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**Sequence stratigraphy and depositional framework of the Upper
Middle Devonian Mahantango Formation in Pennsylvania and
New York**

Slattery, William, Ph.D.

City University of New York, 1993

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**SEQUENCE STRATIGRAPHY AND DEPOSITIONAL FRAMEWORK OF THE UPPER
MIDDLE DEVONIAN MAHANTANGO FORMATION IN PENNSYLVANIA AND NEW
YORK**

by

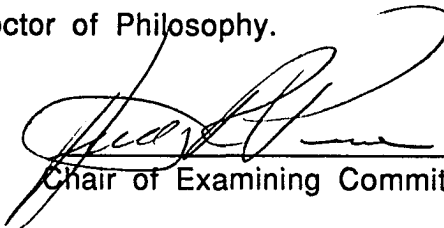
WILLIAM SLATTERY

A dissertation submitted to the Graduate Faculty in the Earth and
Environmental Science Department in partial fulfillment of the
requirements for the degree of Doctor of Philosophy, The City University
of New York.


1993

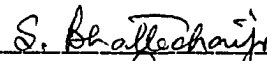
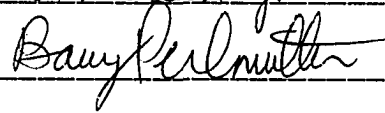
This manuscript has been read and accepted for the Graduate Faculty in Earth and Environmental Science in satisfaction of the dissertation requirement for the degree of Doctor of Philosophy.

4-15-93
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Chair of Examining Committee

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Executive Officer



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THE CITY UNIVERSITY OF NEW YORK

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INTRODUCTION

STATEMENT OF THE PROBLEM

Sequence stratigraphy and its concepts of parasequences, system tracts, and sequences were originally defined using the model of an "Atlantic type" passive continental margin (Mitchum et al. 1977, Vail et al. 1977a). If the concepts of sequence stratigraphy can be extended to mechanically subsiding foreland basins such as the Paleozoic Appalachian foreland basin of eastern North America then light will be shed on some vexing questions such as the rheology of the lithosphere and the origin of depositional cyclicity.

The coarsening upward, cyclic deposits of the upper Middle Devonian Mahantango Formation of Pennsylvania and New York undergo rapid lateral and vertical facies changes, resulting in a confusing assemblage of lithostratigraphic members. If sequence stratigraphy proved to be a useful tool in Appalachian foreland basin analysis then perhaps a simpler, genetic allostratigraphy would emerge, linking various lithostratigraphic formations and members

into parasequences, system tracts, and sequences. As a byproduct of this approach, constraints may be placed on the formation of foreland basins by loading from thrust sheets (i.e. visco-elastic versus elastic rheologies). The goal of this study is to place the Mahantango Formation of Pennsylvania and New York in a sequence stratigraphic framework and then use that framework to approximately determine the rheology of the Central Appalachian lithosphere during the Acadian Orogeny.

STUDY AREA

The study area is in the Appalachian Valley and Ridge province extending from the vicinity of Harrisburg, in south-central Pennsylvania northeastward through Bowmanstown, Milford, and ultimately, Port Jervis, New York (Figure 1). Figure 1 shows the outcrop belt of the Middle Devonian Mahantango Formation and the locations of the measured sections used in this report (Appendix 1)

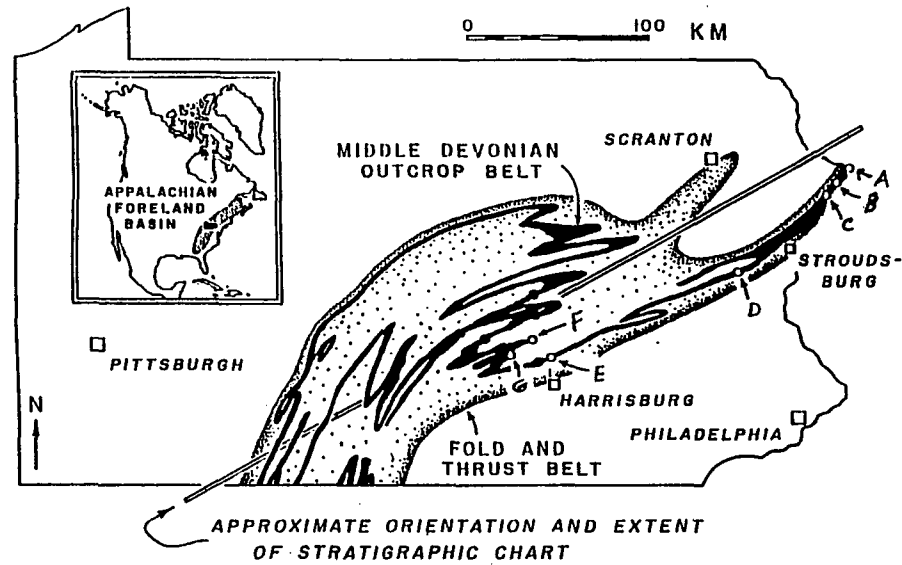


Fig.1. Outcrop belt of the Mahantango Formation in Pennsylvania and immediately adjacent New York. Letters A-G are measured sections. See Appendix 1 for exact locations of measured sections

* APPROACH and METHODS

This study is a field based, outcrop scale study. The sedimentology and facies architecture of seven carefully chosen measured sections and numerous incomplete exposures were used to place the Mahantango Formation in a sequence stratigraphic framework. Therefore, a brief review of the evolution of the paradigm and its concepts are in order. Sloss (1949) was one of the first workers to recognize that the sedimentary cover of cratonic North America contained interregional unconformities. The rocks lying between these unconformities he called sequences. Sloss (1963) refined his earlier sequence boundaries, dividing the sedimentary cover of the North American craton into six sequences (Fig. 2).

