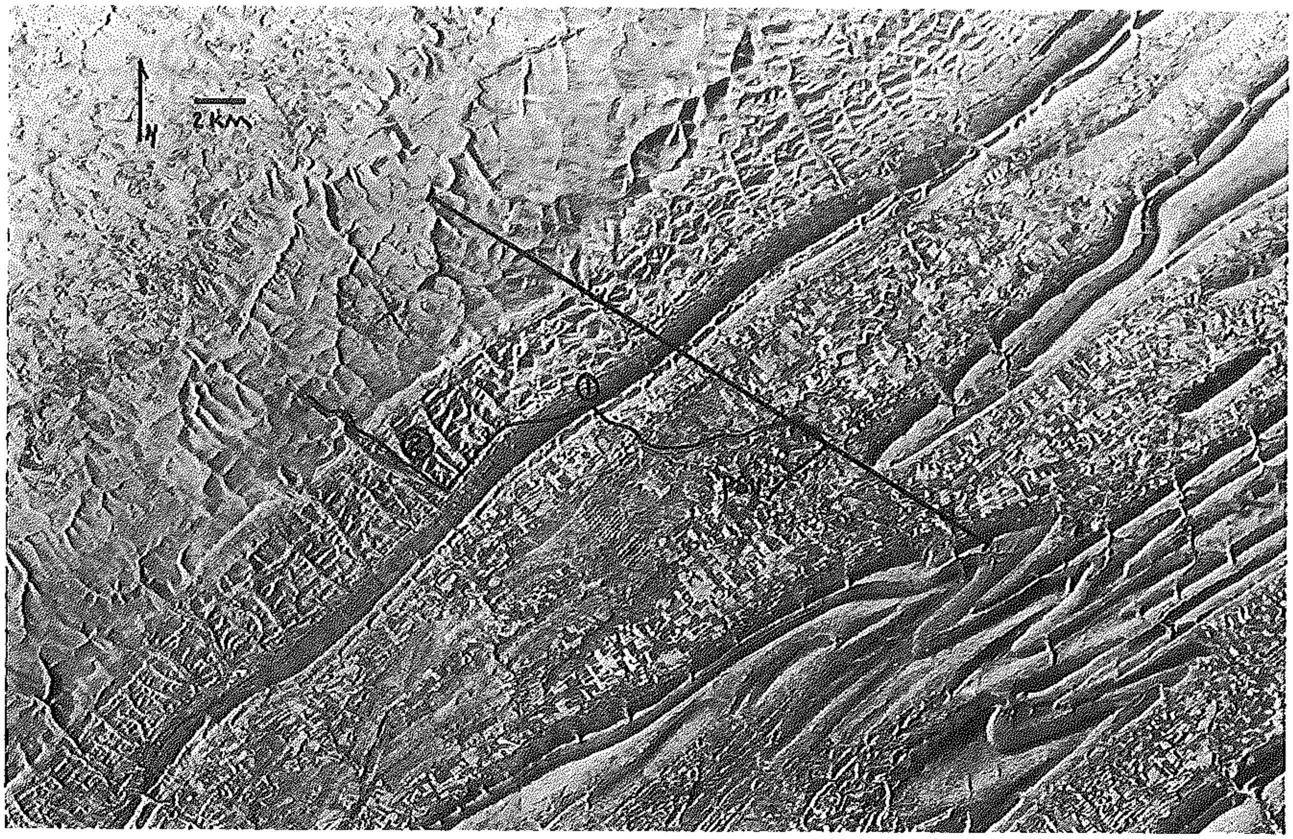


FIGURE 3.1.1 Synthetic aperture radar image of State College, PA, region showing Day 3 stop locations and line of section of Fig. 3.1.2.

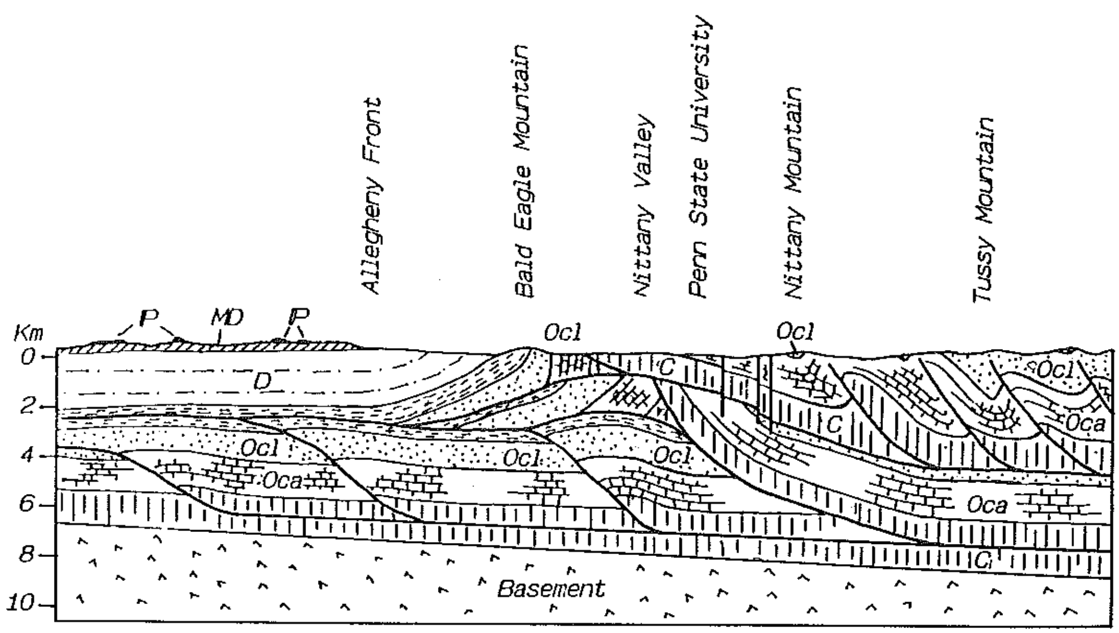


FIGURE 3.1.2 Geologic cross-section along line of section in Fig. 3.1.1 from Northwest (left edge of section) to Southeast (right edge of section). No vertical exaggeration (modified from Berg et al., 1980).

subtle lithologic variation. The folds and faults were formed during the late Paleozoic Alleghanian orogeny, the terminal compressional event of the Appalachian Orogen.

SITE DESCRIPTION

The viewer stands at 456 m elevation on Bald Eagle Mountain and looks northwestward across Bald Eagle Valley towards the Allegheny Escarpment rising to 660 m. Bald Eagle Mountain is the northwestern boundary of the ridge and valley physiographic province, a region of spectacular plunging folds and blind thrusts in variably resistant strata (Fig. 3.1.1). The mountains are underlain by sandstones---usually the Ordovician Bald Eagle and Tuscarora Formations---and the valleys are underlain by lower Paleozoic carbonates and shales. Beyond the Allegheny Escarpment lies the Allegheny Plateau, a region of broad rolling uplands and steep-sided valleys developed upon more gently folded Devonian through Permian foreland strata.

The stratigraphy between Bald Eagle Mountain and the lip of the Escarpment consists of the complete Silurian to Mississippian succession in the area (Fig. 3.1.2). We are standing on vertical to slightly overturned, resistant, Lower Silurian Tuscarora Formation. The Upper Silurian and Lower Devonian carbonates and shales are covered by talus below us. Bald Eagle Creek flows almost exclusively on the Middle Devonian Hamilton Group shales. The Escarpment itself, starting in the low foothills immediately northwest of Route 220, is underlain by the Upper Devonian-Lower Mississippian Acadian clastic wedge. Beds there dip gently northwestward.

The structural relief across Bald Eagle Valley is a surprising 4.5 km. and arises from complicated decollement tectonics (Fig. 3.1.2). As a result of compression from the southeast during the Alleghanian orogeny, cover rocks of the foreland moved to the northwest over a basal decollement in Cambrian strata. In the general area of this stop, numerous splays (localized by regional facies changes?) have stacked at least three thrust slices one above the other, resulting in exposure of Cambrian rocks in Nittany Valley, and overturning of the Silurian strata to form Bald Eagle Mountain.

SITE 2: ACADIAN FORELAND BASIN DEPOSITS AT PORT MATILDA, PA

Rudy Slingerland

LOCATION

Site 2 consists of five outcrops on the westbound lane of Route 322, that are respectively, 17.1 km (10.6 miles), 22.2 km (13.8 miles), 23.3 km (14.5 miles), 24.4 km (15.2 miles), and 26.2 km (16.3 miles) west of the intersection of Atherton and College Ave. in State College, PA (Fig. 3.1.1).

SIGNIFICANCE

These five outcrops present an uncommonly complete view of Acadian foreland fill away from the proximal margin. As presented in Slingerland and Beaumont (this volume), this clastic wedge began forming in earliest Middle Devonian time in response to Acadian orogenesis, with deposition of the Hamilton Group and northwestward progradation of subaerial deposits into the Catskill Epeiric Sea. A latest Middle Devonian eustatic(?) sea level rise abruptly terminated this first phase. Increasingly higher rates of clastic sediment influx in earliest Late Devonian time renewed progradation, producing the Catskill regressive sequence. It is this latter and more voluminous sequence that is exposed at this site. Also of significance, this site is located on one the rare depocenters of what must have been a highly variable coast (Fig. 15, Slingerland and Beaumont, this volume; Fig. 3.2.1). The outcrops start in the middle Frasnian Brallier Fm. deposited by dilute (?) turbidity currents on the basin floor or slope rise, expose shallow shelf storm deposits (Loch Haven Fm.), coastal tidal deposits (Irish Valley Mbr., Catskill Fm.), lower alluvial plain fluvial deposits (Duncannon Mbr., Catskill Fm.), and end in more proximal sandy braided river deposits of the Pocono Formation (Burgoon Ss). Thus the sedimentary character, processes, and environments of a 600 m thick medial basin wedge can be observed here.

SITE DESCRIPTION

Outcrop 1: Basin Floor and Slope Rise Deposits of the Brallier Fm.

This stop is near the base of the Brallier Formation, immediately overlying the Harrell Shale of earliest Late Devonian age. Twenty kilometers to the northeast at Milesburg, the Harrell Shale consists of 15 m of a lower, black carbonaceous paper shale (the Burket Member) and 90 m of upper gray, fissile, somewhat unfossiliferous clay shale. The Brallier

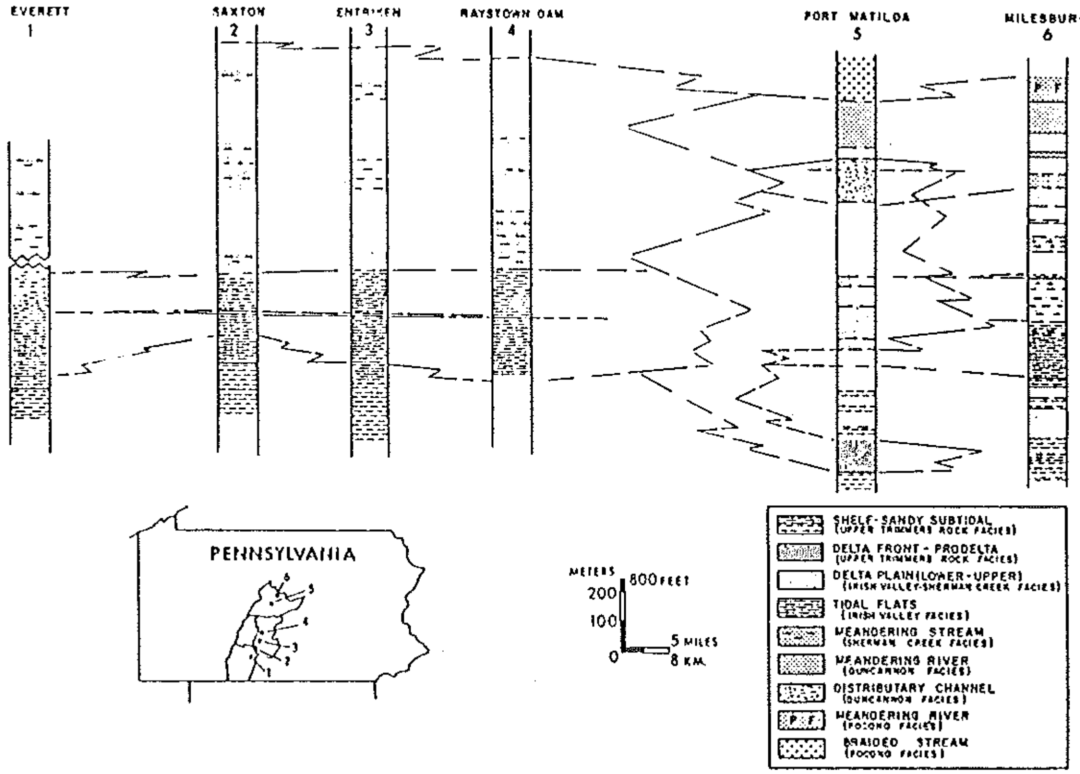


FIGURE 3.2.1 Correlation of Late Devonian sedimentary units and paleoenvironments along depositional strike (modified from Rahmanian, 1979).

there consists of 450 m of interbedded green micaceous silt shales and thin, evenly bedded very fine sandstones. This outcrop possesses a combination of these characteristics and so has somewhat arbitrarily been mapped as lower Brallier.

The accessible portion of the outcrop consists of medium dark gray, platy weathering, thickly laminated siltstones and silt shales, with occasional plant fragments, and sparse brachiopods, pelecypods, and burrows, and occasional thin beds of brownish-gray weathering very fine silty sandstone, internally completely ripple cross-laminated (Fig. 3.2.2). In the sandstone bed a half meter above the base of the outcrop, the cross-laminations are small scale troughs whose asymmetrical infilling indicates a unidirectional paleoflow to the southwest.

The mechanism of emplacement of these strata and their paleoenvironment are problematical. We think these were deposited on the basin floor or slope rise by dilute turbidity currents because they overlie the black shales of the Burket Mbr., and lack wave ripples and hummocky cross-strata---features diagnostic of storm deposition on a shelf.

Outcrop 2: Shallow Shelf Deposits of the Loch Haven Formation

The outcrop consists of silty shale and sandstone-shale facies of the upper Loch Haven (Trimmers Rock) Formation (Fig. 3.2.3). The silty shale facies consists of olive-gray, fossiliferous silty shale to siltstone with thin interspersed sandstone beds. The siltstones are horizontally laminated and the sandstones are small-scale trough cross-stratified and often have interference ripples on their upper surfaces. Slingerland and Loule (in press) suggest this facies was deposited on an open marine shelf, adjacent to shelf sand ridges.

The sandstone-shale facies consists of two sandstone subfacies interbedded with siltstone and silty shale units. The first subfacies is a chocolate brown, thickly bedded and sometimes massive, very

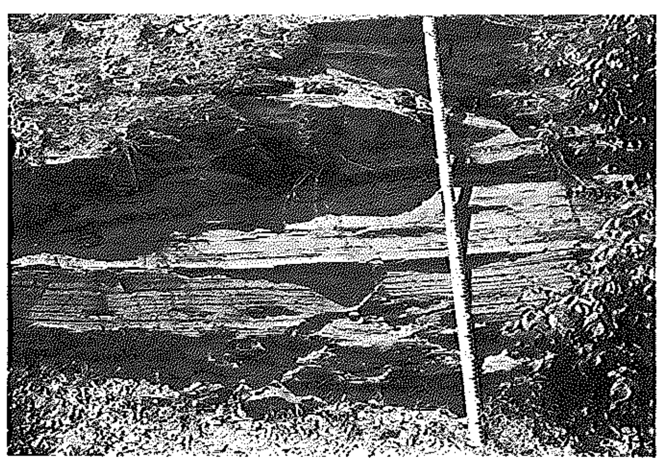


FIGURE 3.2.2 Silt shales and thin silt turbidites of the Brallier Fm., Day 3, Site 2, Outcrop 1.

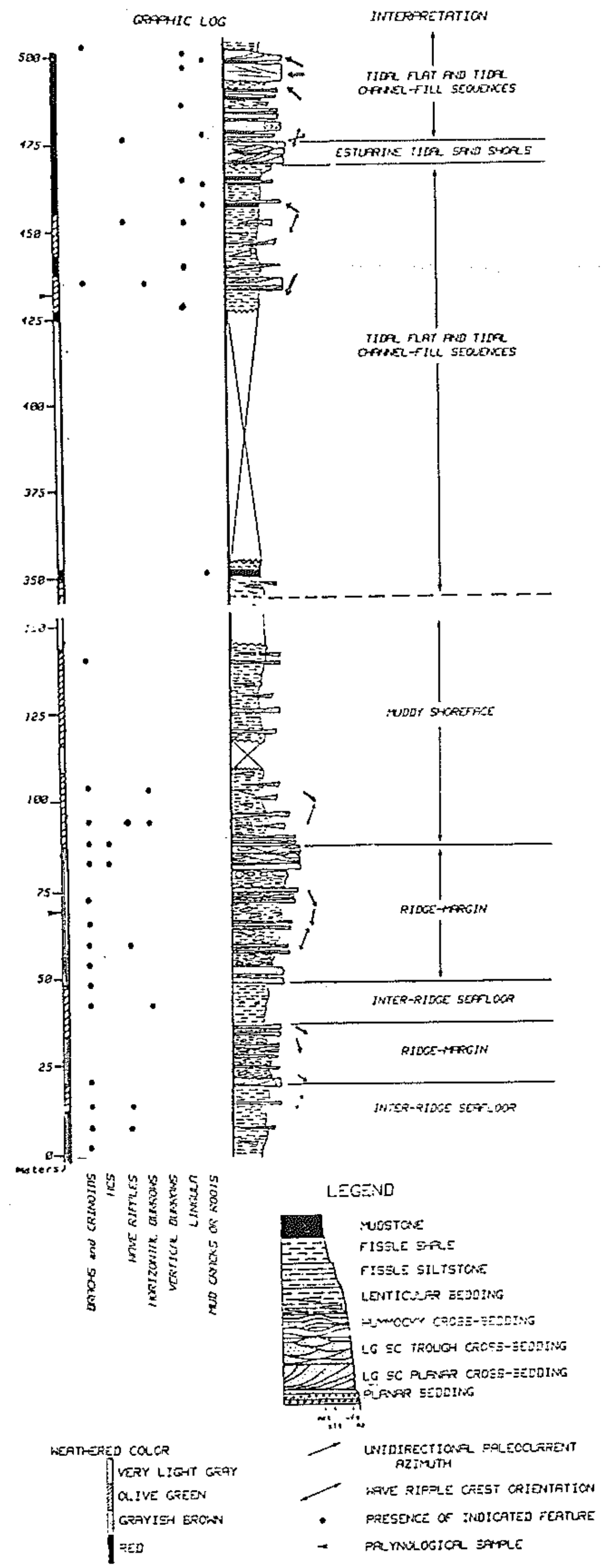


FIGURE 3.2.3 Stratigraphic column of the Lock Haven Fm., Day 3, Site 2, Outcrops 2 and 3(modified from Slingerland and Loule, in press).

fine-grained micaceous sandstone. It is often deformed into ball-and-pillow structures. The second subfacies is an olive-green, regularly bedded, fine-grained fossiliferous quartzose sandstone. Fossils include transported brachiopods, crinoid fragments, gastropods, and traces of Cruziana and Skolithos ichnofacies. Both subfacies commonly display a vertical sequence commencing with a sharp sole-marked base overlain by pockets of coquinite and quartz pebbles and hummocky cross-stratified or wave-ripple cross-laminated fine sandstone, and terminate with wave or asymmetrical ripple forms. By analogy with hummocky sequences reported elsewhere, Slingerland and Loule (in press) interpret these facies to have been deposited on the flanks of a shelf sand ridge complex by storm-driven geostrophic flows. Although both subfacies resulted from rapid deposition by waning flows during storms, we feel the chocolate brown unfossiliferous subfacies was emplaced directly by delta plumes during a rainy season, whereas the olive-green subfacies was emplaced in stages by a number of distinct downwelling flows.

Outcrop 3: Tide-dominated Deltaic Deposits of the Irish Valley Member, Catskill Formation

This revealing exposure occurs within the characteristic alternating red and green mudstones and sandstones of the Irish Valley Member of the Catskill Formation (see Williams et al., 1985a for a discussion on the origin of these colors). The green quartzitic sandstone-mudstone facies in the lower part of the outcrop and in three one-meter thick horizons above (Fig. 3.2.3 and 3.2.4), consists of a vertical alternation of thinly bedded olive-green mudstone and very fine-grained quartzitic sandstone and siltstone. Wavy and lenticular bedding are the dominant sedimentary structures; small brachiopods and fish bone fragments are common. Slingerland and Loule (in press), like Williams et al. (1985a) and Rahmanian (1979), interpret this facies to have been formed by short-lived marine transgressions over parts of the lower delta plain, probably because of variations in local sediment supply.