With the advent of sub surface well logging technology in the nineteen-fifties, geologists had at their disposal a powerful new means of basin analysis. It became possible to trace seismic and electric-log discontinuities across facies and lithologies but the

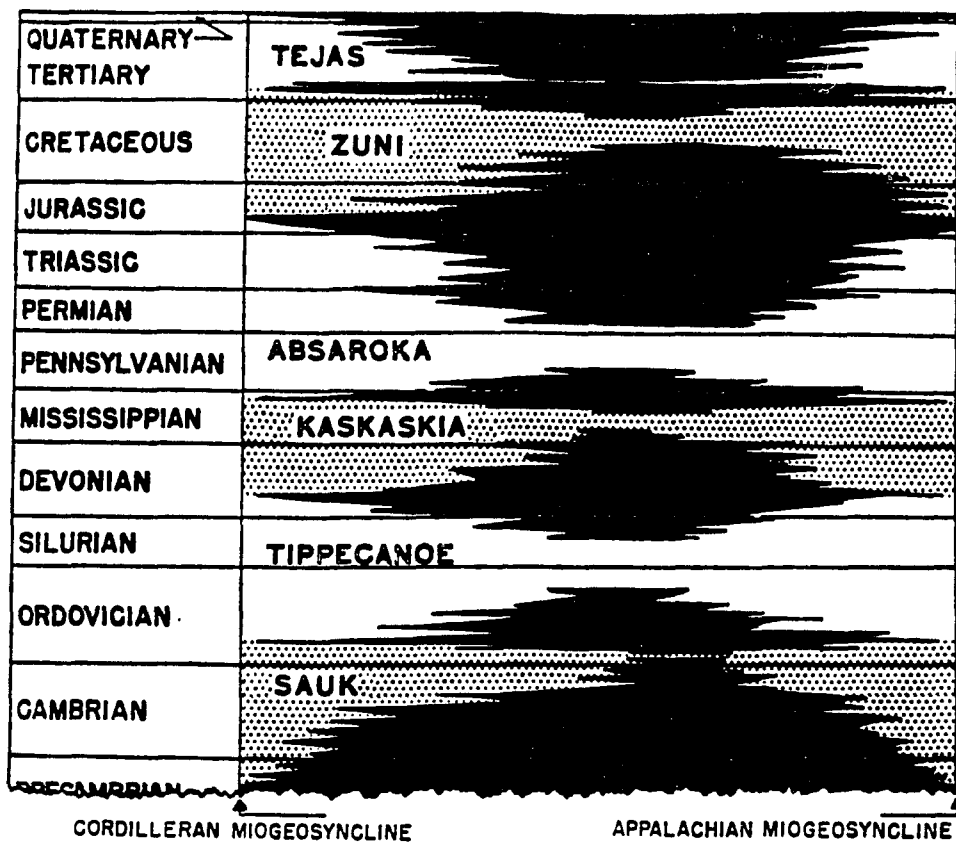


Figure 2. Sloss' sequences on the North American Craton
(From Sloss, L. 1963)

Stratigraphic Code made no allowances for such regional units. Forgotson (1957) coined the term "format" to define a rock unit that crosses facies boundaries and lithostratigraphic units (Figure 3). He suggested that the term apply to operational units defined on the basis of their electric-log signatures. The concept of using marker beds (formats) conveyed an implication that the boundaries of formats were isochronous, or at least penecontemporaneous along their length. Chang (1975) drew on the work of Sloss (1949) and Forgotson (1957) and suggested the term "synthem" for rock units between unconformity bounded surfaces. Gone was the implication of isochronous surfaces but the question of extending synthems to correlative conformities went unaddressed. This was a crucial insight of Sloss' student Peter Vail and others at Exxon Production Research (Mitchum et al. 1977), who helped develop the concepts of sequence stratigraphy. Van Waggoner et al. (1988 p.39) defined sequence stratigraphy as "the study of rock relationships within a chronostratigraphic framework of repetitive genetically related strata bounded by surfaces of erosion or nondeposition, or their correlative conformities."

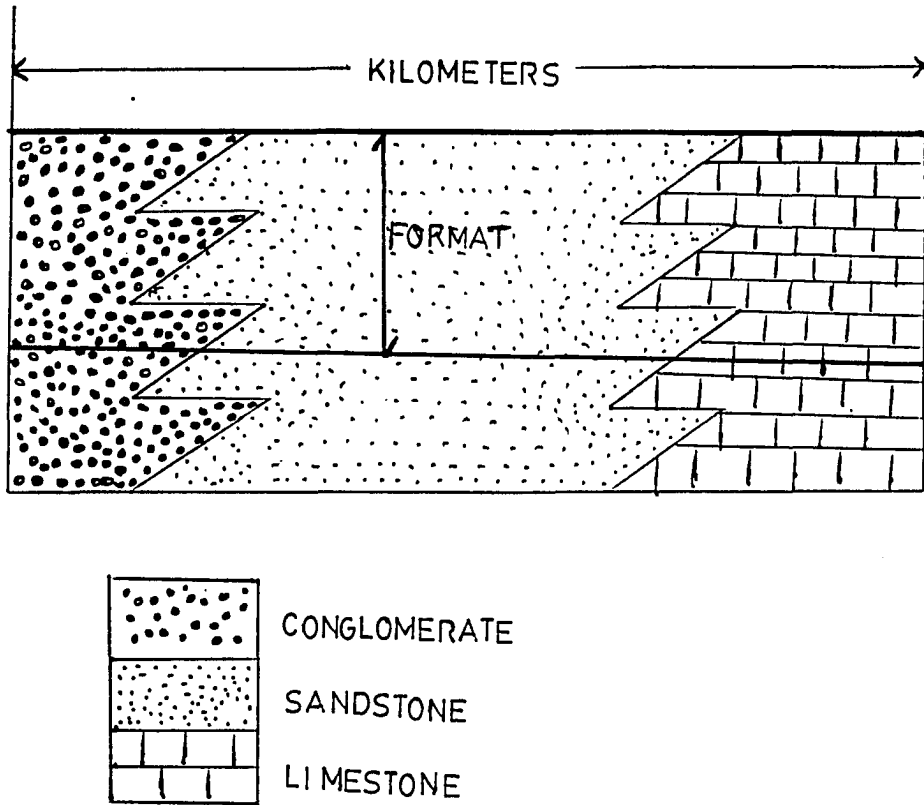


Figure 3. Example of a format crossing facies boundaries.

Parasequences are the fundamental building blocks of sequence stratigraphy. They are defined as relatively conformable successions of genetically related beds bounded by marine flooding surfaces and their correlative conformable surfaces. In clastic basin fills they are progradational and shoal upward. Successive parasequences are stacked in three ways, progradational, aggradational and retrogradational (Fig.4), forming parasequence sets. These parasequence sets form the basis for the next higher ranking of genetically related strata, system tracts. Ideally, there are three system tracts in each of two kinds of sequences. Digressing for the moment, sequences are defined on the basis of their lower bounding surfaces. If the lower bounding surface shows evidence of subaerial erosion and stream rejuvenation it is a Type 1 sequence boundary, if there is evidence of subaerial exposure, but no stream incision, it is a Type 2 boundary (Van Wagoner et al. 1988).

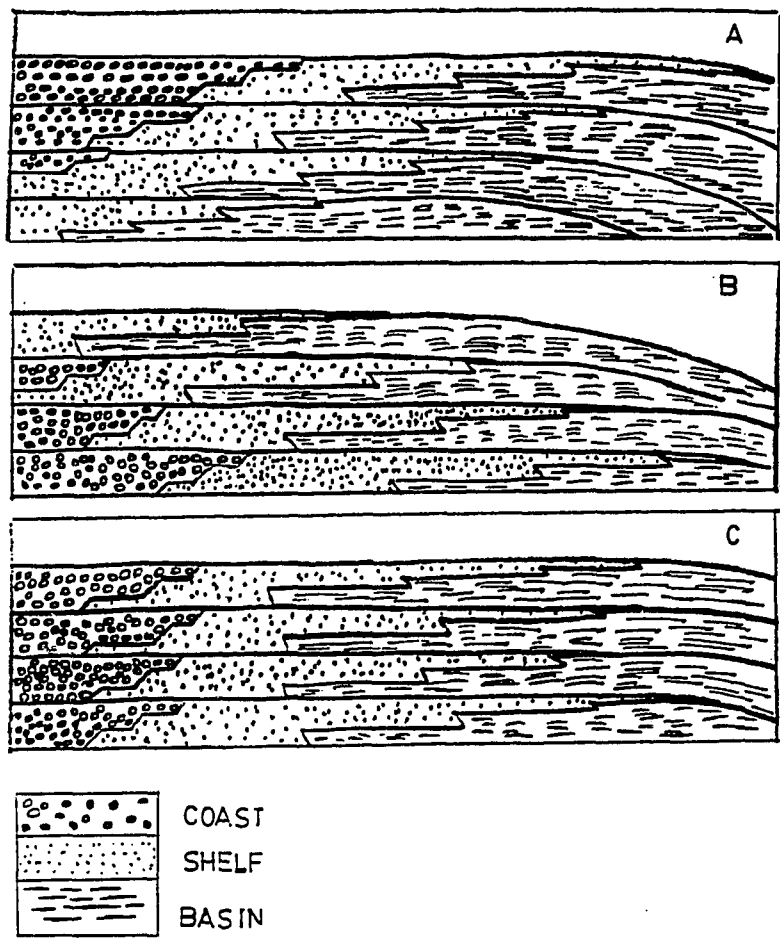


Figure 4. Progradational (a), retrogradational (b), and aggradational (c), parasequence sets. Boundaries between individual parasequence sets are marine flooding surfaces

Type 1 sequences (Posamentier et al. 1988) are composed of Highstand, Transgressive, and Lowstand system tracts. Type 2 sequences share the Highstand and Transgressive system tracts but the "lowest" system tract in a Type 2 sequence is a Shelf Margin system tract. A Highstand system tract is composed of one or more aggradational or progradational parasequence sets. Transgressive system tracts are composed of one or more retrogradational parasequence sets. Within the Transgressive system tract thin, "time-rich" condensed sections occur and consist of abundant and diverse planktonic and benthonic microfossil assemblages, authigenic minerals, and organic matter.

The Lowstand system tract, if deposited in a basin with a shelf break, can be divided into three separate units, a basin floor fan, slope fan, and a lowstand wedge. The fans are deposited at sea level lowstand and the lowstand wedge is deposited during gradual sea level rise. If this system tract is deposited in a basin with a ramp margin it consists of a relatively thin lowstand wedge that contains two parts. The stratigraphically lowest part of the wedge contains evidence of stream incision and sediment bypass of the coastal plain. The sediments of the upper part of the wedge infill

the incised valleys as sea level gradually rises. The Shelf Margin system tract is associated with a Type 2 sequence. It is characterized by one or more weakly progradational to aggradational parasequence sets. The sets onlap onto the sequence boundary in a landward direction and downlap onto the sequence boundary in a basinward direction.