The red rocks can be divided into a red sandstone to mudstone facies sequence and a pink trough cross-stratified, medium-grained sandstone facies. The sandstone to mudstone sequence contains, from an erosive base upwards, a pale pink, large scale trough or planar cross-stratified fine- to medium-grained channel-filling sandstone; inclined heterolithic strata of red very fine-grained sandstone, siltstone, and mudstone arranged in thin interbeds with flaser and lenticular bedding and asymmetrical ripple forms; and red laminated siltstones and mudstones with mudcracks and root traces. The channel sands often contain a hash of brachiopod, bivalve, gastropod, and crinoid fragments, and the heterolithic strata may contain Lingula, whereas the upper units are devoid of body fossils. This sequence is interpreted as the product of tidal channels, possibly with some riverine component, migrating laterally through muddy tidal

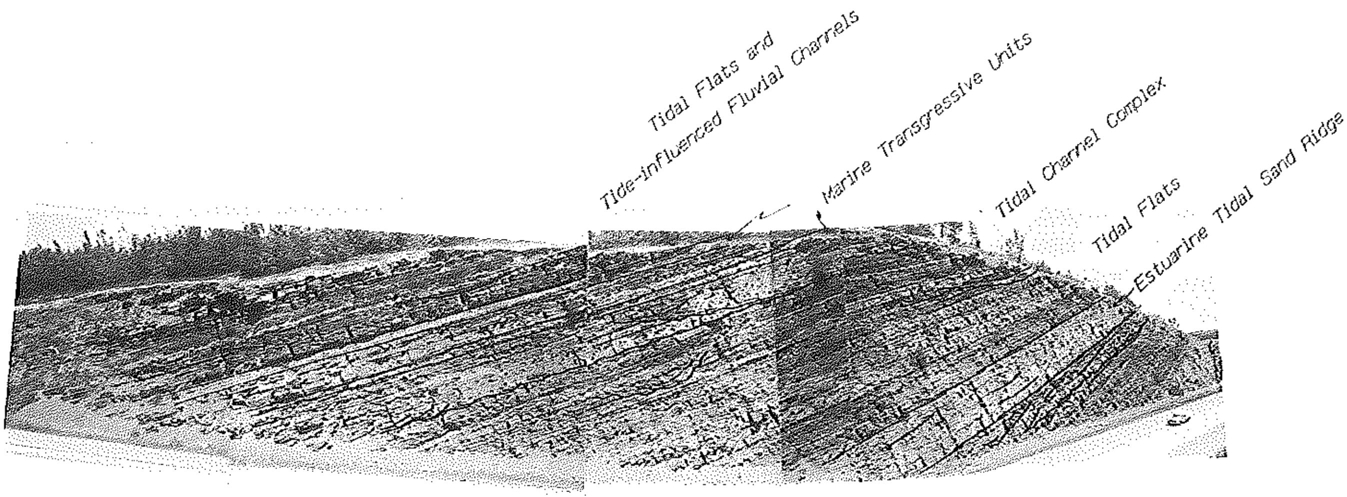


FIGURE 3.2.4 Strip photograph and interpretation of Day 3, Site 2, Outcrop 3.

flats. We feel the heterolithic facies is especially diagnostic of estuary mouths of modern tidally-influenced rivers, where alternating sand and mud is deposited in response to temporary tidal storage of water (Smith, 1985).

The pink trough cross-stratified, medium-grained sandstone facies at this site consists of one 8 m thick sandbody showing multi-directional paleoflows, internal shale drapes, marine body fossils, and quartz pebbles along bedding planes. The base and top are both gradational over a few decimeters into the heterolithic facies described above. Slingerland and Loule (in press) interpreted this facies as an estuarine shoal complex because of its gradational relationship with tidal flat facies, large scale bedforms possibly due to amplified tides, and lower density of burrows possibly due to decreased salinities. See Clifton (1982) for a modern counterpart from Willapa Bay, Washington.

In summary, the assemblage of facies at this and the previous outcrop is believed to represent various subenvironments of a tidally influenced delta (see Fig. 15, Slingerland and Beaumont, this volume). Like its modern counterparts, it probably consisted of an anastomosed distributary system entering a flared estuary. The estuary probably contained shore perpendicular estuarine sand shoals, and was bounded by vast muddy tidal flats along its margins. As noted in Slingerland and Beaumont (this volume), a monsoonal climate led to high river discharges for part of each year, such that plumes of sediment flushed through the estuary onto the shelf where the dominant southwestward shelf circulation created shelf sand ridge complexes downdrift. During the remainder of each year, tidal flows winnowed the estuarine sediment to produce the clean quartz estuarine sandbodies and adjacent mudflats. The paleoclimatology (Woodrow, 1985) and wind- and tide-driven circulation of the Catskill Sea as predicted by numerical modelling (Slingerland, 1986; Slingerland and Loule, in press; Ericksen et al., in press) are consistent with this interpretation (Fig. 3.2.5) as are the paleocurrent data (Fig. 3.2.3).

Outcrop 4: Fluvial Channel and Overbank Deposits of the Duncannon Member, Catskill Formation

We have moved approximately 345 m (1150 ft) upsection through the poorly exposed Sherman Creek Member, consisting of fining-upwards cycles of pink to gray-green fine cross-bedded sandstone, rippled siltstone, and brick red mudstones, the latter often containing root traces, paleosols, and caliche horizons. The inferred environments for these rocks are those associated with shallow, meandering rivers and adjacent flood basins, all on the distal portion of a broad coastal plain, immediately upriver from the estuary mouth.

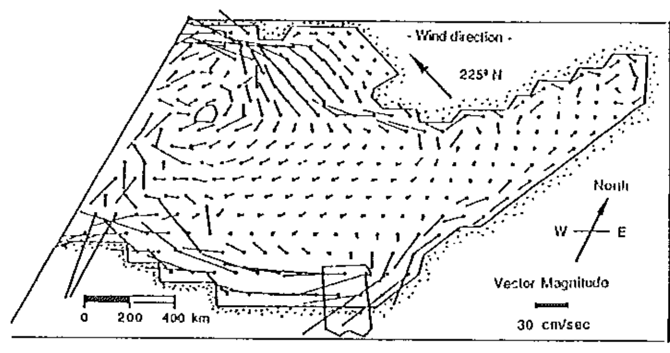


FIGURE 3.2.5 Computed residual circulation in the Devonian Catskill Sea subject to 20 knot southeast winds with an M_2 tide along the open boundary. The predicted flux along the Catskill Shelf in Pennsylvania is to paleowest (modified from Eriksen et al., in press).

The rocks exposed at Outcrop 4 (Fig. 3.2.6) also comprise fining-upwards cycles, but they differ from the subjacent strata in that each cycle is much thicker and contains relatively more sandstone and less red mudstone and shale. They also are thicker than cycles at the equivalent horizon to the north and south, leading Williams et al. (1985a) to infer that the main river system feeding this depocenter maintained a fixed longitudinal course as it crossed the Late Devonian basin in central Pennsylvania. The ultimate cause may have been a basement controlled low (Williams et al., 1985a) or it may have been structural control of the drainage net in the source region, such as in the Zagros Mountains of Iran today.

Each cycle starts at an erosional base, that is overlain by a gray-green, medium-grained, micaceous, large scale trough cross-stratified sandstone, commonly containing a meter basal concentration of conglomeratic sand. The clasts consist of intraformational carbonate (caliche?) concretions, mud chips, and carbonaceous wood and plant fragments, as well as extraformational scattered white quartz pebbles. These major erosive surfaces and associated conglomeratic sandstones also occur within sandstone bodies, dividing each into a number of "storeys". Measurements of trough and planar cross-stratification through all stories give a northwest transport direction (Fig. 3.2.6). The last storey of each cycle often, but not always, fines upwards and grades into a thin interval of red interbeds of siltstone, shale, and claystone containing many root traces and desiccation cracks. Calcareous paleosols also are sometimes preserved in this facies. Of particular interest at this outcrop is a 1.2 m thick green, burrowed, siltstone and silty shale directly overlying the paleosol of cycle 4. It contains calcareous brachiopods and pelecypods and is the last known marine bed of the Acadian clastic wedge at this site.

Although the general character of these cycles suggests deposition by rivers migrating laterally across an alluvial plain, the thick multistorey sandbodies and paucity of overbank muds indicates to us that the rivers were of low sinuosity and possibly braided, at least during the dry season. In this interpretation, the accretion bedding so obvious in the outcrop was formed by lateral migration of mid-channel pebbly sand bars as in the Brownstones of southwestern England (Allen, 1983). The occurrence of marine transgressive units would seem to indicate that the setting was still low on the plain. Alternatively, because the Devonian-Mississippian boundary is only 100 m above this section (Stolar, 1978), this could be the upper alluvial plain manifestation of the abrupt marine transgression at the end of the Late Devonian that deposited the Riceville Shale and Oswayo Formation (see Slingerland and Beaumont, this volume for details).

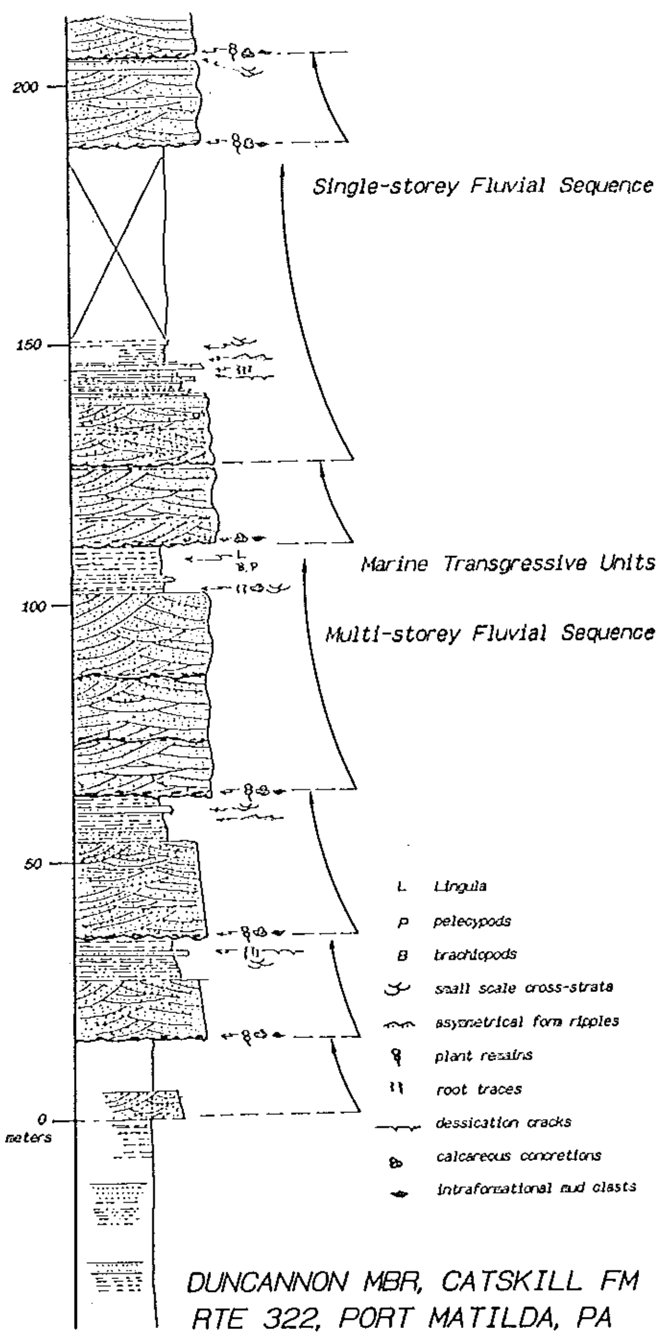


FIGURE 3.2.6 Stratigraphic column for Day 3, Site 2, Outcrop 4.

Outcrop 5: Braided Fluvial Channel Deposits of the Burgoon Sandstone

These strata were mapped by earlier workers as the middle sandstone member of the Pocono Sandstone. They consist of greenish-gray, medium- to coarse-grained, large scale cross-stratified sandstones organized in lozenges 10s of meters wide and up to 10 m thick (Figs. 3.2.7). Lozenge boundaries are defined by many overlapping, concave-upwards, erosive unconformities. If bedding planes are of zeroth order (terminology of Allen, 1983), then the boundary between lozenges is of third order. Within each lozenge are first order surfaces separating trough and planar cross-strata sets, and second order surfaces bounding accretionary sedimentation units. Throughout

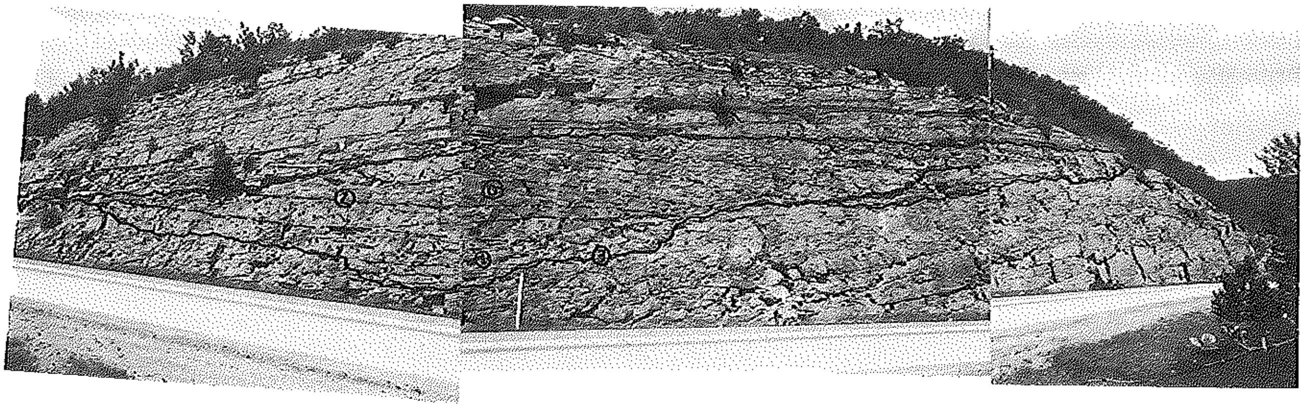


FIGURE 3.2.7. Strip photograph of the Burgoon Ss. at Day 3, Site 2, Outcrop 5, showing an erosively-based channel fill with a hierarchy of internal surfaces in the manner of Allen (1983).

the strata, but especially abundant immediately above the surfaces of higher order, are pebbly sandstones loaded with logs, stems, and comminuted plant debris. Clasts are both intraformational, consisting of gray shale chips and siderite concretions, and extraformational, consisting of quartzose pebbles. Gray silt shale and bone coal lenses occasionally occur above third order surfaces, one with stigmarian root casts preserved, and uncommonly large pyrite nodules are often associated with them.

The environment of deposition can be none other than a low sinuosity, and probably braided, sandy bed stream. The third order surfaces are individual braid channels. The second order surfaces define accretionary cross-strata produced by lateral shifting of mid-channel sand bars and flats. Large scale dunes migrated in the deeper channels and up the backs of the bars. The average paleoflow direction, based on the orientation of channel walls (Fig. 3.2.7) is about due west, consistent with the regional depositional setting of Pelletier (1958) (see Fig. 16 of Slingerland and Beaumont, this volume). Sandy braided streams today possess slopes on the order of 10^{-2}, suggesting by linear extrapolation to a mountain front near Philadelphia, PA, that the elevation there was over two km higher than this site. This is a crude approximation of course, especially because the palinspastically restored distance to the thrust front is unknown, but it does give some indication of relief in front of the Acadian Highlands.

DAY 4

SITE 1: ALLEGHANIAN FORELAND BASIN DEPOSITS AT CURWENSVILLE, PA

Rudy Slingerland

LOCATION

The upper part of the Mississippian Burgoon Sandstone and the Pennsylvanian Pottsville and lower Allegheny[1] Groups are exposed in a riprap quarry, railroad cuts, and roadcuts along Route 969, 3.7 km (2.3 miles) southwest of Curwensville, PA (Fig. 4.1.1) on the west side of Curwensville Reservoir. During the Alleghanian Orogeny this site occupied a medial position in the foreland basin, equidistant from both northern and southeastern sediment sources.

SIGNIFICANCE

These units and the paraconformity between them record the foreland response to waning Acadian orogenesis (Burgoon Ss.) and initiation of Alleghanian orogenesis (paraconformity and Pennsylvanian rocks). As discussed in Slingerland and Beaumont (this volume; see Figs. 13 and 17-20), the Pocono (Burgoon) alluvial plain was prograding westward during Lower Mississippian time in response to downwasting of Acadian thrust sheets. In Meramecian time this

[1]Note that the spelling of the Group name is with an e, after the mountain in Pennsylvania, but the spelling of the orogeny is with an a, after the Alleghany Mountains of Virginia.

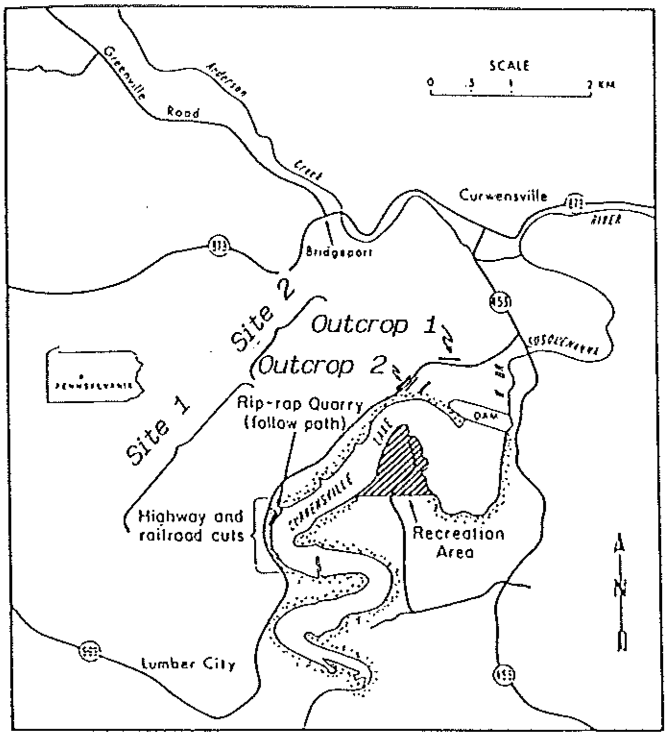


FIGURE 4.1.1 Location map of sites during Day 4 (modified from Berg, 1987).

alluvial plain was flooded along a northwest-southeast axis through Pittsburgh, PA, due to crustal flexture in response to thrust loading to the south. At the same time, the terrane to the north was uplifted as a foreland bulge. Subsequently, this bulge migrated to the south in response to crustal relaxation, and the Curwensville area experienced erosion of all previously deposited Late Mississippian strata (Mauch Chunk and Loyalhanna Fms.) down to the Burgoon Ss. Further to the north in Pennsylvania, the paraconformity cuts even lower, resting on the Upper Devonian Huntley Mountain Fm (Berg, 1987), thus necessitating a minimum of half a kilometer of bulge uplift. The Mercer hard clay developed at this time by deep weathering of still emergent Burgoon and Mauch Chunk highlands (Williams, 1960; Williams and Bragonier, 1974).

By earliest Pennsylvanian time the fluvial Lower Connoquenessing Ss. of the Pottsville Gp. was being deposited by streams draining the craton to the north. The rapid expansion of the piedmont plain to the east and southeast however, soon overran this area such that the Upper Connoquenessing Ss. was deposited by northwestward flowing streams draining the orogenic terrane. By Late Middle Pennsylvanian time, the region was a vast coastal deltaic coastal plain receiving sediments from the thrust sheets to the south-southeast (see Slingerland and Beaumont, this volume, Fig. 20). Numerous transgressive-regressive cycles formed by primarily eustatic sea-level changes produced the many fluvial sandstone---coal---marine or brackish shale sequences higher in the section.

SITE DESCRIPTION

The lower quarry wall (Fig. 4.1.2) exposes a light gray, quartzose graywacke sandstone, identified variously as the Lower Connoquenessing Sandstone of the Lower Pennsylvanian Pottsville Group, a sandstone in the Late Mississippian Mauch Chunk Fm. (Williams et al., 1985b), or most recently (Berg, 1987), as the Early Mississippian Burgoon Sandstone. It is fine- to medium-grained, medium- to thin-bedded, and large-scale trough cross-stratified. Its environment of deposition is in doubt; the quartzose nature and sedimentary structures suggest a weathered fluvial channel deposit.

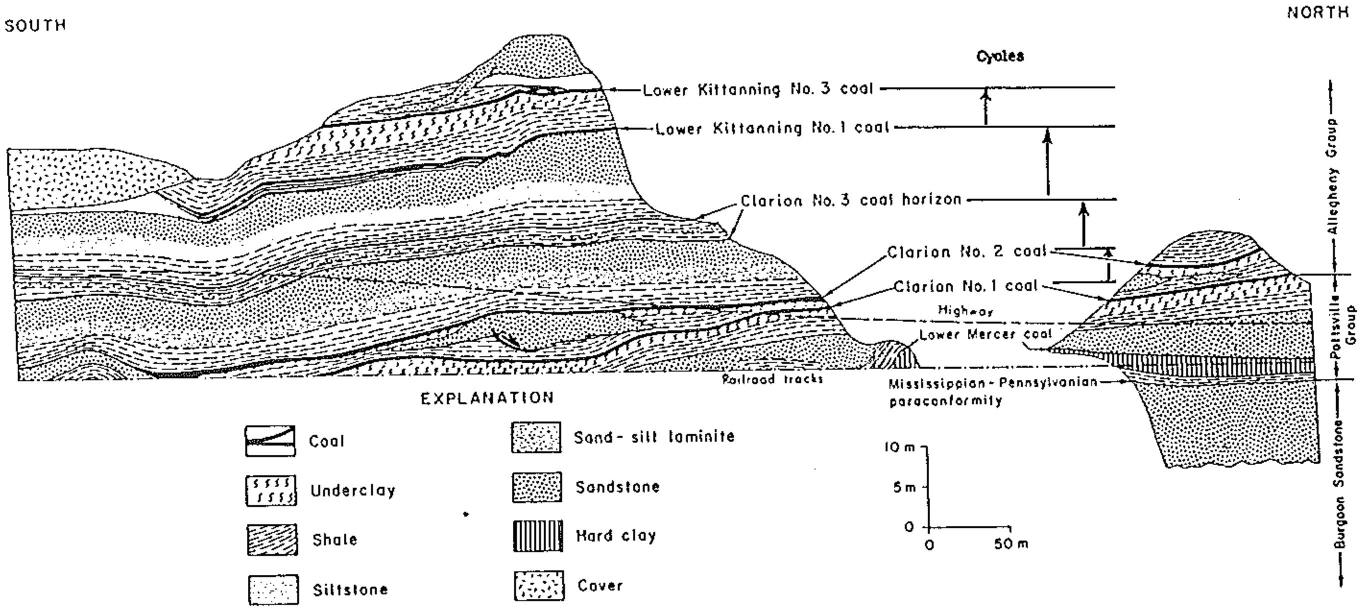


FIGURE 4.1.2 Stratigraphic cross-section of Carboniferous rocks at Site 1, Day 4 (modified from Berg, 1987).