This then is the model of sequence stratigraphy developed on passive continental margins (Vail et al. 1977a, Vail and Hardenbol, 1979). Eustasy is thought to be the driving mechanism in the generation of sequences in these settings (Vail et al. 1977b) however, in tectonic foreland basins, mechanical depression of the lithosphere by thrust sheets is thought to be a first order variable driving the generation of sequences (Jordan et al. 1988).

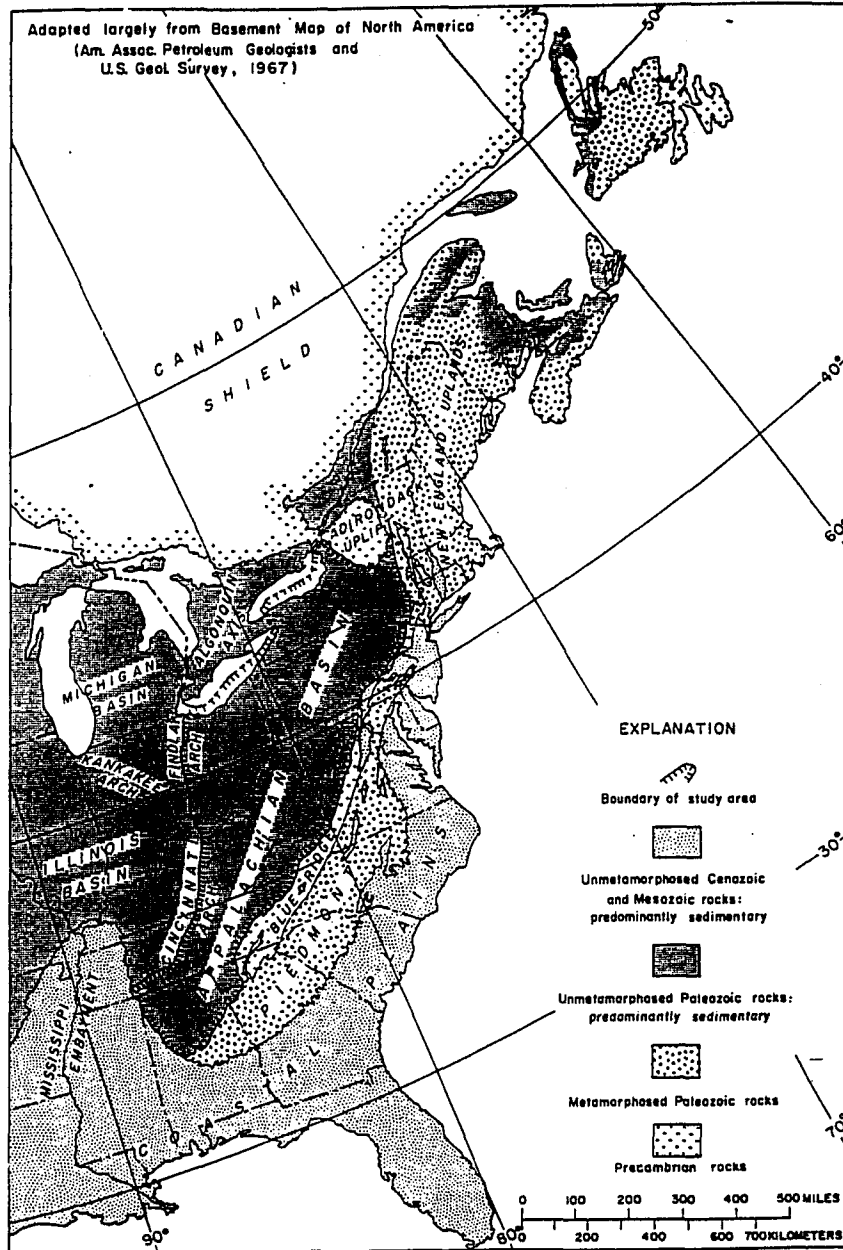


Figure 5. Geological map of the Appalachian foreland basin

THE APPALACHIAN BASIN

INTRODUCTION

The following is a generalized account of the sedimentology, stratigraphy, and tectonics that shaped the Appalachian foreland basin. Every effort will be made to keep the lithostratigraphic nomenclature consistent, but where differences exist the nomenclature used will be that of the Pennsylvania State Geological Survey and / or the New York State Geological Survey. The focus of this study, the Upper Middle Devonian Mahantango Formation, is treated briefly here but will be exhaustively reviewed in a later section.

The Appalachian basin (Fig. 5) is oval in form, and roughly the size of present day Hudson Bay. It extends from the Great Lakes south-westward through southern Tennessee. It is bounded on the west by the Cincinnati Arch and its branches and its eastern edge is buried beneath Piedmont thrust sheets (Cook and Oliver 1981, Milici and deWitt 1988).

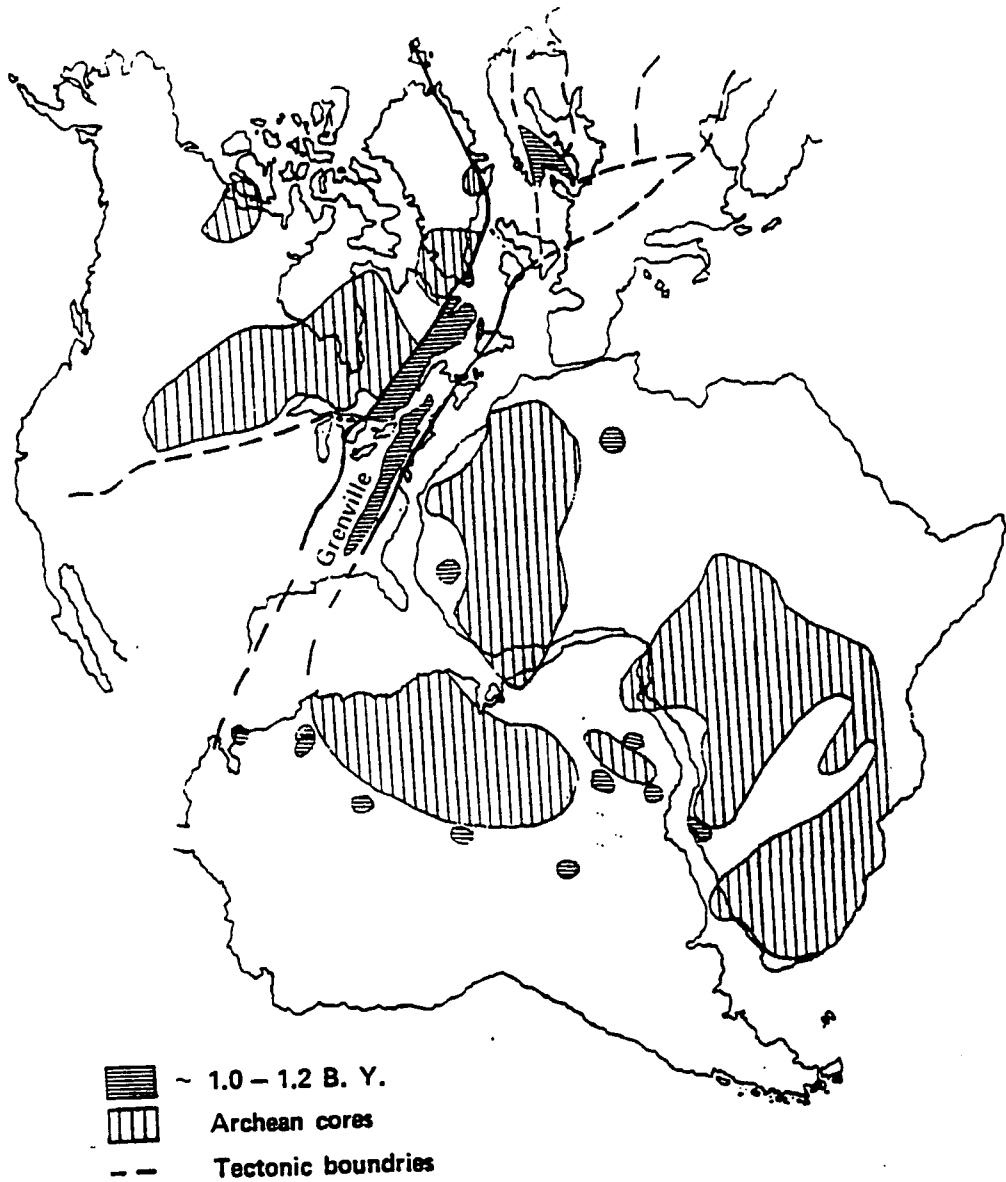


Figure 6. Grenville province in relation to Archean cratons and selected tectonic boundaries. (From Bartholomew, 1984)

The basement rocks upon which the later sediments of the Appalachian Basin were deposited began as passive continental margin sediments that were deformed by the Grenville orogenic event about one billion years ago (Bartholomew 1984, and figure 6, this report). Following this episode of mountain building a long period of erosion ensued with as much as several hundred meters of Grenvillian rock locally removed (Reed 1955; Allen 1967). The latest Precambrian sediments deposited along the eastern border of the Appalachian basin (Fig.7) record the breakup of the North American protocontinent. The Grandfather Mountain Formation and Mechum River Formation were deposited in rift valleys while the Ocoee Supergroup was deposited on an inland sea shelf and rise (Schwab 1986). Interbedded in these earliest Appalachian basin sediments are peralkaline basalts, indicative of tensional rifting.

AGE	GREAT SMOKY MOUNTAINS	NORTHEASTERN TENNESSEE	NORTHERN VIRGINIA, MARYLAND	EASTERN PENNSYLVANIA	SOUTHEASTERN NEW YORK	
EARLY CAMBRIAN	SHADY DOLOMITE	SHADY DOLOMITE	TOMSTOWN DOL	LEITHSVILLE	VINTAGE	STISSING
	HELENMODE FM	HELENMODE MBR	ANTIETAM FM	HARDYSTON FM	ANTIETAM FM	POUGHQUAG QUARTZITE
	HESSE SANDSTONE	HESSE QUARTZITE MBR				
	MURRAY SHALE	MURRAY SHALE MBR				
	NEBO SANDSTONE	NEBO QUARTZITE MBR	HARPERS FM	ARABY FM (LOWER PART)	HARPERS FM	CHICKIES FM
NICHOLS SHALE	HAMPTON FM					
COCHRAN FM	UNICOI FM	WEVERTON FM (*LOUDOUN* FM)				
PRECAMBRIAN	OCOEE SERIES	MT ROGERS FM GRENVILLE BASEMENT	CATOCTIN FM SWIFT RUN FM GRENVILLE BASEMENT	GRENVILLE BASEMENT	GRENVILLE BASEMENT	

Figure 7. Precambrian and early Cambrian stratigraphy of the Appalachian basin, (From Milici, R.C. and deWitt W. Jr. 1988)

CAMBRIAN

As the Iapetus ocean opened a passive continental margin developed and a thick wedge of early Paleozoic strata record its subsidence. The earliest Cambrian sediments are siliciclastics derived from Precambrian crystalline rocks (Milici and deWitt 1988) and the overlying deposits are carbonates. In eastern Maryland and Pennsylvania an embayment (The Glenarm Supergroup) almost isolated a tongue-shaped sliver of continental crust. A modern day analogue to this situation is the peninsula of Korea and the Yellow Sea and the Sea of Okhotsk (Schwab 1986).