Near the top of the quarry, a 1 to 1.5 m interval of silt shale, siltstone, and very fine-grained sandstone is thought by Williams et al. (1985b) and Berg (1987) to represent the lateral equivalent of the Lower Connoquenessing Sandstone, and the base of the Pottsville Group at this site. If true, then a 25 Myr paraconformity occurs between this unit and the subjacent Burgoon Ss. Alternatively, the paraconformity may occur above the fine-grained interval, in the 2.5 m thick, dark gray, high-alumina Mercer hard clay. This seems more reasonable, given the duration of time necessary to produce diaspore hard clays elsewhere. The clay consists from bottom to top of flint clay (well-crystallized kaolinite), .6 m of burnt nodule clay (80% diaspore, 18% pyrite, 2% kaolinite), .3 m of green nodule clay (100% diaspore), and the remainder of flint clay (Williams et al., 1985b). It is one of two occurrences of high alumina clays in the United States and is extensively mined for refractory clay products. Above the clay, and exposed in the railroad and highway cuts, is an additional 6 to 13 m of coals, fluvial sandstones, siltstones, and soft clays of the Pottsville Group. The base of the overlying Allegheny Group is placed at the base of the Clarion No. 1 coal of lower Desmoinesian stage. Note that the Pottsville Group is at maximum, 17 m thick here compared to 332 m eastward at Pottsville.

Above the Pottsville Group rocks, and continuing to the top of the exposure, are strata of the Allegheny Group, the time equivalents of the uppermost Pottsville and lower Llwellyn Formations to the east. The Allegheny Group here consists of at least four transgressive-regressive cycles or cyclothems (Fig. 4.1.2). A typical cycle (Fig. 4.1.3) starts with a coal formed in peat swamps on a delta plain. The coal is overlain by either a lacustrine, bay, or marine shale, depending upon the detailed history of delta lobe abandonment and the magnitude of the marine incursion. Cut into the shales from above are fluvial channel sandstones that in turn grade upwards to overbank siltstones, underclay, and the next higher coal.

The outcrop also dramatically demonstrates the influence of compaction on sedimentation, a feature first pointed out by Williams and Bragonier (1974) (Fig. 4.1.4). Notice that the undulating surface upon which the Mercer No. 2 coal was deposited has affected the thickness and location of overlying units. The sandstone above the Mercer No. 2 coal is thickest where the coal is thickest, because the areas of thicker peat compacted more during sand deposition. Subsequently, the areas of thicker sand compacted less, and the peat forming the Clarion No. 1 coal accumulated to a greater thickness in adjacent topographic lows. This differential thickness provided the conditions necessary to perpetuate the process.

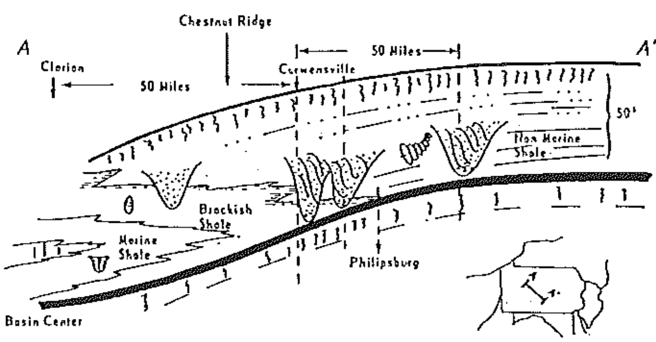


FIGURE 4.1.3. Schematic cross-section into the foreland basin center showing the onlap-offlap origin of a typical Allegheny Group cycle (modified from Williams et al., 1985b).

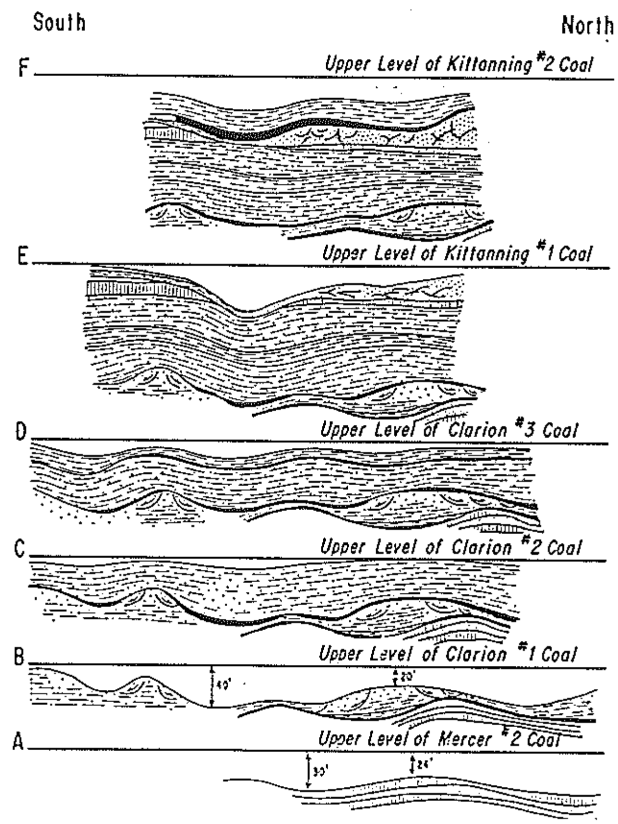


FIGURE 4.1.4. Historical development of the sequence depicted in Fig. 4.1.3 (modified from Williams et al., 1985b).

SITE 2: ALLEGHANIAN FORELAND BASIN DEPOSITS
AT CURWENSVILLE, PA

Rudy Slingerland

LOCATION

The strata surrounding the lower and middle Kittanning coals of the Allegheny Group seen at Site 1 are better exposed approximately 2.4 km to the northeast of Site 1 along Route 969, directly opposite the flood spillway of Curwensville Dam. Two outcrops are available for examination (Fig. 4.1.1 of Site 1): Outcrop 1, consisting primarily of flood basin and bay fill sequences, and Outcrop 2, consisting of the laterally equivalent channel fills.

SIGNIFICANCE

These strata allow a more detailed understanding of the paleodepositional environments and sedimentary processes in the interior of the Alleghanian foreland basin, and illustrate the data from which the paleogeographic reconstructions of Slingerland and Beaumont (this volume) are derived.

SITE DESCRIPTION

The Lower and Middle Kittanning coals are the principal seams exposed in the highwall (Fig. 4.2.1). A local coal at Sections 1 and 2 (Fig. 4.2.1) below the Middle Kittanning thickens into a paleotopographic low. If we assume the peat swamp surface was a horizontal plane from Section 1 to Section 2, we can calculate the compaction ratio for the coal; in this case it is 6 to 1. The coal also is brighter in the low, an observation interpreted by Williams et al. (1985b) to be a result of more water saturated conditions during accumulation and therefore less oxidation.

Above this coal and below the Lower Kittanning coal, such as at Section 2, is a plastic underclay, thought by Williams et al. (1985b) to represent an erosional interval of tens of thousands of years. It was mined at Outcrop 2 for ladle bricks, a product that expands at high temperatures.

Two different facies assemblages overlie the Lower Kittanning coal in this area, a fine-grained, coarsening-upwards assemblage such as seen at Sections 3 amd 13, and a coarser-grained channel sandstone, levee, and abandoned channel-fill sequence such as seen at Sections 4-12. The fine-grained assemblage is representative of three-quarters of the lower and middle Allegheny rocks in this area. It consists of a dark, fissile, Lingula-bearing, silt shale that grades upwards into laminated siltstones and fine-grained, wave-rippled sandstones. This is interpreted as an estuary bay-fill sequence arising from prograding distributary crevasse splays.

The coarse-grained facies assemblage is dominated by a light brown, medium- to coarse-grained graywacke sandstone body, a good example of which occurs at Section 4. There, it is cut into the fine-grained sequence. Gravel-sized intraformational conglomerates and plant debris locally overlie the erosion surface. The dominant internal structures are extensive lateral accretion surfaces dipping northeastward. The accretion beds themselves often have scoured surfaces, that are locally concave upwards. Internally, the beds are large-scale trough cross-stratified (mean paleoflow to N 80^{o} W), and often fine upwards to a siltstone or shale drape. We interpret sandstone bodies of this type to result from lateral deposition of channel-margin bars, probably point bars of a sinuous upper delta plain river, although we can not rule out side-bars of a lower delta plain distributary.

The features at Section 5 are interpreted to be slump blocks produced by compaction and lateral flow of the subjacent mud due to loading by the sand. The mud was squeezed up into the channel at Section 6 by slumps on both the channel-margin bar and the cut bank of the stream (Section 7). Interestingly, if the slumps are rotated back to their original position, they extend above the present position of the Middle Kittanning coal. Thus, an interval of erosion must have occurred after slumping, but prior to peat formation. The pure quartzite gannister at Section 9 is thought by Williams et al. (1985b) to have formed during this interval. Following slumping, and possibly because of it, the channel was abandoned, and filled with a black carbonaceous shale.

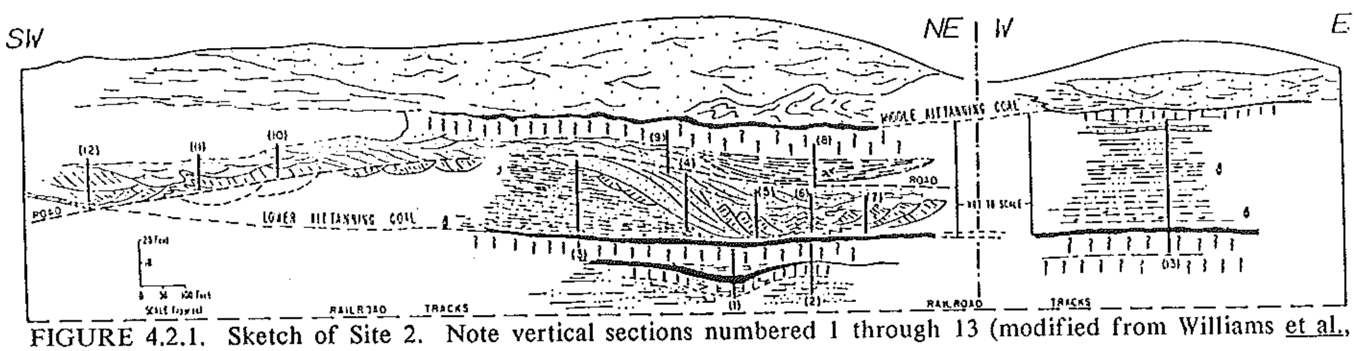


FIGURE 4.2.1. Sketch of Site 2. Note vertical sections numbered 1 through 13 (modified from Williams et al., 1985b).

DAY 5

SITE 1: ACADIAN FORELAND BASIN DEPOSITS AT PORT ROYAL, PA

John Diemer

LOCATION

This site is on Route 322, five kilometers south of Millerstown, Pennsylvania, and was first described by Faill and Wells (1974). The rocks are exposed in one long roadcut on the northbound lanes and in three smaller outcrops on the southbound lanes. We will visit the outcrops on the southbound lanes first and then climb onto the median strip to view the large outcrop on the northbound lanes. Please view the outcrop on the northbound lanes from a distance. Rockfalls are a common occurrence!

SIGNIFICANCE

The rocks exposed here in the nose of the Buffalo Mountain syncline are part of the Duncannon Member of the Catskill Formation, last seen on Day 3, Site 2. This site allows us to better describe the more proximal portion of the alluvial plain that stretched from the Acadian thrust belt to the Catskill Sea during Late Devonian time.

SITE DESCRIPTION

The Duncannon Member in south-central Pennsylvania is composed of two lithofacies

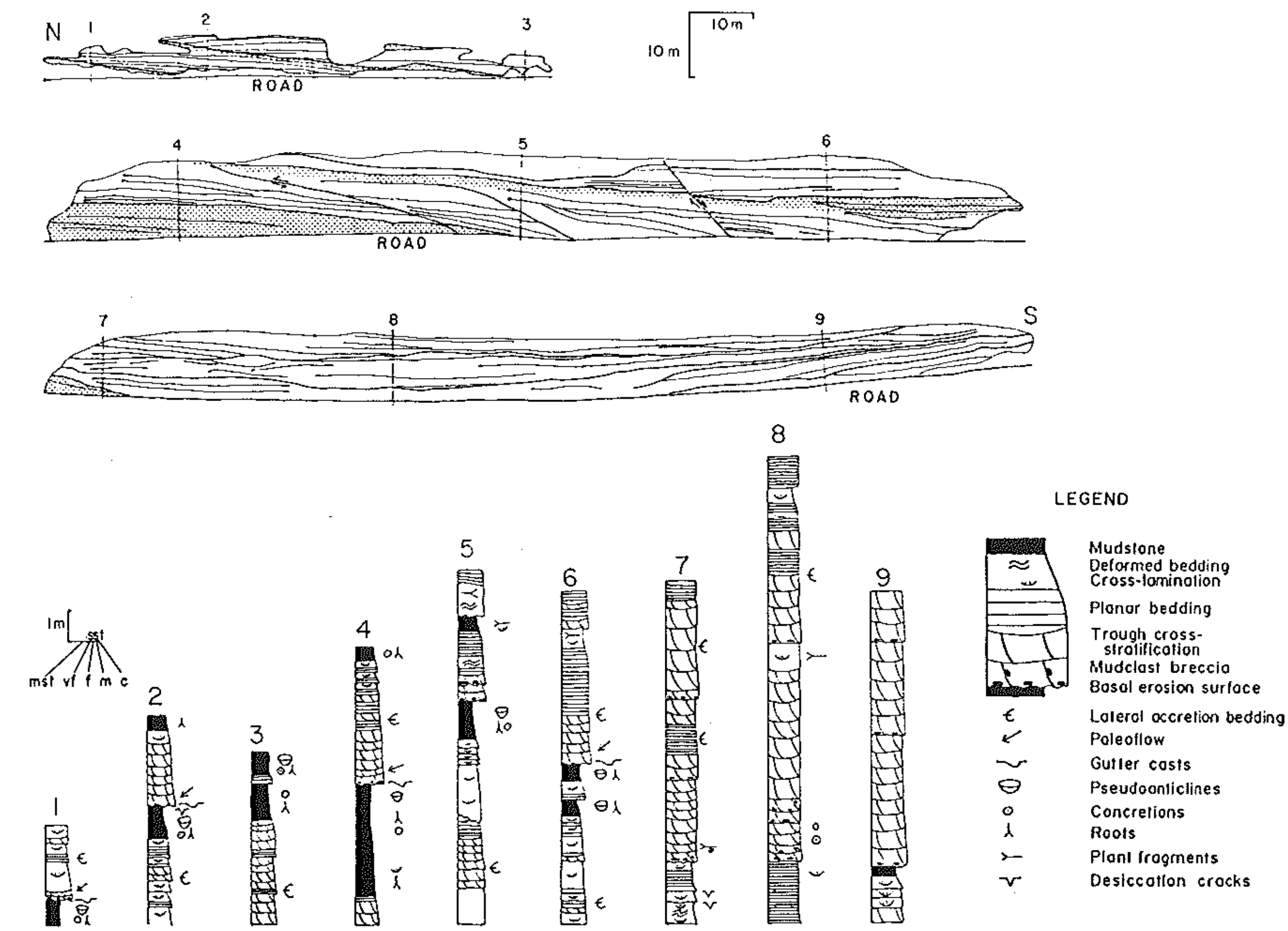


FIGURE 5.1.1 Line drawing overlay and sedimentologic logs of outcrops on the southbound lanes. Sandstone bodies (non-stippled) and sandstone-mudstone interbeds (stippled) are shown. Important erosion surfaces, lateral accretion surfaces and faults are indicated.

associations: sandstone bodies and sandstone-mudstone interbeds (Diemer, 1986; Diemer and Phillips, 1987). The sandstone bodies are composed of medium to very fine-grained, cross-stratified, horizontally-stratified, and cross-laminated sandstones (Figs. 5.1.1 and 5.1.2). Sandstone bodies are underlain by laterally extensive erosion surfaces with decimeters of relief. Basal breccias composed of centimeter-scale, sub-angular mud clasts and rounded concretions occur locally immediately above erosion surfaces. Sandstone bodies typically fine upward and consist of bedsets separated by lateral accretion surfaces. In places, the lateral accretion bedding grades laterally into coarse- or fine-grained channel fills delineated by channel abandonment surfaces. Channel initiation surfaces occur locally (Fig. 5.1.2). In places, tops of sandstone bodies are truncated by erosively-based channel fills which are up to one-third the thickness of the sandstone body. Sandstone bodies are both single- and multi-storeyed. Paleoflows are dominantly westward and most lateral accretion surfaces dip northward (Figs. 5.1.1 and 5.1.2).

The sandstone-mudstone interbeds are erosively-based, fine to very fine-grained, cross-stratified, horizontally-stratified, and cross-laminated sandstones which grade up to cross-laminated, bioturbated or massive mudstones (Figs. 5.1.1 and 5.1.2). These upward-fining bedsets are spatially organized into sheets or channel fills and occur individually or in stacks of bedsets up to several meters thick. The stacks of bedsets may fine or coarsen upward. Transported plant fragments are common above erosion surfaces and on bedding planes of stratified sandstones. Downward-branching, millimeter-scale, clay-lined tubules centimeters in length are common in mudstones. In situ calcareous concretions are common in laterally extensive, decimeter-thick horizons in mudstones. The concretions are centimeter-scale nodules of clastic material cemented by calcite. Some concretions contain millimeter-scale tubules filled by blocky calcite cement. Mudstones also commonly exhibit decimeter to meter thick horizons with blocky weathering surfaces. The blocky weathering surfaces are likely due to the presence of centimeter-scale blocks of mudstone (peds) separated by shiny fracture surfaces with slickensides. In places, large-scale (decimeters to meters in length) slickensided fracture surfaces defining bowl-shaped structures occur in mudstones.

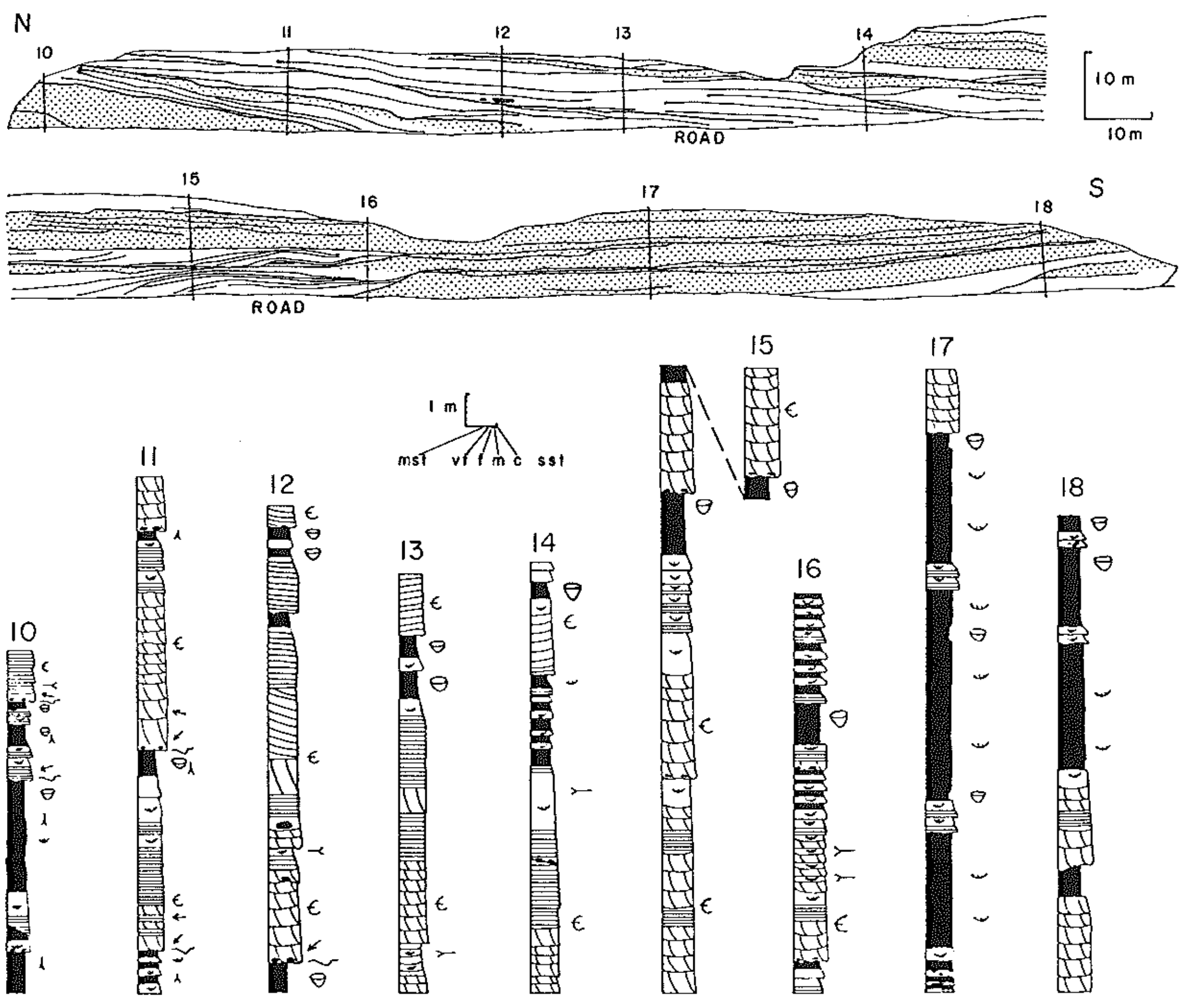


FIGURE 5.1.2 Line drawing overlay and sedimentologic logs of the northbound lanes outcrop. See Figure 5.1.1 for legend.

Trace fossils have been found in horizontally-stratified sandstone float blocks which can be assigned to either the sandstone body facies association or the coarse-grained portion of the sandstone-mudstone interbeds. The trace fossils occur as elliptical, vertical burrows 5 to 8 centimeters in diameter and 5 to 10 centimeters in length. The burrows occur in groups, locally cross-cut each other, are filled with pelleted sand, and some exhibit a faint meniscus structure.

The sandstone bodies are interpreted as point bar deposits of laterally migrating and aggrading single-channel rivers flowing across an alluvial plain (Fig. 5.1.3). Basal erosion surfaces were produced by talweg scouring during lateral migration. Basal breccias are interpreted as reworked cutbank material. The tops of point bars were vegetated and locally incised by chute channels. Single storey sandstone bodies record deposition from single lateral migration events. The consistent grain size and bedset scale within multistorey bodies suggests that down-valley migration of meander bends between avulsion events accounts for multistorey sandstone bodies. The common northward dip of lateral accretion bedding suggests structural control of the direction of channel migration (Leeder and Alexander, 1987).

The sandstone-mudstone interbeds are interpreted as overbank flood deposits on levees, crevasse-splays, and flood basins (Fig. 5.1.3). Each fining upward bedset likely records deposition during a single flood event. The overbank areas were vegetated as evidenced by plant fragments, downward branching rootlets, and concretions (rhizocretions of Klappa, 1980). The overbank areas were subaerially exposed for sufficiently long periods under warm, seasonally wet and dry climates to form calcrete-bearing paleosols (Reeves, 1976; Goudie, 1983; Allen, 1974; Leeder, 1975). The bowl-shaped structures in the mudstones resemble pseudoanticlines (Gordon and Bridge, 1987; Bridge, 1988) and are interpreted as lateral displacement structures developed in soils containing swelling clays subjected to repeated wetting and drying events

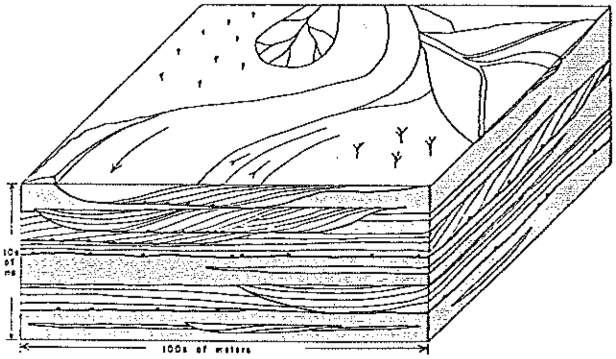


FIGURE 5.1.3 Depositional model for the Duncannon Member of the Catskill Formation. Lateral accretion bedding and channel initiation and abandonment surfaces are shown for the (non-stippled) sandstone bodies.

(Goudie, 1983). The spatial organization of facies within this association and between associations can be explained by channel migration and periodic channel-belt avulsions.

The elliptical, vertical burrows are interpreted as the escape structure of a freshwater, filter-feeding mollusc, Archanodon, oriented with its commissural plane parallel to paleoflow (see Bridge, Gordon and Titus, 1986). As the sand was deposited, Archanodon migrated upward in order to remain at the sediment-water interface. The size of the burrows indicates mature individuals and long periods of sustained (perennial) flow, either on point bar surfaces or in major overbank channels.

Paleohydraulic reconstruction (see Bridge, 1978, 1982; Bridge and Diemer, 1983) of channels suggests two distinct sizes: small channels which were about 45 meters wide and had mean depths of about 1.85 meters, and large channels which were about 90 meters wide and had mean depths of about 3.5 meters. Paleoslopes for both sizes of channels were about .00009 and sinuosities were about 1.20. The smaller channels may be tributaries of the larger channels.

SITE 2: PROXIMAL LLEWELLYN FM. OF THE ALLEGHANIAN CLASTIC WEDGE AND THE HISTORY OF ANTHRACITE MINING IN THE UNITED STATES

Rudy Slingerland

Fourty years I worked with pick and drill
Down in the mines against my will,
The Coal King's slave, but now it's passed;
Thanks be to God I am free at last.

-From the tombstone of an anthracite miner

LOCATION

This site is in the historic town of Ashland, PA, in the western middle anthracite field. The western middle is one of four fields in eastern Pennsylvania, each located in a synclinorium of the fold and thrust belt that preserves Upper Paleozoic strata (Fig. 5.2.1).

SIGNIFICANCE

The anthracite region of Pennsylvania contains the world's largest deposit of hard coal. Because of that fact, and its location adjacent to the urban centers of early 19th century America, an Anthracite Empire developed, of staggering wealth and power for its barons and grinding poverty for its workers. Here was the cradle for the heavy industrial revolution in America. Here men like Asa Packer and Charles Brodhead first learned to vertically integrate an industry, owning coal mines in the eastern middle field, carbonate quarries in the Lehigh Valley, iron ore mines in the Lehigh Valley, the Lehigh Valley railroad, and the Bethlehem Iron Company. Here too, was waged "the greatest conflict between capital and labor which the world has ever seen," in the words of Clarence Darrow. It started in the 1850s with the terrorist activities of the Molly Maguires, a secret society dedicated to improving the lot of the Irish immigrant workers, and culminated in the Great Strike of 1902 called by the United Mine Workers charismatic president, John Mitchell. Finally, here was a new life, cruel as it was, for millions of European immigrants.

Now it would seem that the kingdom of coal is gone. Production peaked in 1917 at 99.6 million tons and has declined steadily ever since to a present annual 6 million or so tons. The black diamond that broke America's dependency on foreign coal, fueled an industrial revolution, kept millions warm, created great wealth, and gave birth to a vibrant immigrant culture, now has only a ravaged landscape for a legacy.

SITE DESCRIPTION

The Anthracite Museum of Ashland focuses on the mining and processing of anthracite coal. We will visit a restoration of an early mining operation, the Pioneer Tunnel. It represents a horizontal drift mine, extending for 364 m (1200 ft) into Mahanoy Mountain.

The coal seams occur in the Pottsville and Llewellyn Formations of Pennsylvanian age (Fig. 5.2.1); the Buck Mountain is thought to be coeval with the Lower Kittaning coal seen on Day 2. The most important seams are the Mammoth, generally consisting of two splits, each 1 to 4 m thick, and the Buck Mountain, averaging 1 to 2 m thick. The coal is a high carbon (92-98%, moisture and ash free), high BTU (14,430-15,450, moisture and ash free), low-volatility (2-8%), and low sulfur (.3-1.2%, as received) anthracite.

Unlike the bituminous fields further west in the foreland, where the strata are almost horizontal, the anthracite fields were complexly folded and faulted during the Alleghanian Orogeny (Arndt and Wood, 1960), making mining expensive and dangerous. The 2nd and 3rd order synclines, and to a lesser extent the anticlines, are upright or slightly overturned tight folds verging to the northwest. Most folds follow a parallel style up to the point when appression of the limbs creates a space problem; then the folds become disharmonic with the coals and shales acting as glide planes. Some mines encountered folds in which all the coal in the limbs had been squeezed into the fold nose. High-angle and low-angle thrust faults, bedding plane faults, underthrust faults, and tear faults are also common. The high angle south-dipping thrust faults commonly originate in the cores of tight synclines, and rise to the north, cutting the north limb of the syncline. Under these conditions a breast and pillar method of underground mining was employed. Edmunds (1972) estimated that, of the remaining in-place coal within 300 m of the surface, over 7 billion short tons were recoverable in 1972.

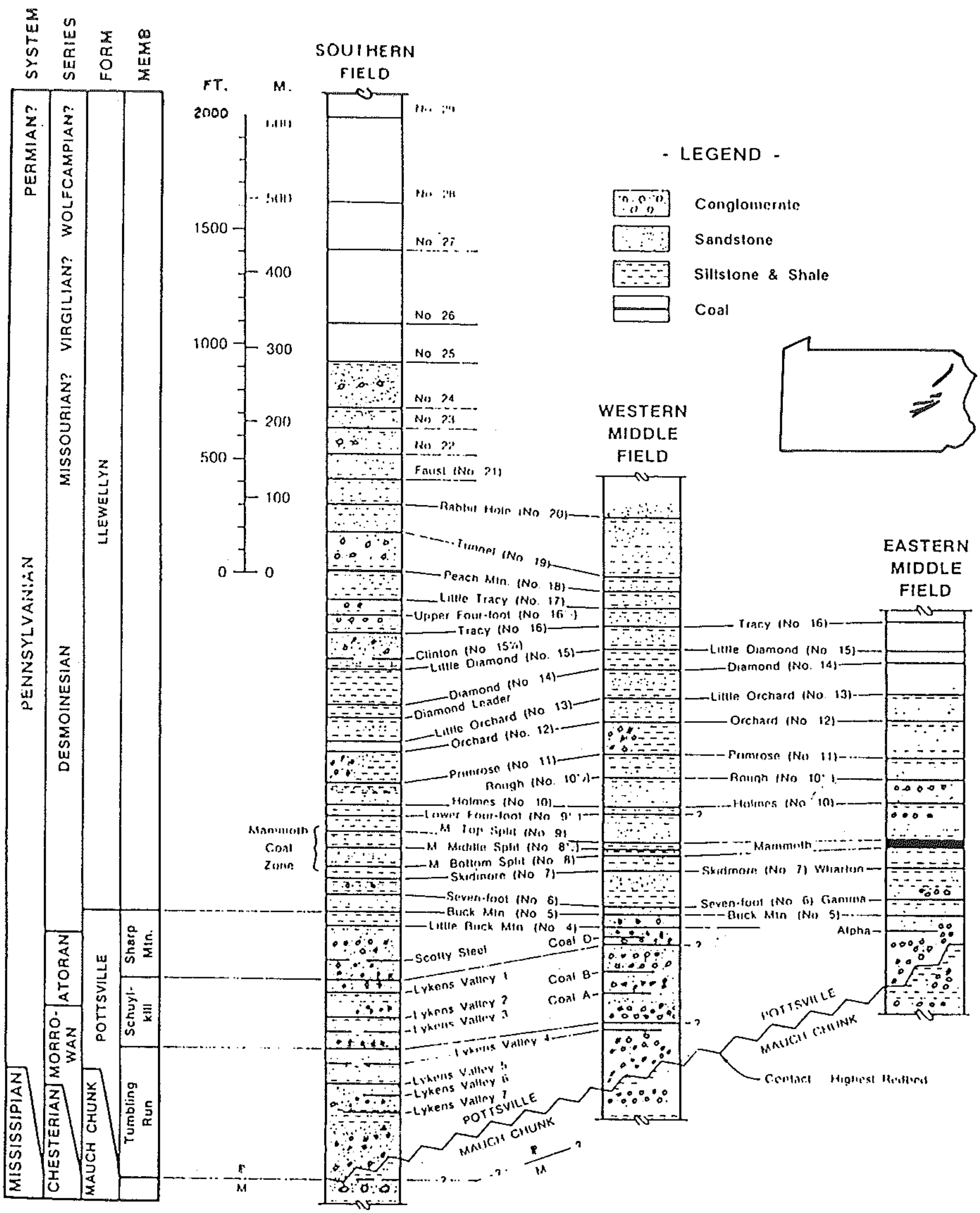


FIGURE 5.2.1 Representative columns of upper Paleozoic strata found in the middle and southern anthracite fields of Pennsylvania (modified from Eggleston and Edmunds, 1981).

DAY 6

SITE 1: ALLEGHANIAN FORELAND BASIN DEPOSITS AT POTTSVILLE, PA

Excerpted from
Levine and Slingerland (1987)

LOCATION

The rocks at this site are exposed along a road cut on the eastern side of Pennsylvania 61, 0.4 to 0.8 km south of Pottsville, Pennsylvania (Fig. 6.1.1), on the southern margin of the Southern Anthracite field where the Schuylkill River has cut a deep gap in Sharp Mountain.

SIGNIFICANCE

The outcrop exposes a 600+ m thick section of upper Carboniferous molasse, representing the northwestward influx of clastic detritus into the Alleghanian foreland basin from an orogenic source terrane formerly situated to the south-southeast. The alternation of facies (Fig. 6.1.2) reflects the gradual but progressive evolution of depositional environments from a semi-arid alluvial plain (Mauch Chunk Formation), to a semi-humid alluvial plain (Pottsville Formation), to a humid alluvial plain dominated by peat swamps (Llewellyn Formation). This transition, documented by dramatic changes in sedimentary facies, facies sequences, and maximum clast sizes, clearly reflects regional (perhaps even world-wide) climatic changes occuring near the end of the Mississippian, incipient Alleghanian tectonism, and the evolution of many new plant groups.

Subsequent to their deposition, the Carboniferous sediments were deeply buried, metamorphosed, tectonically deformed in the Alleghanian orogeny, uplifted, and largely eroded. The Southern Anthracite field now preserves the thickest, coarsest-grained, most proximal to the source, and most stratigraphically continuous occurrence of upper Carboniferous molasse in the central Appalachians.

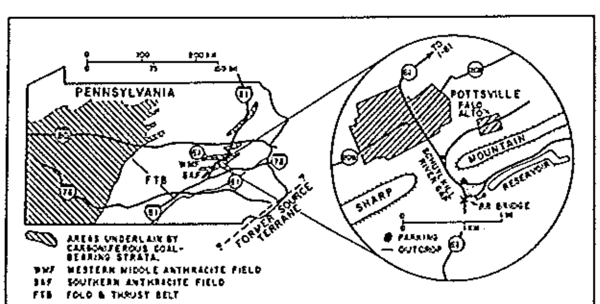


FIGURE 6.1.1 Field location of Pottsville section (Day 6, Site 1) (from Levine and Slingerland, 1987).

SITE DESCRIPTION

Stratigraphic and Geomorphic Overview

Molasse sediments of the Anthracite region are stratigraphically subdivided on the basis of grain size and predominant coloration (Wood and others, 1969). The fine-grained, red Mauch Chunk Formation (Middle to Upper Mississippian) intertongues with and is replaced by the coarse-grained, gray Pottsville Formation (Lower to Middle Pennsylvanian), which in turn gives way to the finer-grained, gray to black, coal-rich Llewellyn Formation (Middle Pennsylvanian), representing the youngest extant molasse in the region. The former presence of many miles (kilometers) of overlying rocks is implied by the high coal rank and compaction of the Llewellyn sediments (Paxton, 1983; Levine, 1986).

The Mauch Chunk Formation is informally subdivided into three members (Wood and others, 1969). The middle member represents the 'type' Mauch Chunk red bed lithofacies. The lower and upper members represent respectively the zones of intertonguing with the underlying Pocono Formation and the overlying Pottsville Formation. The upper contact of the Mauch Chunk is defined as the top of the uppermost Mauch Chunk-type red bed (Fig. 6.1.2).

The Pottsville Formation is formally subdivided into three members (Wood and others, 1956), each representing a crudely fining-upward megacycle. Of the three, the Tumbling Run and the Sharp Mountain members are the coarser-grained, while the intervening Schuylkill Member is finer-grained and contains a greater proportion of coal. The lower contacts of the Schuykill and Sharp Mountain members are defined at the base of major conglomeratic units. The base of the Schuylkill Member is by no means obvious at the outcrop, but the "Great White Egg" quartz pebble conglomerate at the base of the Sharp Mountain Member is very distinctive. The contact between the Pottsville and Llewellyn Formations is placed at the base of the lowermost thick, stratigraphically persistent coal horizon, the Buck Mountain (#5), which has been correlated over large areas of the Anthracite fields (Wood and others, 1963).

Chronostratigraphic age designations in the Anthracite region, based upon the 13 upper Paleozoic floral zones defined by Read and Mamay (1964; also see Edmunds and others, 1979 , Fig. 11), indicate the Pottsville section is conformable, extending from Zone 3 in the upper Mauch Chunk Formation (Chesterian

Series) to Zone 10 in the lower Llewellyn Formation (Des Moinesian/ Missourian Series); however, Zones 7 and 8 have not been explicitly recognized at this site. The Mauch Chunk/Pottsville contact, occurring between Zones 3 and 4, corresponds roughly to the Mississippian/Pennsylvanian systemic boundary. In areas of the central Appalachians other than the Southern and Middle Anthracite fields, Zones 4, 5, and 6 are absent, suggesting the presence of a significant disconformity between the youngest Mississippian and oldest Pennsylvanian strata (see discussion in Slingerland and Beaumont, this volume).

The strata exposed at the site are slightly overturned and comprise part of the southern limb of the Minersville Synclinorium, forming the southern margin of the Southern Anthracite field. They attained their present attitude during the late Paleozoic Alleghanian orogeny when northwest-directed tectonic forces produced a progression of deformational phases that migrated northwestward across the foreland basin. At the Pottsville site all structural phases are superposed (Wood and Bergin, 1970; Nickelsen, 1979).

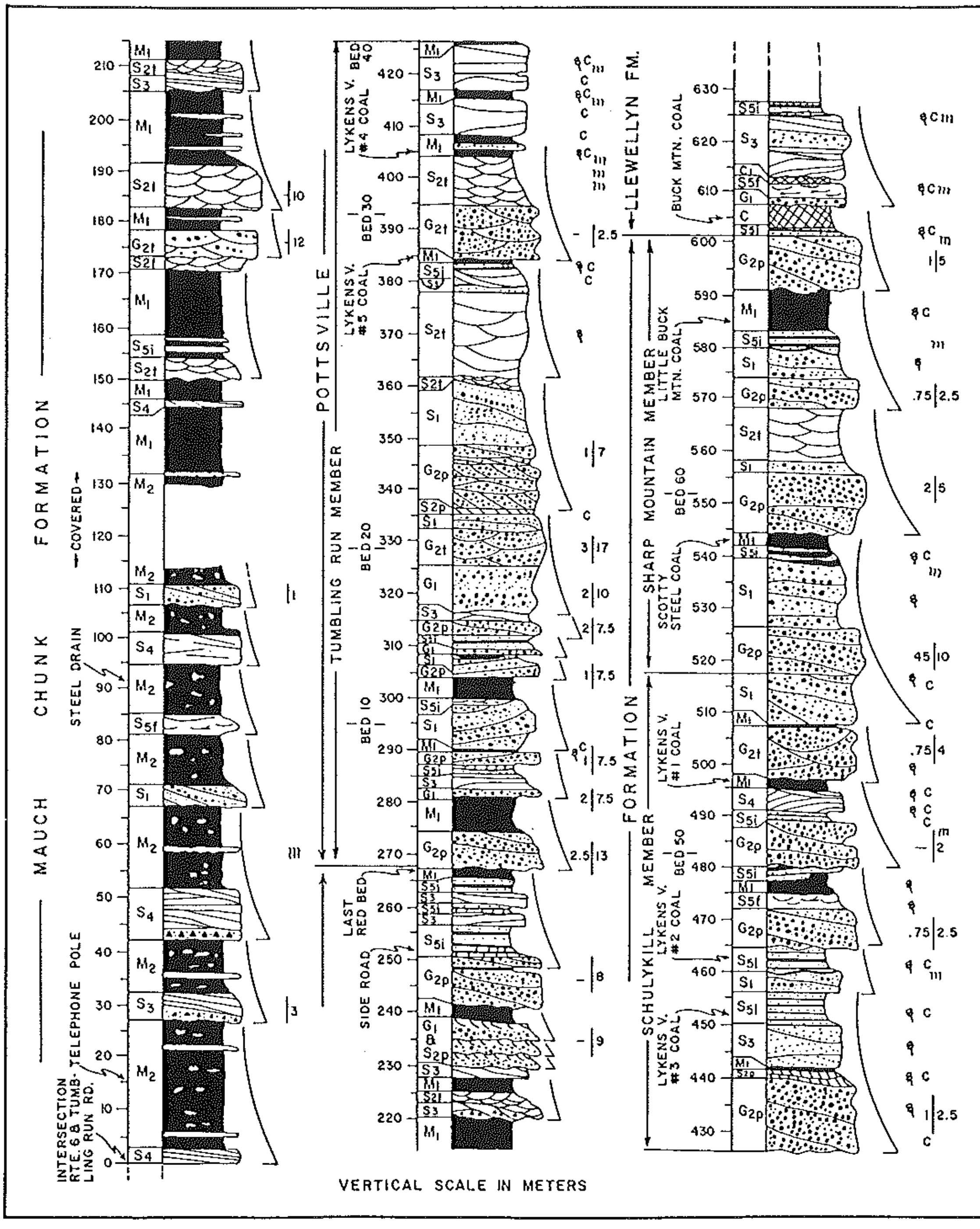


FIGURE 6.1.2 Stratigraphic column of Pottsville section (from Levine and Slingerland, 1987).