The earliest siliciclastics of Cambrian? age are those of the Chilhowee Group. The base of the Chilhowee forms the base of the Sauk sequence (see Figure 2). It unconformably overlies sediments of the Precambrian Ocoee Series in the Great Smoky Mountains of north Georgia and Tennessee, and steps stratigraphically down to rest on the Proterozoic Grenville basement in eastern Pennsylvania, New Jersey, and New York (Fig. 8).

AGE	EASTERN TENNESSEE AND SOUTHWESTERN VIRGINIA	NORTHERN VIRGINIA AND WESTERN MARYLAND	FREDERICK VALLEY, MARYLAND	VALLEY AND RIDGE, PENNSYLVANIA	EASTERN PENNSYLVANIA	OHIO	NEW YORK			
LATE CAMBRIAN	COPPER RIDGE DOLOMITE	CONOCO-CHEAGUE DOLOMITE	FREDERICK LIMESTONE	SEERHARTOWN (PART)	CONOCO-CHEAGUE GROUP	KNOX DOL (PART)	LITTLE FALLS DOLOMITE			
	WARRIOR FM	WARRIOR FM		GATESBURG FM			CONOCO-CHEAGUE GROUP	KERBEL FM	QALWAY FM	
MIDDLE CAMBRIAN	ROCKWELL DOLOMITE	ELBROOK DOLOMITE	ARABY FORMATION (PART)	PLEASANT HILL FORMATION	ELBROOK FORMATION	MOUNT SIMON SANDSTONE	POTSDAM SANDSTONE			
	MARYVILLE LIMESTONE							WARRIOR FM	CONOCO-CHEAGUE GROUP	"ROME" FORMATION
	ROGERSVILLE SHALE							WARRIOR FM	CONOCO-CHEAGUE GROUP	"ROME" FORMATION
EARLY CAMBRIAN	RUTLEDGE LIMESTONE	WAYNESSBORO FORMATION	WAYNESSBORO FORMATION (PART)	WAYNESSBORO FORMATION	LEITHVILLE FORMATION	MOUNT SIMON SANDSTONE	POTSDAM SANDSTONE			
	ROME ("WAUTAUGA") FORMATION	WAYNESSBORO FORMATION						WAYNESSBORO FORMATION	LEITHVILLE FORMATION	
	SHADY DOLOMITE	TOMSTOWN DOLOMITE		TOMSTOWN FORMATION						

Figure 8. Cambrian stratigraphy of the Appalachian basin (From Milici, R.C. and deWitt W. Jr. 1988)

In eastern Pennsylvania the Chilhowee is a tripartite unit with conglomerates and arkosic sands in the lower part, Skolithos burrowed quartzites in the middle part and phyllite grading into quartzites and dolomites at the top (Milici and deWitt 1988). In east-central Pennsylvania and New Jersey, the Chilhowee equivalent, the Hardyston Quartzite, thins to about 240 meters (Aaron 1969) but retains the same impure dolomitic top. In the Adirondack region of New York the basal siliciclastic unit is the Potsdam Sandstone. The Potsdam is divided into a lower portion of non-fossiliferous quartz arenite of early Late Cambrian age while the upper portion of calcareous sandstone may be latest Late Cambrian or Early Ordovician in age, consistent with a postulated Late Cambrian eustatic lowstand (Fortey 1984, Miller 1984). This unconformity may be related to the development of the Rome trough, the eastern expression of the Eastern Interior aulacogen of Harris (1978, and Fig. 9, this report) The earliest sustained carbonate deposition in the study area is the Leithsville Formation in Pennsylvania and its equivalents.

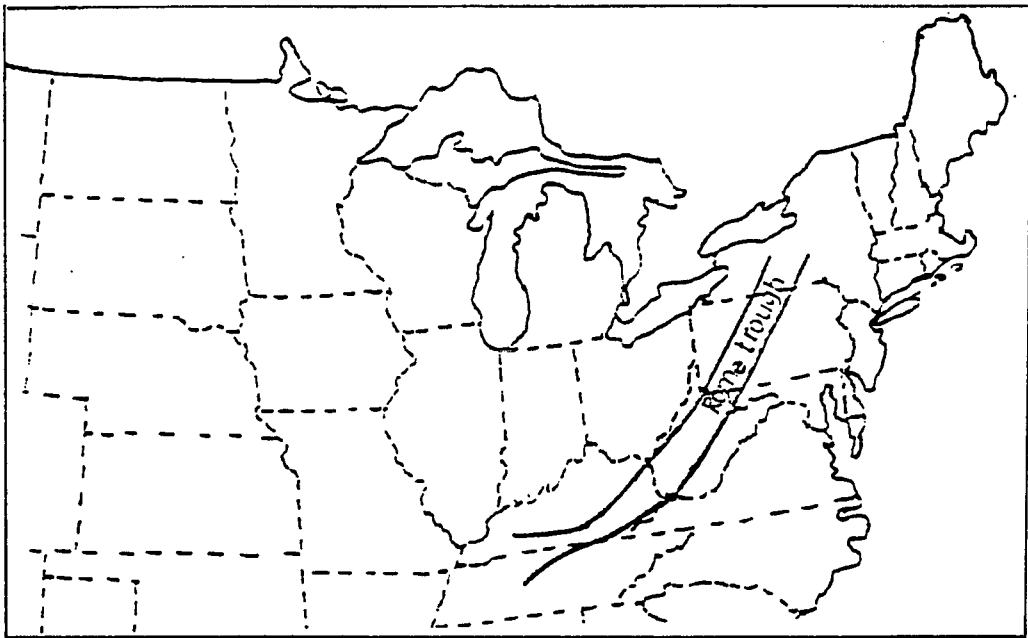


Figure 9. The Rome Trough in Early Paleozoic time.