The structure and stratigraphy of the upper Paleozoic molasse sequence are revealed geomorphically by the relative resistance to erosion of the near-vertical component units. The Pocono sandstone, subjacent to the Mauch Chunk Formation, upholds Second Mountain, the major ridge visible to the south of the Pottsville section. The Mauch Chunk Formation underlies the valley between Second and Sharp mountains. The distinctive double ridge of Sharp Mountain is formed by the Tumbling Run and Sharp Mountain members of the Pottsville Formation. The Schuylkill River, which excavated the gap in Sharp Mountain, flows southeasterly across the Valley and Ridge Province on its course to the Delaware Bay, opposite to the streams that originally deposited the Pottsville sediments.

Sedimentology of the Pottsville Section--Facies States and Composition

Sedimentary bedforms, sediment composition, facies sequences, and paleobotany reveal a significant alteration in paleoclimatic conditions across the Pottsville section, ranging from generally semi-arid, poorly vegetated conditions at the base to perennially humid, lush conditions at the top. Ten general facies have been defined at this site and are described in Table 6.1.1. Transition matrix analysis reveals two repeating motifs, one characteristic of the Mauch Chunk and one of the Pottsville. When compared to facies sequences from modern environments of deposition, the Mauch Chunk sequence is similar to that of Bijou Creek, Colorado, a sandy, braided, ephemeral stream subject to catastrophic floods (Miall, 1977). Facies S3 and S4 probably comprised sand flats or shallow channel deposits; S5i and S2t comprised waning flow deposits or overbank deposits more removed from the active channel. M1 represents intra-channel, slack water deposits and M2 represents overbank soils.

The Pottsville sequence is similar to that produced by the Donjek River, Yukon Territory, a gravel-sand mixed bedload, perennial braided stram (Miall, 1977). Facies S2t and S5t formed in the upper parts of active channels or minor channels and on the tops of braid bars. Facies S5i and M1 formed on bar tops, abandoned channels, and overband areas, and facies C was deposited in inter-channel swamps. The channels forming the Pottsville Formation were deeper with greater cross-sectional areas, and lower width/depth ratios than those forming the Mauch Chunk Formation. In consequence, maximum clast size is greater as is the thickness of cross-bed sets.

Sandstone petrology, organic matter content, clay mineralogy, and features of the paleosols (Table 6.1.1) all show a progressive trend to more highly leaching, less oxidizing (i.e., more humid) conditions higher in the section. Sandstones are compositionally mature throughout the section but become even more mature up section. The Tumbling Run Member of the Pottsville Formation contains the highest variety and

TABLE 6.1.1

CODE:	G_1	$G_{2T\&P}$	S_1	$S_{2T\&P}$	S_3	S_4	$S_{5F\&I}$	M_1	M_2	C
SYMBOL:		t p		t p			f i			
NAME:	Crudely bedded sandy conglom.	Cross bedded sandy conglom.	Plane bedded pebbly sandst.	Cross bedded pebbly sandst.	Coarse, plane bedded sandst.	Fine, plane bedded sandst.	Flaser or interbedded sandst. & mudst.	Noncalcareous mudst.	Calcareous mudst.	Coal
COLOR:	Pale gray ss. with variegated conglom.	Dusky yellow conglom. with gray sandst.	Light olive gray	Grayish orange to pink	Pale red (Mauch Ch.) Pale olive (Pottsv.)	Grayish to pink	Gray red (Mauch Ch.) or dark gray (Pv.)	Ruddy (MC) to light brown to black (Pv.)	Ruddy to brown	Black as coal
GRAIN SIZE:	Coarse sand to pebble conglom.	Pebble conglom. to very coarse sand	Coarse to granule with pebbly stringers	Coarse to granule	Coarse to very coarse	Very fine to fine sand	Fine sand with interbd. mud	Fine clay to silt	Fine clay & silt with carbonate concretions	Finely macerated plant fragments
INTERNAL BED FORMS:	Subhoriz. interstrat. coarse sand & pebble conglom. Lenticular medium-thick beds	Medium to very thick lenticular beds. Matrix supported conglom. (Note 1)	Low angle laterally contin. wedge sets. Medium to thick, concave upward	Cross bedding Lg. scale trough (S_{2t}) or Sm. scale planar (S_{2p})	Decimeter-thick tabular or wedge sets. Lateral cont. to 10s of m. Mass. or par. laminated	Tab. or wedge-shape beds. Laterally extensive thin to thick parallel laminae	Small scale cross-strata with mud fasers (S_5f) or laminated w/ mud layers (S_5i) (Note 2)	Finely laminated or rooted see Note 3.	Stratified or paleosol features (see Note 3) M. C. only	Finely stratified but sheared during tectonic deform.
COMPOSITION MAUCH CHUNK TYPE:	(No specific data)		Maturity: mature	Monoxl quartz: 60%	Feldspar: <1%	Rock fragments plus mica. Avg. 15% Generally sedimentary or low-grade metamorphic origin; primarily zircon & tourmaline.		90% of M. C. Shales are red beds (org.-void) Relatively Fe-rich. Clays: 80% illite; 20% chlorite; $CaCO_3$ & SiO_2 concretions in M_1.		None
PV.–LLEWELL.–TYPE:	Vein qtz > sandstone, chert, conglom., silt, and shale > low to med grade metamorph. incl. phyllite, slate and schist.		more mature	60%	<1%	(3-30%) Med. to high-grade metamorphic minerals first appear in upper Mauch Chunk indicating unroofing of these rocks in source terrane.		All are grey to black org.-bearing Fe-poor, Al-rich. Primarily ill. & chlor. with → 40% kaol. & pyrophyllite. No caliche.		> 92% fixed carbon, dry mineral matter-free.
TYPICAL BASE:	Undulatory, sharp	Undulatory, sharp, erosive	Undulatory, sharp, erosive	Gradational from S_1 or erosive from M_1	Gradational or sharp	Sharp, planar dips to 15°	Gradational from S_3 or S_4	Gradational from S_3 or S_5	Gradational	Gradational from M_1
TOP:	Gradational	Grades to S_{2t}	Sharp, undulatory, [illegible]	Grad. to S_{5f} or M_1	Grad. to S_{5f} or S_{5i}	Grad. to M_2	Grad. to S_1 or M_2	Sharp or erosive	Sharp or erosive	[illegible] to G_1, Grad. to M_1
NOTES:	1. G_2 bed continuity ranges from 3m to across outcrop. Lenses subparallel to 2nd-order truncation surfaces; either laterally extensive tangential low angle (<15°) planar cross-strata (G_{2p}) or large-scale (2-4 m) cut-and-fill troughs (G_{2t}) 2. S_1 unit contains abundant dark brown calcareous root traces, desiccation cracks, and raindrop impressions in Mauch Chunk or plant fragments in Pottsville. 3. Paleosol features of M_1 & M_2: In Mauch Chunk, paleosols are "vertisols" with wetting/drying features, including wedge-shaped peds, blocky joints, mud cracks, raindrop impressions, pretectonic slickensides, & caliche. Paleosols in Pottsville Fm. are "underclays" with abundant root impressions, leached clay minerals, & no bedding.									
OTHER SYMBOLS:	▲▲▲ Intraformational conglomerate; Calcareous concretions; Root traces; Plant fragments; C Carbonaceous; Erosional contact; 3|10 average (3) and maximum (10) clast size in centimeters; Fining and thinning upward cycle; Disconformity									
TRANSITION MATRIX SUMMARY SEQUENCES:	Mauch Chunk No. 1 $S_4 \rightleftarrows M_2$; Mauch Chunk No. 2 $S_3 \rightarrow S_{5f} \rightarrow M_1$, $S_3 \rightarrow S_{5i} \rightarrow M_1$; Pottsville $G_{2p} \rightarrow S_1 \rightarrow S_{2t} \rightarrow M_1$, $S_1 \rightarrow S_{5f} \rightarrow M_1$, $S_3 \rightarrow S_{5i}$									

(Compositional information from Meckel, 1967; Wood et al., 1969; Holbrook, 1970; Hosterman et al., 1970) proportion of non-quartzose fragments while the Sharp Mountain Member contains the highest proportion of vein quartz (Meckel, 1967). Preservation of organic matter in the upper part of the section implies conditions of low Eh, maintained by continuous saturation by stagnant or slowly moving water. Clay minerals are enriched in alumina and depleted in iron higher in the section indicating a greater degree of chemical and biological leaching.

Paleosols occurring throughout the section are particularly useful in revealing paleo-environmental conditions. Most paleosols of the Pottsville and Llewellyn Formations formed as underclays beneath peat swamps and, therefore, must have been water-saturated during most of their development. In contrast, paleosols of the Mauch Chunk Formation, classified as vertisols by Holbrook (1970), exhibit a variety of features indicating episodic wetting/drying cycles (Table 6.1.1).

Caliche, occurring as thin, bed-parallel laminae or in nodular layers less than 3 ft (1 m) in thickness is common in the middle member of the Mauch Chunk (Fig. 6.1.2) and occurs occasionally in the upper member. Caliche forms in seasonally arid conditions when surface evaporation produces supersaturation of dissolved salts, especially calcium carbonate and silica. The laminar caliche is interpreted to have formed at the sediment surface in shallow ponds during evaporative cycles (Holbrook, 1970). A surface or near-surface origin is indicated for the nodular caliche as well (Holbrook, 1970) based on: (1) sedimentary laminations that pass from the surrounding sediment into the concretions, (2) nodules occurring as intraformational clasts in conglomerates, (3) the presence of carbonate as nodules in the shales but not as cement in the adjacent sandstones, and (4) ball and pillow structures occurring between the nodules and the underlying (but not the overlying) sediments.

The composition of the organic matter and clay minerals has been strongly influenced by diagenetic conditions during burial. The coal has been elevated to anthracite rank. Expandable layer clays are not present and illite is of the highly ordered 2-M form, representing "anchizone alteration. Pyrophyllite is an anchizone alteration product of kaolinite that forms only in Fe-depleted rock (cf., Hosterman and others, 1970, Table 1). Ammonium illite is thought to form at high coal rank in organic matter-rich sediments by nitrogen released during late stages of coalification (Paxton, 1983). These transformations imply temperatures of ca. 225-275^{o}C and 4 to 6 mi (6 to 9 km) of burial.

Tectonic Significance of the Pottsville Section

During deposition of the Pottsville section the source region lay south of Philadelphia as indicated by paleocurrent directions, regional trends in maximum grain size (Pelletier, 1958; Meckel, 1967; Wood and others, 1969), and the modelling presented in Slingerland and Beaumont (this volume). North-northeast flowing streams carried sediments toward the basin axis, which trended northeast-southwest across western Pennsylvania. Time equivalent upper Carboniferous rocks are alluvial in eastern Pennsylvania and deltaic and shallow marine to the west (Edmunds and others, 1979). The influx of coarse clastics in the Pottsville interval has traditionally been ascribed to tectonic uplift in the source (e.g., Meckel, 1967), but it may also reflect the change to more humid climatic conditions in the Pennsylvanian that produced larger sediment yields and stream discharges.

An additional factor influencing the stratigraphic succession may have been the diversification and proliferation of terrestrial plants during the middle Carboniferous. Plant evolution could have helped to stablilize stream banks, allowed peat accumulation rates to equal or exceed basin subsidence, and influenced climatic patterns.

The intertonguing of Mauch Chunk and Pottsville facies in the upper member of the Mauch Chunk clearly indicates an alternation of depositional environments, but it is problematical whether this represents the lateral migration of two co-existing subenvironments in the sense of Walther's Law, or the sedimentological adjustment of an entire depositional system to cyclic climatic changes. In the former case, the Pottsville Formation would represent a higher elevation, proximal, more humid facies, subject to wetting more by flooding than by rainfall.

The interpreted tectonic and paleoenvironmental setting during Mauch Chunk deposition would have resembled in many respects the current alluvial plain extending from the Zagros Mountains to the Persian Gulf where arid conditions produce little clastic influx from the tectonically active mountain belt. The adjacent foreland basin axis--lying parallel to the mountain belt--receives primarily carbonate sedimentation. Were a future global climatic change to transform the Middle East into a humid region, the margins of the Persian Gulf could perhaps evolve into a broad peat-forming environment such as existed in the Appalachian basin during Pottsville and Llewellyn times.

SITE 2: MESOZOIC RIFT BASIN STRUCTURES READING TO POTTSTOWN, PA

Warren Manspeizer and Mark Lucas

LOCATION

Our field stops at this site are made in the Jacksonwald Syncline, along US route 422, from Reading to Pottstown, Pennsylvania (Fig. 6.2.1). The structural and tectonic setting of this site is given in Figure 6.2.2 and the stratigraphic position of our stops is given in Figure 6.2.3.

SIGNIFICANCE

We will examine the deformational structures that formed along the border fault of the Narrow Neck, or corridor, linking the Newark with the Gettysburg Basin, and relate these features to basin-forming processes. The Newark Basin, the largest of the exposed rift basins in eastern North America, is filled with Late Triassic-Early Jurassic fluvial-lacustrine and volcanic synrift strata that have been intruded by tholeiitic diabases and structurally deformed. Deformation, in part synchronous with deposition, began in the Carnian and ended in the Middle Jurassic with the onset of sea-floor spreading, and thereby may have been related to both basin-forming and basin-filling processes.

The Narrow Neck displays compressional (e.g. folds, cleavage, and thrust faults), extensional (e.g. veins and normal faults), and strike slip structures (Reidel shears) that may have formed in a predominantly left-lateral wrench or strike-slip zone, or perhaps in a sinistral transtensional zone (Lucas and others, in press). Note that in both the wrench and transtensional models, deformation in the Narrow Neck may be synchronous with the opening of the main Newark and Gettysburg rift basins. Thus strike slip in the Neck creates the transtensional Newark and Gettysburg Basins, and sinestral transtension in the Neck creates the extensional main basins.

Sedimentary basins forming within strike-slip regimes and along faults systems that are en-echelon, curved or otherwise offset (as in the Newark Basin), develop complex compressional or transpressional structures adjacent to, and within, regions of extension or transtension. Rocks in the Narrow Neck have these characteristics and are similar to those in the foreland of fold and thrust belts. The strata

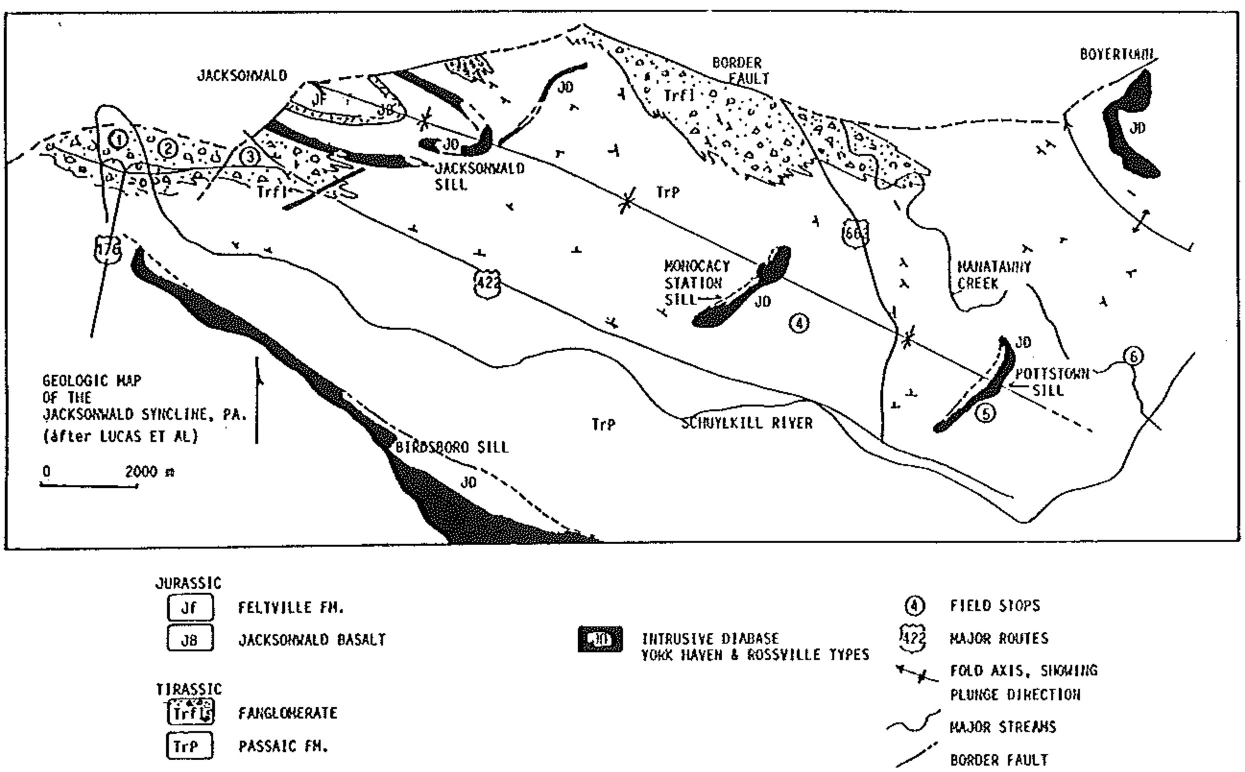


FIGURE 6.2.1 Generalized Geologic Map of the Newark Supergroup along U.S. route 422, from Reading to Pottstown, Pennsylvania. (After Lucas et al., In Press). Map also shows location of field trip stops on Day 6, Site 2.

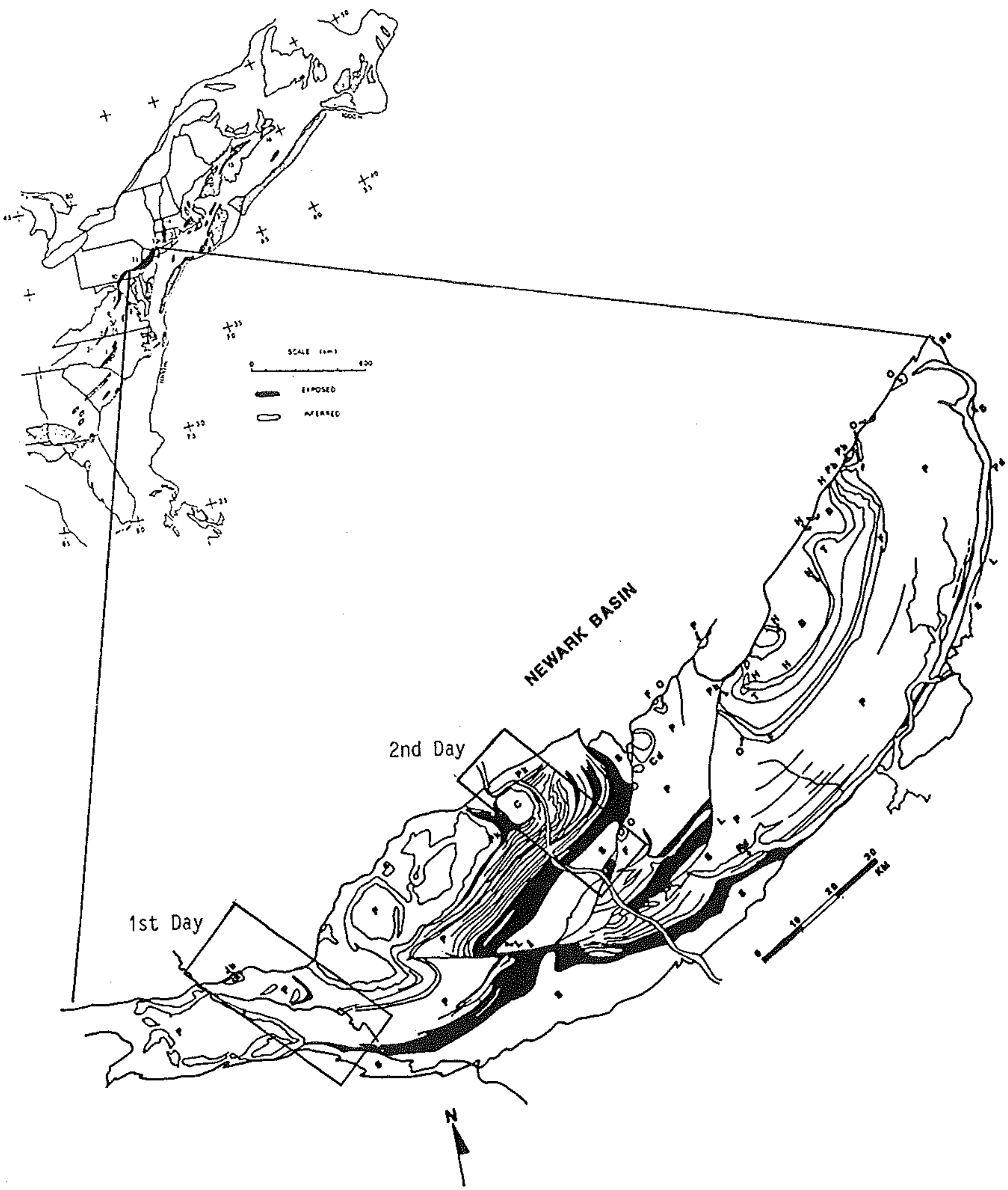


FIGURE 6.2.2 Geologic Map of the Newark Basin and eastern edge of the Narrow Neck, showing general location of stops for Days 6 and 7, and regional tectonic setting of the Eastern Mesozoic rift basins. See also Figures 6.2.1 and 7.1.1 for more detailed geologic maps and field trip routes. Geologic map of the Newark basin shows distribution of formations and clusters of detrital cycles (parallel black lines) in the Passaic Formtion (from Olsen, 1980). Abbreviation of formations and diabase bodies as follows: B = Boonton Formation; H = Hook Mountain Basalt; Hd = Haycock Mountain Diabase; Jb = Jacksonwald Basalt; L = Lockatong Formation; O = Orange Mountain Basalt; P = Passaic Formation; Pb = Preakness Basalt; Pd = Palisades Diabase; Pk = Perkasie Member of Passaic Formation; Rd = Rocky Hill Diabase; S = Stockton Formation; Sc = carbonate facies of the Stockton Formation; Sd = Sourland Mountain diabase; T = Towaco Formation.

typically dip about 10-20^{0} towards the basin-bounding faults, where they are deformed into broad open folds or more highly compressed into en-echelon folds, as exemplified by the Jacksonwald Synform and its related structures along the border fault.