Elsewhere in Pennsylvania the basin, slope, and shelf deposits of the Vintage, Kinzers, and Ledger Formations of Early and Middle Cambrian age occupy the stratigraphic position of the Leithsville (Gohn 1976, Berg et al. 1983). Transgression in the late Cambrian produced the shallow water carbonates of the Allentown Formation and its equivalents, the Gatesburg Formation of central Pennsylvania and the Galway Formation of the Mohawk Valley, New York. By the end of the Cambrian, the passive continental margin was well established.

ORDOVICIAN

Carbonate deposition continued into the Lower Ordovician with deposition of the Beekmantown Group limestones and dolostones (Fig. 10). In central Pennsylvania the lower beds of the Beekmantown thin to the north and west and are absent in western Pennsylvania, New York and northeast Ohio. Wagner (1966) suggests that this reflects uplift of the Cincinnati arch.

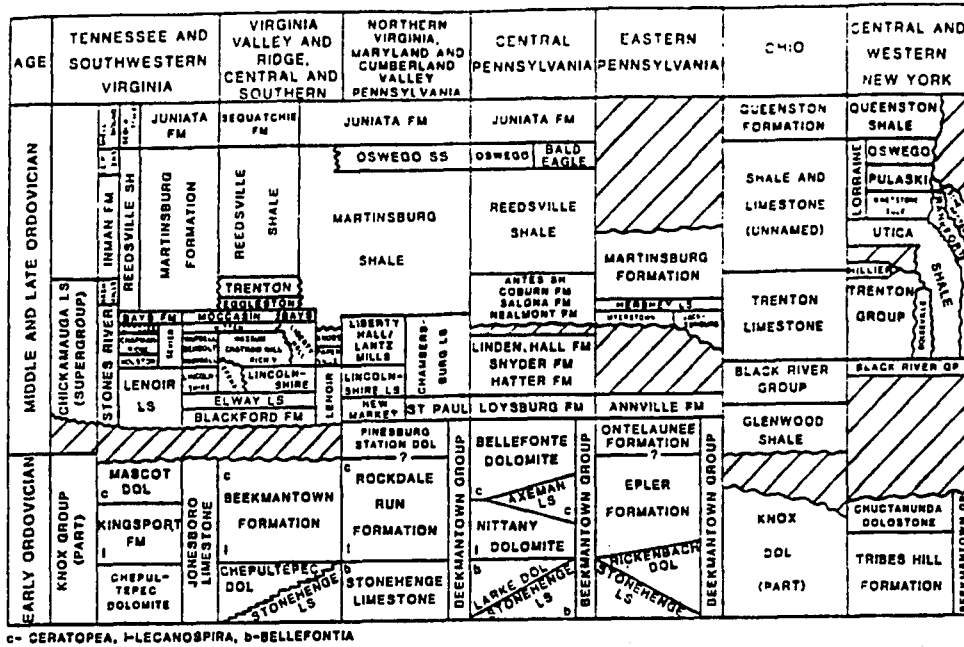


Figure 10. Ordovician stratigraphy of the Appalachian basin (From Milici, R.C. and deWitt W. Jr. 1988)

Except in the deepest parts of the basin, the Lower-Middle Ordovician boundary is marked by an unconformity (Bridge 1955, Milici 1973). This unconformity marks the top of Sloss' Sauk sequence (see Figure 2) and the base of the Tippecanoe sequence. This unconformity has been attributed to the collapse of the carbonate shelf related to the first collisional events of the middle Ordovician Taconic orogeny (Jacobi 1981) that involved a collision between a volcanic arc and the North American craton (Wehr and Glover 1985; Van der Pluijm 1987). This event is recorded in the Appalachian basin by the progressive westward progradation of siliciclastics on the collapsing carbonate platform. Deposition of the Martinsburg Formation and equivalents (Austin Glen, Reedsville etc.) took place in slope and deep basin settings (flysch), followed by the basin filling sediments ("molasse") of the "Queenston Delta". The "Queenston" rocks reach their maximum thickness in two depocenters, one in Pennsylvania and the other in West Virginia. In central Pennsylvania, the sediments of the Bald Eagle, Juniata and Tuscarora Formations are the expressions of the prograding wedge, but along the basin margin areas of eastern Pennsylvania and New Jersey an unconformity exists through the Late Ordovician. This

unconformity, caused by the uplifting of the eastern margin of the basin by Taconic orogenesis changed the shape of the basin from the gently sloping east facing monocline of the early Ordovician to an oval-shaped foreland basin. The basin was to retain this general physical configuration throughout the remainder of its depositional history.

SILURIAN

Molasse deposition in the Appalachian Basin continued with the emplacement of the Shawangunk conglomerates and sandstones that rest in angular unconformity on the Martinsburg Formation (Epstein and Epstein 1972). To the west, in central Pennsylvania the Shawangunk equivalents are the Keefer Formation and the slightly younger Tuscarora and Rose Hill Formations (Fig. 11). Further to the north and west, the Medina and Clinton Groups are stratigraphic equivalents.