Lucas and others (in press) report that the Jacksonwald Synform has characteristics of folds in the foreland fold and thrust belt. The shallow plunging (15^{0} towards 305^{0}) upright synform has a subangular hinge and straight planar limbs. Early formed clastic dikes were rotated towards the hinge, while mud cracks were stretched parallel to the hinge during buckling. An axial plane spaced cleavage is well-developed and maintains a constant orientation, suggesting late-stage formation. Fibrous quartz and calcite formed in veins, oriented sub-perpendicular to the fold axis.

These workers also report that the strata have been deformed by three sets of steeply dipping, predominantly strike-slip faults. West-northwest-striking sinistral faults along the border and in the basin are coeval with north-striking minor faults (showing both senses of shear). These early faults are cut by northeast-striking sinistral and dextral faults. The population of minor faults produced bulk strain with subhorizontal northeast-shortening and northwest extension, essentially coaxial with the fold structure.

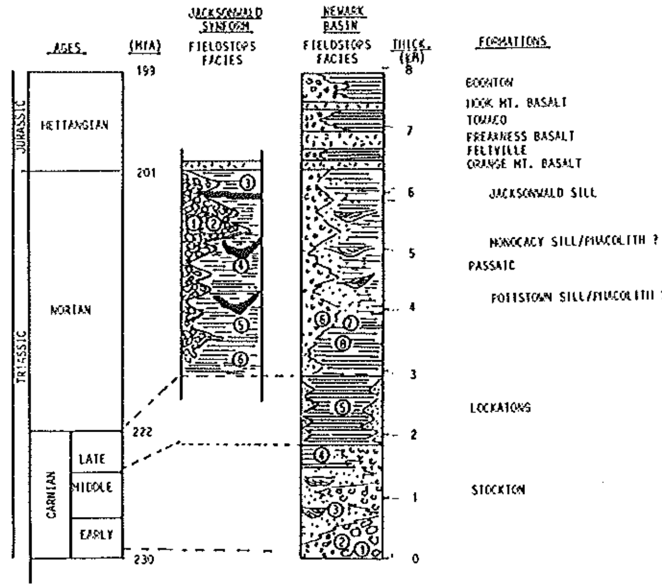


FIGURE 6.2.3. Stratigraphic column of the Jacksonwald Synform and the Newark Basin, showing stratigraphic location of fieldstops for days 6 and 7. Age and thickness data from Olsen (Written comm., 1988).

SITE DESCRIPTION

Outcrop 1: Mesoscopic structures in Triassic border conglomerate

This outcrop lies about 250 meters south of the northern border fault. The conglomerate occurs stratigraphically high up in the Passaic Formation, and is composed primarily of clast-supported, sand-to-boulder-size carbonate detritus with red-brown sandy matrix, and minor interbeds of sandstone. The objective at this stop is to examine the chronology of tectonic events that occurred along the border faults. The mesoscopic structures are described here in chronologic order, inferred from cross-cutting relationships.

1) Up to 50% shortening, due to pressure solution, can be viewed here by sutured and deeply embayed clast boundaries (Fig. 6.2.4). The greatest amounts of embayed and sutured contacts occur along clast boundaries which are perpendicular to bedding. This general trend of pressure solution deformation indicates that shortening was subparallel to bedding, and thus was due primarily to tectonic strain and not to lithostatic loading.

2) Pressure solution surfaces are cut by evenly spaced (10 cm) subparallel, gently dipping contraction faults. The faults are defined by offset clasts along straight or slightly anastomosing fractures. The fractures are usually thin (7.5 cm) and not filled, although some contain small amounts of gouge and tectonic breccia. Maximum offset observed along the contraction faults is approximatey 5 cm. Slickenside surfaces are not observed at this location so that the exact direction of movement cannot be determined.

3) Pressure solution surfaces and contraction faults are cut by northeast-striking veins, ranging in width from 2 mm to 1 cm that are filled solely with calcite. Microscopically, the calcite appears as large equant, euhedral crystals exhibiting mechanical twinning. A small overhang at the outcrop reveals an updip view of these veins, which are spaced approximately 10 cm apart. A second set of short gash-shaped veins, arranged in an en-echelon left-stepping array, occur within the areas bounded by the first set of veins (Figure 6.2.4). The second set of veins seem consistant with northeast directed dextral shear.

Outcrop 2: Northern Border Fault at Big Dam Quarry

At this outcrop a west-northwest striking segment of the border fault juxtaposes Triassic fanglomerates against Lower Cambrian carbonates. The fault zone is defined by thin (2 to 3 cm thick), steeply dipping, sheet-like horses of fanglomerate bounded by slickensides and narrow zones of cataclasite. The fault zone is at least 15 m wide in the hanging wall (the width of the quarry exposure), with the intensity of brittle deformation decreasing towards the southwest. The fault consists of 2 to 3 m of pebbly breccia with a gouge matrix. The Cambrian carbonates are deformed into cataclastic breccias adjacent to the fault, though deeply embayed clasts in the fanglomerate suggest pressure solution accompanied the brittle deformation. The slickensides are marked by subhorizontal "tool and groove" type slickenlines (corrugations) and a faint mineral streaking

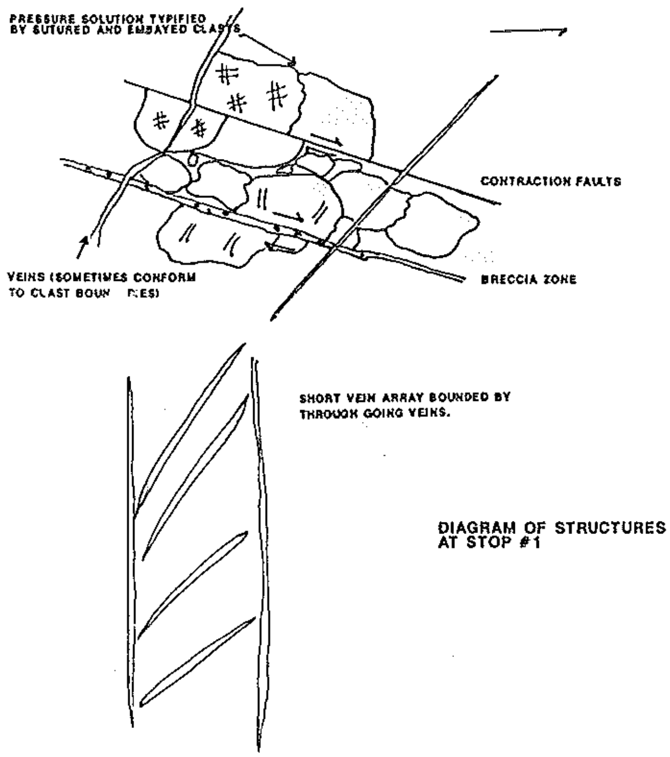


FIGURE 6.2.4 Sketches from Outcrop 1 (Site 2, Day 6) of pressure solution surfaces cut by contraction faults that are in turn cut by northeast-striking veins.

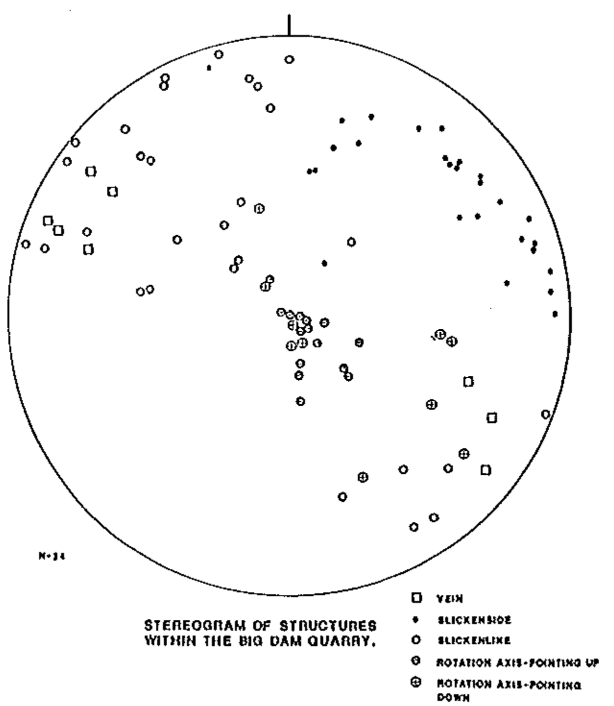


FIGURE 6.2.5. Stereogram of structures within the Big Dam Quarry (Outcrop 2, Site 2, Day 6).

is present on some surfaces.

Based on the orientation of the horses and their bounding faults, this border fault segment dips 70^{o} towards 210^{o} (Fig. 6.2.5), and therefore strikes subparallel to the trend of the synformal axis. The girdling of poles to slickensides reflects the anastomosing character of the fault zone. Based on the orientation of partially girdled slickensides, the direction of net slip rakes less than 20^{o} on the fault plane. Clasts within the fanglomerate are offset in a sinistral sense in map view. Using a right hand rule convention and rotation axis, the rotation axes for this fault are steeply plunging and upward pointing (Figure 6.2.5). The bulk deformation is therefore one of plane strain, simple shear in a left lateral strike slip zone. The fault zone is cut by small, subhorizontal contraction faults and steeply dipping, calcite filled veins that are mutually crosscutting.

Triassic limestone conglomerate, cropping out about 80 m southwest of the border fault, exhibits a general decrease in pressure solution deformation and fracture density with increasing distance from the fault. The conglomerate consists of Early Paleozoic limestone and dolomite clasts in a matrix of red-brown, fine-grained sand and silt. The conglomerate is notably clast-supported near the border fault, and matrix-supported away from the fault.

Mesoscopic structures here include brittle deformation fabrics in the form of small faults, slickenside-coated shear fractures and veins. In addition pressure solution features, such as embayed and sutured clasts, occur within the conglomerate. Three minor sets of faults cut the conglomerate. Two sets, seem to exhibit an Anderson-type conjugate.

The first set strikes N80^{o} W and dips steeply to the southwest (Fig. 6.2.6). This set exhibits a sinistral sense of shear, defined by offset clasts in map view and slickenline direction. The second set of faults, is mutually cross cutting with the first, striking N20^{o} W and dipping to the northeast. This set exhibits a dextral sense of shear in map view. Both fault sets exhibit thin (1 mm to 1 cm) zones of deformation defined by tectonic breccia or tool and groove type slickensides. Areas between the zones are undeformed.

A third set of minor faults is defined by gently, southwest-dipping contraction faults, which are similar in morphology and orientation to the faults observed at Outcrop 1. The contraction faults are not as abundant, or as closely spaced, as those found at the previous outcrop, and thus are not a major portion of the bulk strain at this locality.

Both the contraction faults and conjugate shear fractures are transected by calcite-filled veins, which strike N30^{o} E and dip steeply to the southeast. The veins range in thickness from 1 mm to 2 cm, vary in spacing from 1 to 10 m, and cut both clasts and matrix. Vein density increases away from the border fault, with the largest density of veins along the southern reaches of the outcrop. A second, less abundant, set occurs at the southern end of the outcrop. These veins are oriented subparallel to bedding and are mutually cross cutting with the first set. This second set is relatively thin (1 mm to 5 m) and randomly spaced.

Outcrop 3: Syndepositional Folds

This outcrop, exposed at a residential construction site, consists of about 60 cm of tan-buff, micaceous sandstone that occurs about 3 m below the Jacksonwald Basalt (top of Passaic Formation). The bed, situated on the south limb of the syncline, dips steeply (80^{o}) to the northeast, and is internally deformed into a series of minor antiforms and synforms. The fold limbs appear planar with rounded hinge zones. The folds are tight to isoclinal and appear asymmetric with fold axes that trend and plunge subparallel to the Jacksonwald axis. Axial surfaces range from moderately inclined (60^{o}-30^{o} dip) to recumbent (10^{o}-0^{o} dip).

These minor folds, the product of layer-parallel shortening, probably formed due to syndepositional compression, prior to the main folding event. Their sense of asymmetry is consistant with layer parallel slip, which accommodates the buckling of beds in a syncline.

Outcrop 4: Spaced Cleavage in the Passaic Formation

Although reported from the Fundy Basin (Latjai and Stringer, 1979), spaced cleavage has not been previously described in the Newark Basin (Lucas, 1985). The cleavage is expressed in outcrop by a characteristic fabric of penetrative anastomosing fractures; weathering of the cleavage produces a scree of short stubby pencil-shaped objects. Cleavage surfaces are discontinuous and spaced approximately 1 to 3 cm apart. Mineral lineation or slickenlines have not been observed on the cleavage surfaces. In thin section, the cleavage is defined by dark, opaque and phyllosilicate-rich seams usually 1 mm thick. The seams are roughly 2 to 3 cm in length and terminate in feathery splays. Large quartz crystals are often sutured along irregular boundaries that parallel the cleavage. Microlithons between the cleavage seams occasionally exhibit crenulated or disturbed bedding, though bedding shows little displacement across the seams.

At this location the pencil habit of the cleavage is confined to the shaly-siltstone fabrics, although the dark feathery cleavage seams are found in coarser grained rocks such as fine sandstones.

Sinks for the dissolved minerals are not known, and overgrowths or fibrous beards are not present. Veins containing fibrous quartz are common at the site, and commonly cut the cleavage at high angles.

Solution surfaces usually form perpendicular to the principal shortening direction (Groshong, 1975), indicating horizontal shortening along a 35^{o} azimuth for the cleavage forming event. Reks and Gray (1982) have shown that pencil cleavage in mudstones of the Valley and Ridge Province represents shortening values between 10 and 25%, where strain increases with increasing aspect ratio of the pencils. Short pencils in the Jacksonwald synform suggest shortening strains of about 10% ($e = .1$) normal to the cleavage.

The cleavage is vertical and uniform in orientation throughout the fold (Fig. 6.2.7), and is not fanned about the fold axis. The cleavage is partially girdled about a vertical axis, which reflects the change in trend of the fold axis along the hinges. The absence of fanning suggests the solution cleavage developed in the later phases of folding, rather than during early layer parallel shortening. If the fold is symmetric, then the cleavage is subparallel to the axial plane,

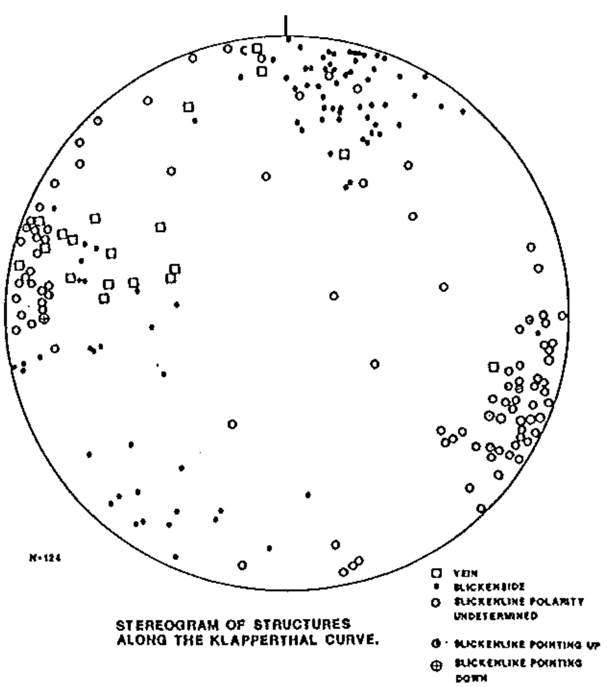


FIGURE 6.2.6 Stereogram of structures along the Klapperthal Curve at Big Dam Quarry (Outcrop 2, Site 2, Day 6).

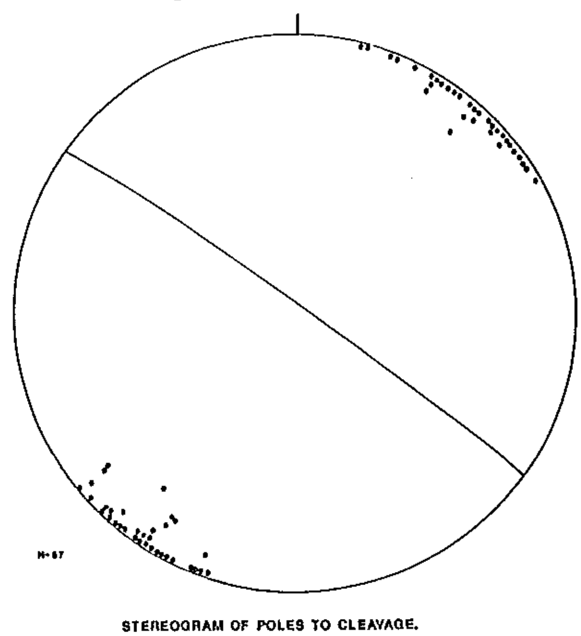


FIGURE 6.2.7 Stereogram of poles to cleavage at Outcrop 4 (Site 2, Day 6).

however, the orientation of the axial plane is not independently known, and the fold may be transected by the cleavage.

Outcrop 5: Tectonic Structures, Hornfels, and Sill at Pottstown Quarry

The walls of the Pottstown Quarry reveal a 45 m section of stratified hornfels in the Passaic Formation that has been baked by the overlying Pottstown Sill. The Pottstown Sill, composed of York Haven type diabase (Smith, et al., 1975), resembles a phacolith that was intruded along the hinge of the synform. Both the Pottstown and Monocaty Station sills (located up section within the Passaic Formation) may have been intruded either before folding or, perhaps during folding ("Saddle reefs"). Exposures of the "phacoliths" are poor and their relationships to the different structural elements within the syncline have not been assessed at this time.

Mesoscopic structures within the quarry are expressed in two large fault zones. The fault zones range in width from 20 to 40 m. The 40 m fault zone thins rapidly to the south, and is 5 m thick in the south wall of the quarry. The fault zones consist of steeply dipping horses of cataclasized hornfels bounded by zones of clay gouge. Bounding surfaces are marked by both highly polished, "tool and groove" type slickenlines (with some epidote) and chlorite fibers.

The fault zones dip 70^{o} towards 35^{o} (Fig. 6.2.8) and exhibit subhorizontal lineations which rake gently to the north. Steps on chlorite slickenlines and offset gas vesicles define a uniform, dextral sense of shear. Given the predominance of the single shear sense, this minor fault system does not represent an

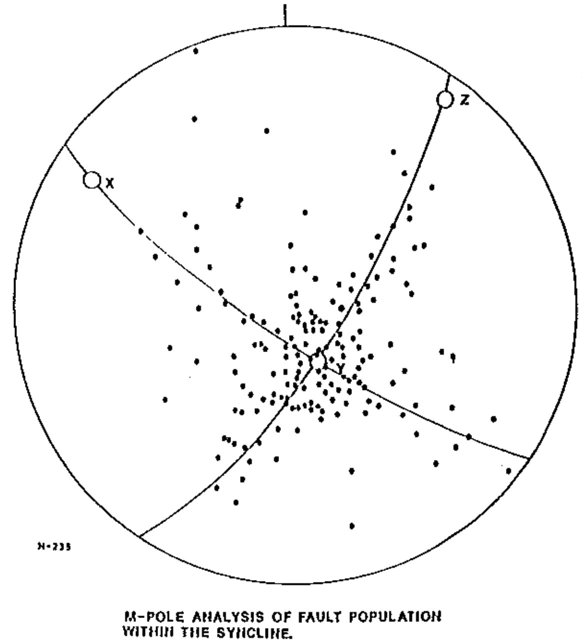


FIGURE 6.2.9 M-pole analysis of fault population within the syncline at Pottstown Quarry (Outcrop 5, Site 2, Day 6).

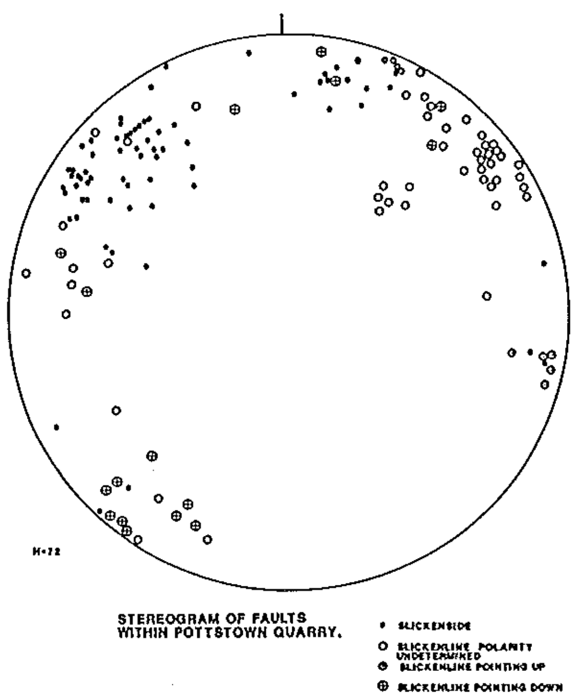


FIGURE 6.2.8 Stereogram of faults within Pottstown Quarry (Outcrop 5, Site 2, Day 6).

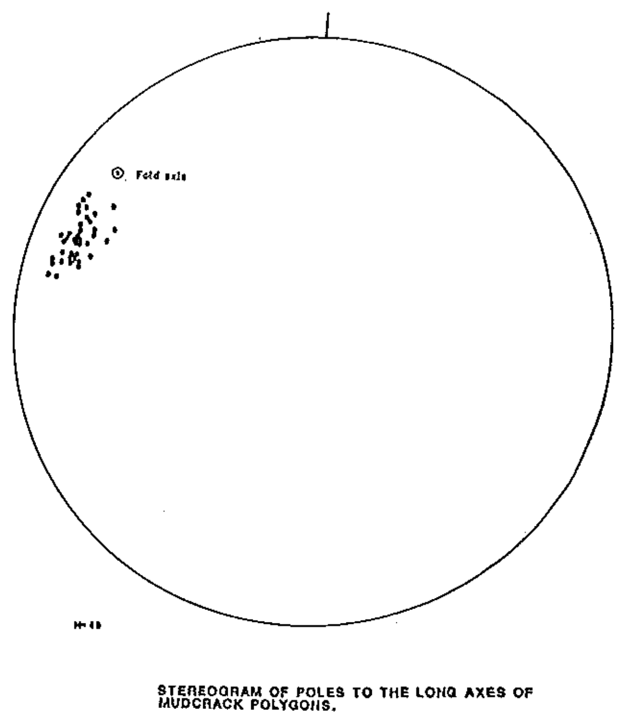


FIGURE 6.2.10 Stereogram of poles to the long axes of mudcrake polygons at Manatway Creek (Outcrop 6, Site 2, Day 6).

Anderson-type conjugate, as was seen on the Klapperthall curve at Outcrop 2.

The bulk strain produced by the minor fault population within the synform (Klapperthal curve, Pottstown Quarry, etc.) is difficult to assess. The diffuse point maximum of steeply plunging rotation axes suggests a plane strain deformation (Fig. 6.2.9), with the intermediate principal strain direction (Y) parallel to this maximum. The rotation axes (or "M-poles", poles to movement planes) also define two perpendicular girdles, and the principal diretions of shortening (Z) and extension (X) lie at 90^{0} Y on either girdle (Arthaud, 1969). The distribution of minor faults and rotations suggests northwest directed extension and northeast directed shortening. Unfortunately, the fault population may not represent a single bulk deformation. Cross cutting relationships among the different sets in this fault population are seldom seen, and the faults may have developed at different times. In addition, the amount of displacement or net slip on each of the fault sets is not well known. Preliminary measurements suggest that slip on the east-striking sinistral faults is about 5 times that of the north- to northeast-striking sets (Lucas, 1985); therefore the faults should not be equally weighted in the strain analysis.

Outcrop 6: Penetrative Strain in the Passaic Formation at Manatway Creek

The floor of the quarry is subparallel to bedding and exhibits three distinct penetrative fabrics, namely: spaced cleavage, fiber filled veins, and a uniquely-formed shape lineation of mudcrack polygons that have been deformed into ellipses. Found primarily within the hinge area, the lack of deformed mud polygons on the limbs may reflect poor exposure.

The long axes of the polygons are subparallel to the fold axis, the cleavage trace and poles to veins, but are not strictly coaxial with these structural elements (Fig. 6.2.10). The average ellipse ratio is 21 and does not vary with lithology. The mud cracks record a small amount of penetrative strain (grain scale) within beds.

DAY 7

SITE 1: MESOZOIC RIFT BASIN DEPOSITS ALONG THE DELAWARE RIVER, STOCKTON TO MILFORD, NJ

Warren Manspeizer and Mark Lucas

LOCATION

Intrabasinal faults have cut the Newark Basin into three major fault blocks (Fig. 6.2.2), repeating the stratigraphic section three times. Today's traverse examines the synrift strata (Fig 6.2.3) on the western fault block, along the Delaware River from Stockton to Milford, N.J., a distance of about 30 km on NJ route 29 (Fig. 7.1.1). See Van Houten (1969 and 1980) and Manspeizer and Olsen (1981) for additional field descriptions.

SIGNIFICANCE

This section is considered by many workers to be the classic basin fill for the East Coast Mesozoic basins. The sequence (Fig. 6.2.3) begins with a fluvial conglomerate, the Stockton Formation (1840 m thick), whose upper members seem to grade laterally into lacustrine mudstones and carbonates of the Lockatong Formation (1145 m thick) and vertically into fluvial-lacustrine mudstones and sandstones of the Passaic Formation (about 2000 m thick). Time-transgressive border conglomerates of the Hammer Creek Formation interfinger with the above formations along the western border fault, and a diabase intrudes the sequence along the basin axis. In the northern part of the basin, Triassic strata are overlain by a thick Early Jurassic succession of tholeiitic lava flows and fluvial-lacustrine clastics that, prior to erosion, probably extended to the Delaware River Section.

The synrift strata rest with a profound unconformity on Precambrian and Paleozoic basement rock of the Piedmont Province along the southeastern margin of the basin, and are in fault contact with Precambrian and Paleozoic rocks of the New England Upland Province along its northwestern margin. The sequence is wedge-shaped, being thicker and coarser-grained along its western margin and finer grained and thinner along the eastern hinged margin. Conglomerates, however, are common around the margins of the basin and thin out towards the basin center. Axial deposits are substantially finer-grained and consist of lacustrine mudstones and carbonates. Paleocurrent directions yield a pattern of transverse infilling from the northwest and southeast, and axial infilling from the northeast. Maximum clasts sizes typically increase towards the margins of the basin with the largest clasts occurring along the northwest fault margin.

Proprietary seismic lines show, through sedimentary overlap, that the depositional centers have migrated

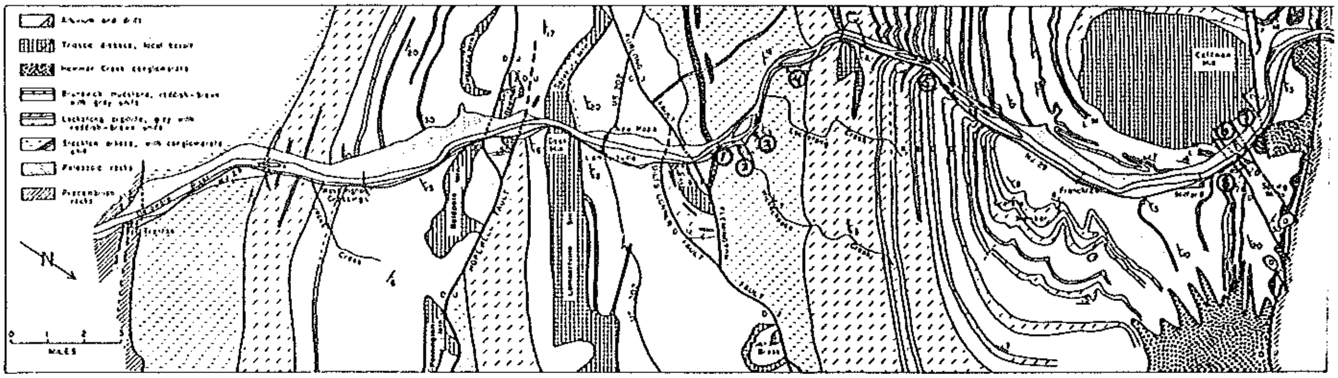


FIGURE 7.1.1 Geologic Map of the Newark Supergroup along the Delaware River, west central New Jersey and Pennsylvania (from Van Houten, 1969, 1980). Map also shows location of field trip stops at Day 7, Site 1.

away from the basin axis, so that the stratigraphic thickness is greater than the maximum thickness of strata at any one place. Basin migration in this context is presumably by growth faulting along the southeastern margin (see Cloos and Pettijohn, 1973; Van Houten, 1980), and by listric faulting along the northwestern margin of the basin.

Newark strata along the Delaware River section have been studied for many years (see references in Manspeizer and Huntoon, this volume), and while important issues go unresolved (e.g. facies vs. formation), the major patterns of sedimentation are fairly well understood. The depositional model, presented here and tested in our field trip, is a modification of the one by Turner-Peterson (1980). In general, debris flow alluvial fans dominated the northwest margin of the basin, while debris and streamflow alluvial fans and lacustrine-deltas dominated the southeastern margin. The axial part of the basin was dominated by a exceedingly large and moderately deep lake that was fed primarily by perennial streams flowing along the axial margin of the basin. Fluvial sedimentation occurred on a broad sandy alluvial plain there, while lacustrine and marginal shoreline sedimentation occurred in a large and moderately deep-water lake that underwent major and minor episodes of transgression and regression. The basic factors controlling facies distribution were the geometry of the basin and fluctuations in lake level.

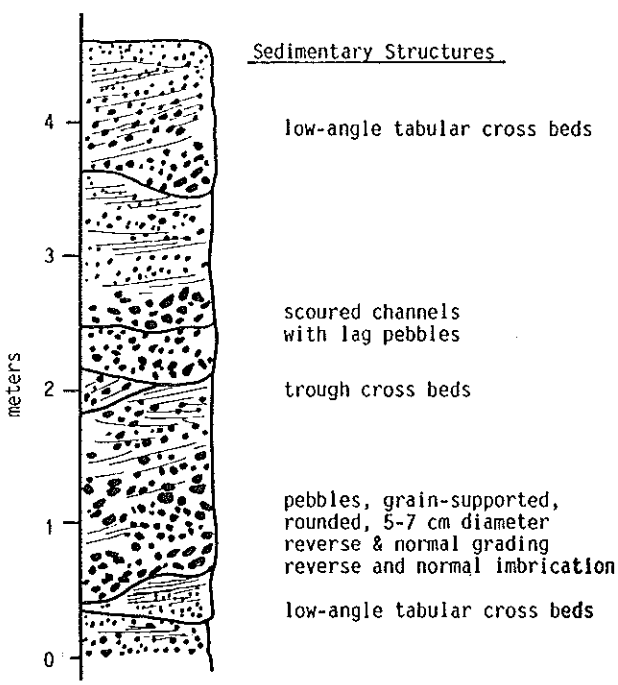


FIGURE 7.1.2 Stratigraphic column of proximal midfan deposits, Solebury Mbr., Stockton Fm. (Outcrop 1, Site 1, Day 7).

SITE DESCRIPTION

Outcrop 1: Proximal Midfan Deposits, Solebury Member, Stockton Formation

This section lies unconformably about 200 m above the Paleozoic limestone basement (Allen, 1979). The outcrop is marked by multiple scoured surfaces and deeply eroded conglomeratic sequences (Fig. 7.1.2). The bedrock consists primarily of white and gray, very coarse-grained quartz-arkose conglomerates with cobbles, averaging about 5-7 cm, but ranging up to 14 cm in diameter. Arkosic sands and fine gravels (1-3 mm) comprise the matrix for the conglomerate. The beds are commonly graded, but well-stratified, and contain well-rounded pebbles that are both matrix-and-clast-supported. Pebble lag concentrates, within scoured pockets, often are imbricated to the west, which is contrary to the general paleocurrent trend. Scattered pebbles occur throughout the bed, and in places show reverse grading. Large-scale tabular cross-stratification is common and yields paleocurrent data, indicating west and northwest flow. A modal analysis (Allen, 1979) of sandstone shows 63% quartz, 31% plagioclase and 2% orthoclase, indicating a granitic or gneissic source rock to the southeast.

Outcrop 2: Distal Midfan Deposits, Solebury Member, Stockton Formation

This section (Fig. 7.1.3) of the Solebury Member consists of coarse-grained pebbly sand to very fine-grained (3-4 mm), well-rounded, feldspathic-rich quartz gravel with scattered and rounded pebbles up to

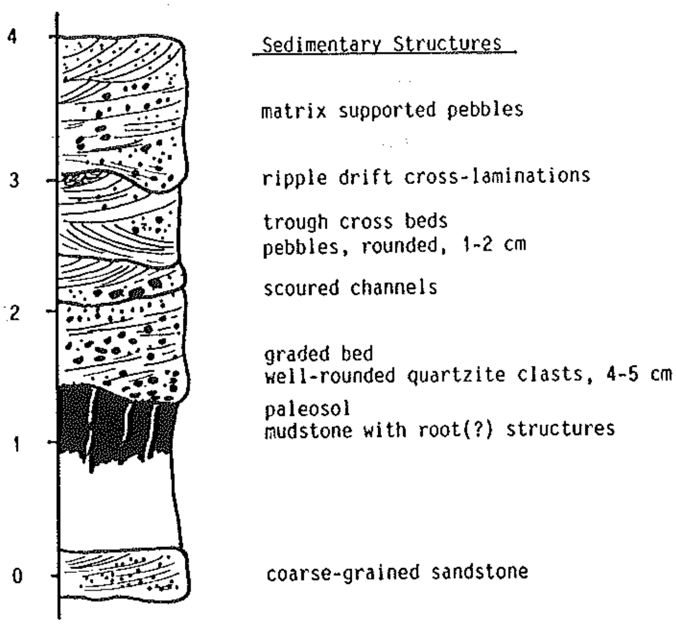


FIGURE 7.1.3 Stratigraphic Column of distal to midfan deposits, Solebury Mbr., Stockton Fm. (Outcrop 2, Site 1, Day 7).

8 cm in diameter. Multiple graded cycles of scoured basal contacts with lag pebbles are common, and grade upward to finer sands with festoon-cross-beds that are overlain by ripple-drift cross-lamination. Although poorly exposed, red-brown mudstones with root structures of the overbank deposits (?) occur at the base of the section. Paleocurrent measurements indicate west and northwest current directions.

Differences in Outcrops 1 and 2 are found in average and maximum grain size, types and sizes of cross-bed sets, and the 'completeness' of the depositional cycle. Whereas Outcrop 1 is interpreted as proximal midfan, this Outcrop is interpreted as distal midfan.

Outcrop 3: Fluvial Deposits, Prallsville Member, Stockton Formation

In this quarry, about 2 km north of Stockton, the Prallsville Member is 60 m thick and about 760 m above the base of the Stockton Formation. The section (Fig. 7.1.4) is difficult to interpret because soil-forming processes, manifested by intensive bioturbation by large and small plant roots and by the arthropod Scoyenia have destroyed, or otherwise obscured, many earlier formed primary structures. Both structures also attest to the high organic productivity in shallow water, fluvial and/or deltaic, environment. Secondly, it seems that the fluvial system, which laid down this sequence, may have meandered through an earlier channel fill, thus juxtaposing diachronously formed features. Although the base of the fluvial cycle is not seen in section, the shale chip conglomerates are interpreted as lag concentrates laid down in the main channel. Lateral accretion in point bars is recorded by the lower arkosic sandstones, which grade upward from medium-to fine-grain size, have thinly-stratified low angle planar beds, and hint of large-scale trough cross-beds. Reddish-brown siltstones with mudstone interbeds, trough cross-beds, rippled surfaces and ripple-drift cross lamination, comprise much of the section, and mark the upper part of the point bar. Vertical accretion on flood plains is marked by extensively burrowed red-brown, micaceous and calcareous mudstones with root traces, concretions and thin, ripple-marked siltstone lenses. The basal sands of a younger fluvial sequence are seen at the top of the quarry wall, and are coarse-grained, pebbly (2-5 cm), and buff-colored with large-scale trough cross-bedding. The paleocurrent trend is to the northwest, and consistent with progradation direction of the older alluvial fans.

Outcrop 4: Lacustrine Delta Deposits, Raven Rock Member, Stockton Formation

The Raven Rock Member at this stop occurs about 1300 m above the basal Stockton Formation. Thin interbeds of dark gray siltstone and sandstone that gradually, and almost imperceptibly, coarsen upward into massively-bedded sandstones, characterize this section (Fig. 7.1.5). The ideal and most complete

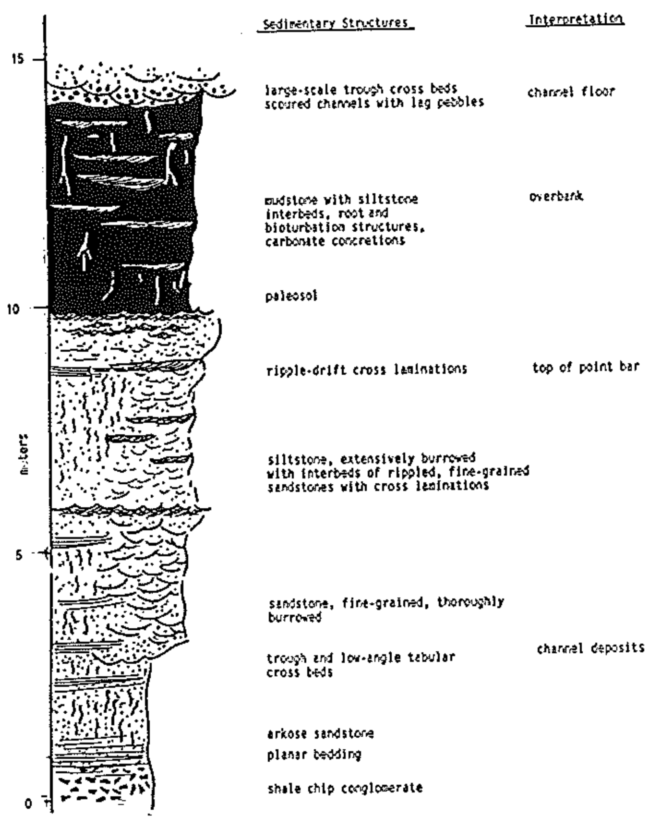


FIGURE 7.1.4 Stratigraphic Column of fluvial-deltaic deposits of Prallsville Mbr., Stockton Fm. (Outcrop 3, Site 1, Day 7).

sequence, seen at this outcrop, consists of: a finely laminated black shale; thin interbeds of dark gray siltstone and sandstone with small trough cross-bedding; and dark gray, calcareous, fine-grained, massively-bedded sandstones with root structures, large elliptical voids of carbonate nodules (?), convolute bedding and both small- and large-scale trough cross-bedding. Coarse-grained arkosic sandstones, seen as float in the field, probably comes from the upper part of this depositional cycle.

The section is notable because of the coarsening-upward deltaic-type sequence. The absence of inclined foreset beds, however, leads us to conclude that sedimentation was dominated by hypopycnal flow, i.e. where the river water is less dense than the basin water. Under these conditions fine sediment is transported in suspension some distance from the river mouths and forms deltas fronts sloping at about 1^{o}, in contrast to most Gilbert-type deltas, which slope between 10^{o}-20^{o} and are the product of homopycnal flow. High density lacustrine waters, envisioned for this setting, may be a consequence of high evaporation rates in closed basins, as in the Dead Sea. Large voids, presumably of weathered carbonate nodules, lend support to this thesis. Slightly higher water densities for Newark-type closed basins may explain why Gilbert-type deltas are rarely identified in this synrift sequence.

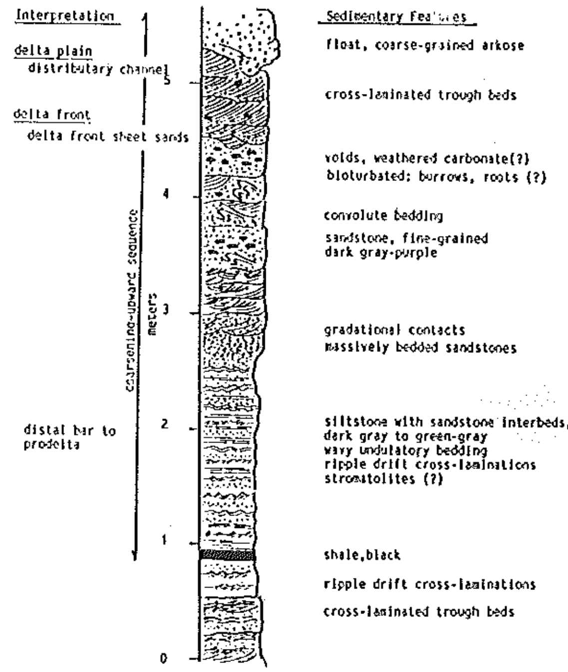


FIGURE 7.1.5 Stratigraphic Column of lacustrine delta deposits in Raven Rock Mbr., Stockton Fm. (Outcrop 4, Site 1, Day 7). Section generalized from a detailed measured section by H. F. Houghton and K. W. Muessig.

Outcrop 5: Lacustrine Cycles, Upper Lockatong Formation

This long road cut in the Upper Lockatong Formation and Byram diabase sheet is described in detail by Van Houten (1962, 1969, and 1980). The section, showing well-marked sedimentation cycles, termed Van Houten cycles by Olsen (in press, b), can be grouped into 25 m and 100 m cycles. The log and following description by Olsen (Manspeizer and Olsen, 1981), show that a typical cycle can be divided into 3 lithofacies or divisions, that correspond to the transgressive, high-water stand, and regressive phases of Lake Lockatong (Fig. 7.1.6; see discussion in Manspeizer and Huntoon, this volume). A comparison of cycles at this outcrop records a broad range in sedimentary features that, in a two-member system, corresponds to Van Houten's chemical and detrital cycles. Each end member will be examined in the field.

At this roadcut detrital cycles average 4.5 to 6 m thick, and are divisible into: a) a lower massive to platy dark gray siltstone; b) laminated black siltstone that commonly has very small burrows and delicate shrinkage cracks; and c) massive-bedded feldspathic siltstone and fine-grained sandstone that are intensively disrupted by shrinkage cracks and show a variety of sole marks, including reptile footprints (Olsen, in press, b).

Chemical cycles averages 5 m thick, and consist of two major varieties. One type resembles detrital cycles in having a lower black or gray platy siltstone. The middle parts of this cycle contain massive beds of

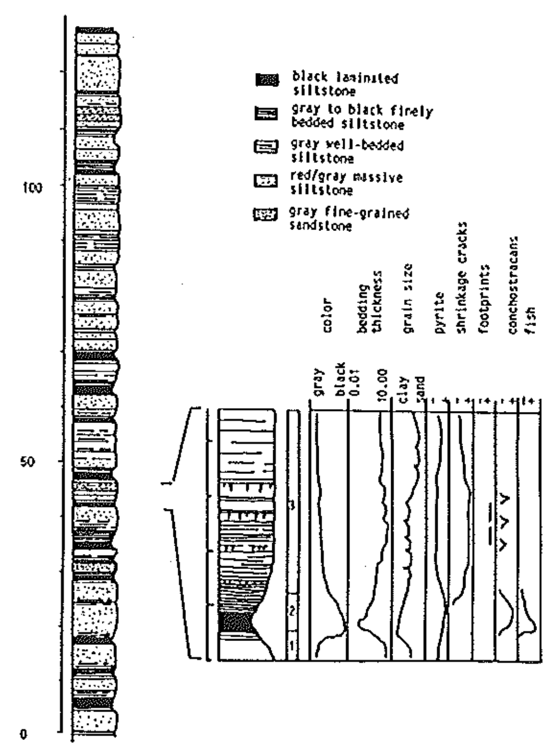


FIGURE 7.1.6 Stratigraphic Column of lacustrine cycles in the Upper Lockatong Fm. (Outcrop 5, Site 1, Day 7). Section modified from Manspeizer and Olsen (1981).

gray very hard and blocky siltstone (argillite) speckled with analcime and dolomite. The other variety has a lower platy gray to gray-green often dolomitic siltstone that is usually disrupted by shrinkage cracks. This basal unit grades upward into a blue-gray or red massive siltstone with abundant analcime.

Outcrop 6: Hammer Creek Fanglomerate at Pebble Bluffs

This conglomerate, exposed in a steep roadcut at the RR milepost 37 and less than 8 km from the border fault, occurs within a southward projecting lobe of quartzite-rich fanglomerates that pinch out rapidly to the south into a distal finer grained facies and coarsen upsection, indicating fan progradation with time (Van Houten, 1969; Arguden and Rodolfo, 1986).

The conglomerate (Fig. 7.1.7) is arranged in multiple cycles of crudely fining-upward sequences about 3-5 m thick, wherein clast-supported pebbles occupy deeply scoured basal channels, and grade upward to coarse-grained pebbly sandstones that are overlain by medium-to-fine-grained sandstones and locally mudstones. Locally the upper half of these cycles is riddled with caliche nodules that, in places, have coalesced into large vertical rhizoconcretions that are about 2-4 cm wide and 50 cm long. Multiple cut-and-fill channels are vertically juxtaposed against each other, attesting to deep scour across the fan with partial erosion of the previous depositional cycles. The clasts, primarily Paleozoic quartzite with some dolomite, are well rounded and have maximum sizes that average about 7 cm, ranging up to 10 cm. Sheared, deeply embayed pock-marked, and half rounded pebbles are present within this section, according to Arguden and Rodolfo (1986). Whereas sheared and pocked-marked clasts indicate in situ post-depositional deformation, half-rounded carbonate clasts indicate that these pebbles underwent syndepositional deformation.

An alluvial fan setting is interpreted by many workers for this section. Arguden and Rodolfo (1986), for example, suggest that most of the coarse clastics were deposited during major floods, whereas the finer beds were laid down during declining flow from the same flood, or during episodes of quieter runoff. Clast-supported conglomerates are interpreted as reworked channel sands laid down as stream flood deposits, whereas coarse pebbly sandstones represent braided stream channel facies, and the finer-grained sandstones are interpreted as a sheetflood deposit that are normally succeeded by thin mudstones of the waning flood facies. Fan buildup was intermittent, as suggested by the calcrete paleosols, which record long periods of non-deposition in a warm arid climate, followed by intermittent flooding and erosion of the fan. Penecontemporaneous deformation and uplift along the border fault is recorded by coarsening-upward megacycles, and by clasts that are shattered, pock-marked and half-rounded.

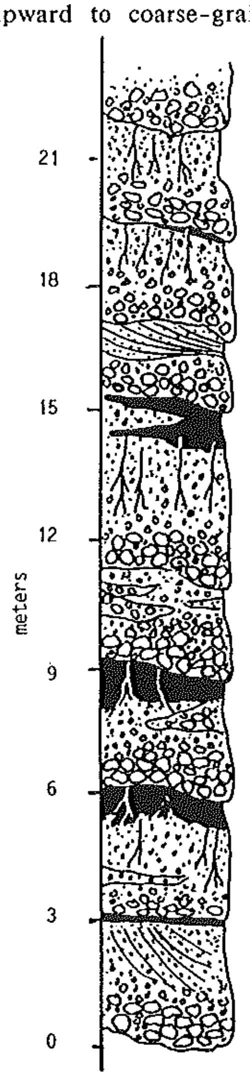


FIGURE 7.1.7 Stratigraphic Column of Hammer Creek Fanglomerate at Pebble Bluffs (Outcrop 6, Site 1, Day 7).

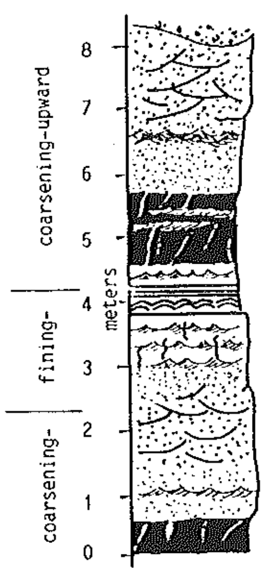


FIGURE 7.1.8 Stratigraphic Column of lacustrine shoreline deposits in the Passaic Fm. (Outcrop 7, Site 1, Day 7).

Outcrop 7: Lacustrine Shoreline Deposits, Passaic Formation

This section of alternating red-brown mudstones and siltstones with fine-grained sandstones occurs about 950 m above the base of the formation, and about 1.0 km east or basinward of the alluvial fan conglomerates of Outcrop 5, Site 2, Day 6 (Picard and High, 1963). Although these sections are in normal fault contact, conglomerate tongues of the Hammer Creek Formation (seen here about 3 m above the roadbed) attest to the general contemporaneity of these stratigraphic sequences (see Drake and others, 1961).

Preliminary studies indicate that the section consists of subtle coarsening-upward sequences (Fig. 7.1.8.), that record marginal shoreline progradation into a lake. Although the section is tentatively identified as deltaic, it has structures and sequence similar to prograding beach or barrier island. Multistory fluvial scoured surfaces are locally present and show that deep erosion, perhaps due to ephemeral lake levels, followed deltaic deposition. In general the contacts are ill-defined, except where lacustrine-deltaic sequences overlie fluvial deposits. There it is sharp, and marked by thinly-laminated siltstones and/or stromatolites(?). The onset of lacustrine transgression was rapid, and most likely due to flash flooding, or perhaps tectonic subsidence. Ripple-drift cross laminations are common; they are particularly noteworthy in the mudstones and fine-grained siltstones that are extensively bioturbated, indicating they formed in mudflats that became sites of high organic productivity. Large-scale trough cross-bedding in the fluvial sequence records westward fluvial-deltaic progradation and basin filling at the same time that alluvial fans advanced eastward across the basin. Deposition may have occurred in shallow areally extensive mud-filled playas.

Outcrop 8: Deltaic Cycles, Perkasie Member, Passaic Formation

Larger, and perhaps deeper, lakes existed during Passaic time. One of these lakes is recorded in the greenish-gray and red-bed sequence of the Perkasie Member at this stop, which is about 2 km east of, and stratigraphically only slightly higher than, the section at Outcrop 6, Site 2, Day 6 (see Drake and others, 1961). The Perkasie Member includes a sequence of recurring gray units in the upper part of the Passaic Formation in the Delaware River area. To the northwest the member interfingers with quartzite conglomerates of the Hammer Creek Formation. Here a tongue of the Perkasie interfingers with redbeds, showing as many as 13 deltaic cycles within a 60 m section, as described in Figure 7.1.9. Each cycle begins with mudstones that are interbedded with ripple drift cross-laminated siltstones, and grades gradually upward into massively-bedded siltstones that become coarser-grained and trough cross-bedded upward in section. Scoured distributary channel sandstones typically cap the cycle. The key to this section seems to rest with interpretations given to the origin of the overlying mudstones. Burrows and mud cracks suggest that these mudstones formed as interdistributary deposits. However, because they have sharp contact with underlying sandstones and gradational contact with overlying sandstones, it is thought that these mudstones formed on delta fronts that were subsequently exposed to the air as the lake level dropped.

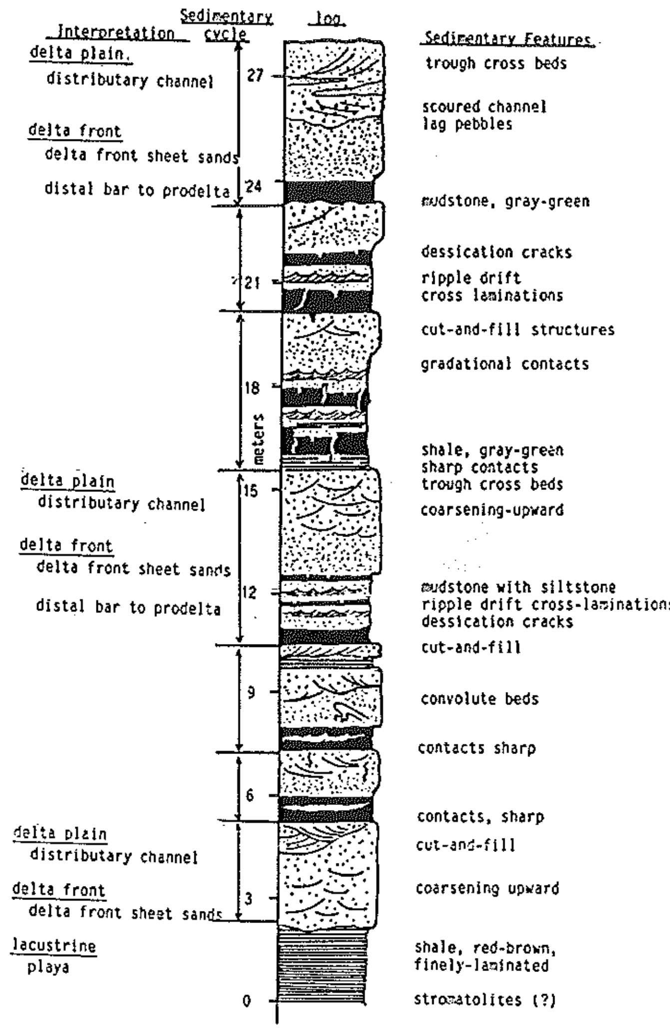


FIGURE 7.1.9 Stratigraphic Column of deltaic cycles within the Perkasie Mbr., Passaic Fm. (Outcrop 8, Site 1, Day 7).

DAY 8

SITE 1: PASSIVE MARGIN DEPOSITS AT CHINCOTEAGUE ISLAND, VA: THE HOLOCENE TRANSGRESSION AND RESULTING DEPOSITS

Rudy Slingerland

LOCATION

This site is located on the Delmarva Peninsula, a topographically low portion of the coastal plain separating the Atlantic Ocean to the east from the drowned valley of the Susquehanna River, Chesapeake Bay, to the west (Fig. 1 of the introduction to this volume and Fig. 8.1.1). The peninsula is underlain by up to 24 m (80 ft) of middle Pleistocene (184,000 years B. P.) unconsolidated sands, gravels, and lesser silts and clays (Mixon, 1985), and 2400 m of Tertiary passive margin fill. The Pleistocene strata are thought by Mixon (1985) to have been deposited by a transgressive barrier island system. Along the Atlantic margin of the penisula is the Holocene transgressive barrier island system, consisting of the islands of Chincoteague, Assateague, and Wallops and their associated inlets, marshes, and bays.

SIGNIFICANCE

The Holocene barrier system presents the opportunity for us to better understand the origin of the unconformity-bounded sequences that are so characteristic of passive margins. Here we can see the sedimentary processes that are responsible for the various facies and erosive surfaces of shore and backshore environments.

SITE DESCRIPTION

Outcrop 1: Fishing Point of Asseateague Island

We are standing on the largest recurved spit of the US coastline (Fig. 8.1.1). Radiocarbon dates of the underlying deposits by Goettle (1978) show that the islands and ridges in front of Chincoteague Island were deposited as a series of southward migrating spits over the last 2000 years, even as regional sea level rose 7 m over the last 4000 years (Belknap and Kraft, 1985). This is in contrast to barriers both north and south which experienced shoreface retreat of a few kilometers over that time (Belknap and Kraft, 1985) and serves as an example of the local variability that may be expected in a transgressive facies sequence.

Barrier sedimentary facies observed in trenches and vibra-cores include: 1) coarse-grained, large-scale planar and finer-grained, small-scale trough cross-stratified sands of offshore ridges and runnels, 2) low angle, planar-laminated, heavy mineral rich, medium sands of the swashface, berm and back berm environments. 3) well sorted fine sands of the dunes, shelly, heavy mineral rich, fine to coarse sands of washovers, and fine, bioturbated sands of the back barrier bay.

Offshore on the storm-dominated shelf of the Mid-Atlantic bight is a classic shoreface-connected sand

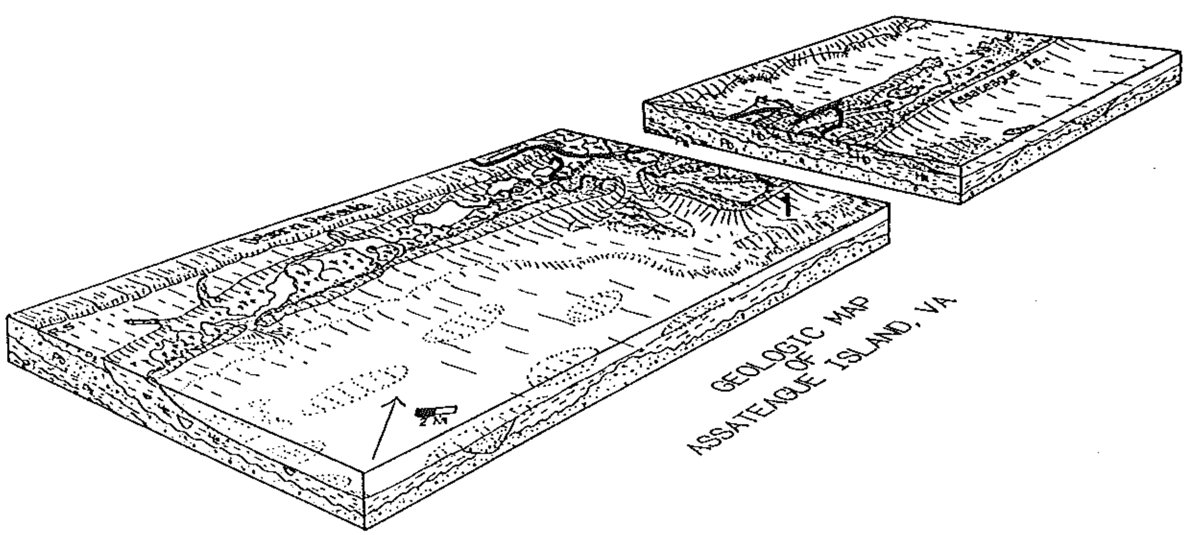


FIGURE 8.1.1 Two-point perspective block diagram of Assateague Island, VA area showing present geomorphology and Cenozoic deposits.

ridge complex composed of the "sawdust of marine planation" (Veatch and Smith, 1939) (Fig. 8.1.2). Just to the north off New Jersey, vibra-cores and seismic lines show that post-transgressive marine sands are typically less than 3 m thick and overlie eroded Pleistocene and early Holocene deposits (Niedoroda et al., 1985). The Holocene deposits are typically back-barrier bay, marsh, and tidal flat muds, and more rarely, channel or washover sands and gravels (Fig. 8.1.3). These overlie the Pleistocene deposits described earlier. Note that the subaerial portion of the barrier is rarely, if ever, preserved.

Outcrop 2: Leading Edge of the Holocene Transgression

At this "outcrop" we can investigate the back barrier facies of marsh, tidal creek, and bay, and examine the character of the Holocene-Pleistocene unconformity. A typical core in this area consists of homogeneous dark gray clayey silts with scattered Spartina stem and leaf fragments, and occasional beds of homogeneous, lighter gray, silty clays, containing shell fragments. The former are interpreted as salt marsh deposits and the latter as bay deposits. Depending upon location, at about 3.5 m a gray to reddish brown fine quartz sand stops penetration. Occasionally it contains granules and rare quartz pebbles. We interpret this to be either in situ or slightly reworked Pleistocene sands, and thus to define the unconformity.

The geometry of the unconformity deserves some discussion. Its trace in plan view is remarkably straight from here southward and in cross section it it steepens along its western boundary (Fig. 8.1.1). This has been interpreted by Demerest and Leatherman (1985) to arise from passive onlap of the mainland fringing marsh against a 60,000 year B. P. barrier shoreface. But mapping by Mixon (1985) discredits a continuous Pleistocene shoreface in this area. We prefer an interpretation wherein the feature is a degraded wave-cut cliff formed prior to 2000 yrs B. P. when the barrier islands were only partially emergent shoals (Morton and Donaldson, 1973). The surface itself is a subdued replica of the subaerial drainage net (Slingerland, 1977; Niedoroda et al., 1985).

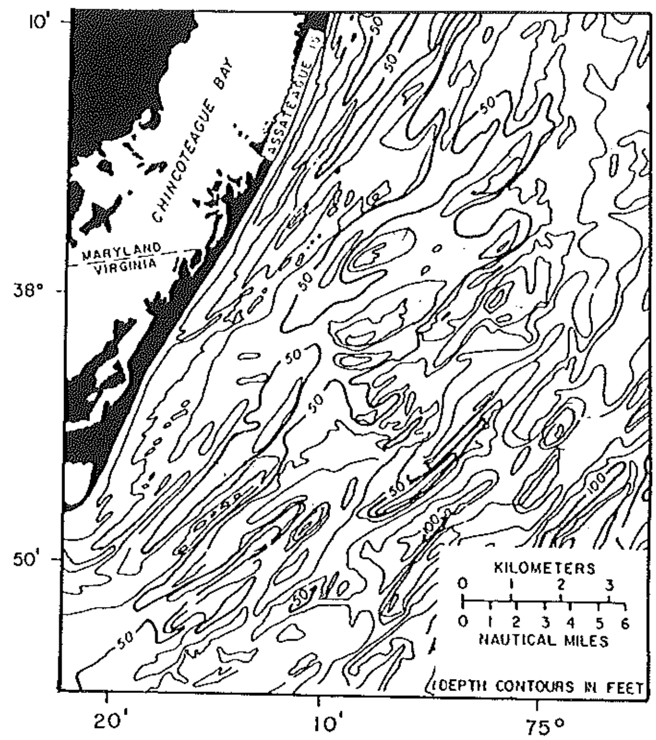


FIGURE 8.1.2 Bathymetric map of Atlantic continental shelf off Assateague Island, VA showing shoreface connected sand ridges trending obliquely offshore (modified from Niedoroda et al, 1985).

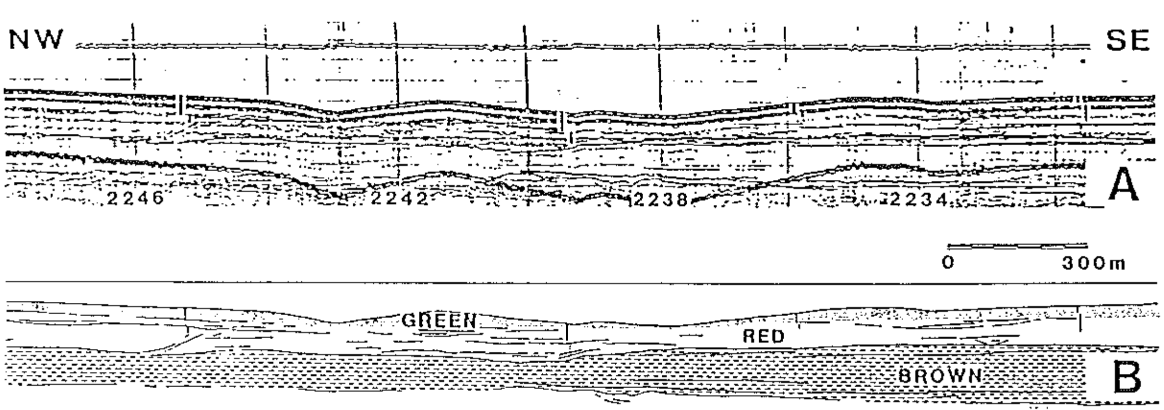


FIGURE 8.1.3 Seismic line and interpretation along a shore-normal transect off New Jersey, a setting similar to that depicted in Fig. 8.1.2. The brown unit consists of late Miocene strata. It is unconformably overlain by the red unit, consisting of shelly sands and thin-bedded alternations of sand and mud dating to 3,000 to 11,000 yrs BP and interpreted as Holocene back barrier deposits. The red unit in turn, is unconformably overlain by the green unit, consisting of a shelf sand making up the shelf ridges, and of similar age or slightly younger than the red unit (modified from Niedoroda et al., 1985).

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