Overlying the Shawangunk and equivalent rocks is the Bloomsburg Formation, a red bed deposit that extends across much of the central Appalachians, from eastern Pennsylvania and New Jersey south-eastward through Maryland into parts of West Virginia and Virginia.

The Bloomsburg consists of a variety of lithologic units ranging from conglomerate to claystone. In western New York the Bloomsburg stratigraphic equivalent is the Lockport Group consisting of biohermal limestones, dolostones and shale. Between the prograding Bloomsburg in the east and the reefs and banks to the west a restricted sea developed in which the thick evaporites of the Salina Group were deposited, although these are for the most part younger than the Bloomsburg Formation. Within the restricted Salina basin, limestone, dolostone, anhydrite, and halite formed in concentric patterns (Rickard 1969).

The continued erosional beveling of the Taconic thrust sheets resulted in diminished clastic influx thereby allowing carbonate production to resume in central and eastern Pennsylvania with deposition of the Poxono Island, Bossardville, and Decker Formations. Epstein et al. (1967) concluded that these impure dolomites and limestones were deposited in shallow water shelf to lagoonal environments. During the Silurian, the Appalachian Basin lay in a subtropical to tropical setting in low southern latitudes (Van Der Voo 1988)

DEVONIAN

Shallow marine carbonate production continued into the early Devonian with deposition of the Coeymans and New Scotland Formations in the study area (Fig. 12). Above these carbonates is the quartzitic, marine Oriskany Sandstone that records a relative drop in sea level and marks the top of Sloss' Tippecanoe sequence (Fig. 2). The overlying strata of the Needmore Shale in central Pennsylvania and Onondaga Limestone in central and eastern New York signal the beginning of the Kaskaskia sequence (Fig.2). In early Middle Devonian time, the effects of the Acadian orogeny were beginning to be felt within the Appalachian foreland basin. This tectonic event has been alternately attributed to the collision between Armorica, Laurentia (ancestral North America) and Baltica (Perroud et al. 1984) or the oblique convergence of Laurentia and Avalonia along a major sinistral strike slip fault (Ettensohn 1985; Van Der Pluijm 1987). During the Middle Devonian, the Appalachian Basin rapidly subsided and was infilled with the siliciclastics of the Hamilton Group that were shed from the advancing Acadian thrust sheets.

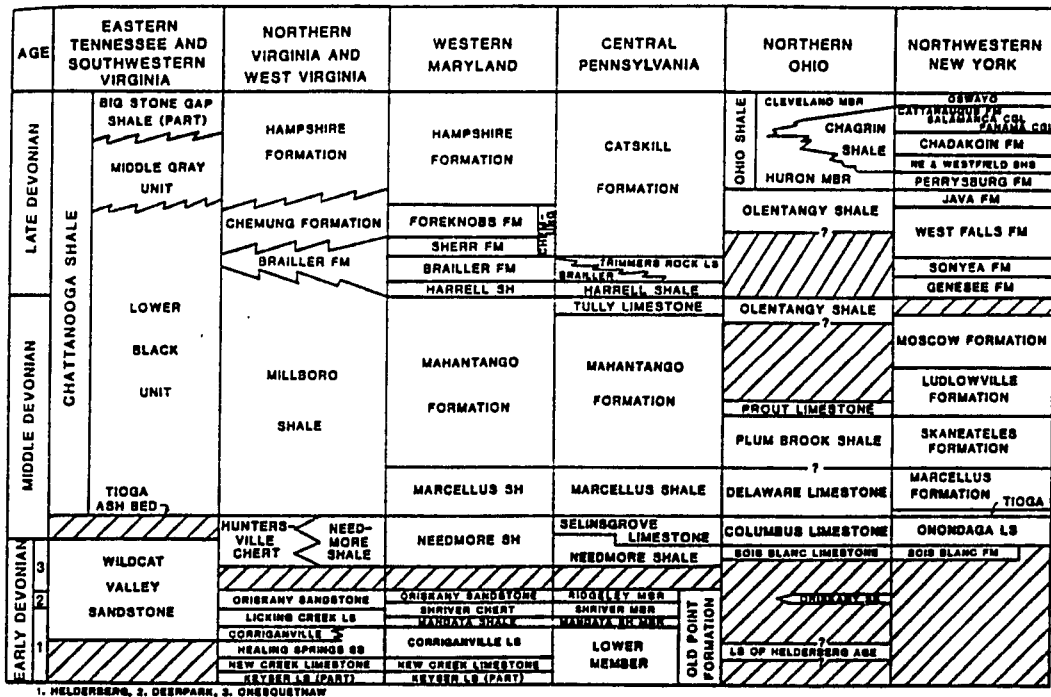


Figure 12. Devonian stratigraphy of the Appalachian basin (From Milici, R.C. and deWitt W. Jr. 1988)

The Marcellus Shale and its equivalents were followed by the cyclic deposits of the Mahantango Formation. The lithology of the Mahantango Formation in the study area ranges from conglomerate and coarse sand in south-central Pennsylvania to finer siltstone and shale elsewhere. In central and western New York, the Hamilton Group consists of dark gray shale and siltstone with interbedded carbonate and comprises one of the most complete and fossiliferous sections of middle Paleozoic strata in the world (Brett and Baird 1989). Overlying the Mahantango Formation in central Pennsylvania are the Harrell and Burket Shales that record a deepening event. Within the study area of eastern Pennsylvania and extreme southeastern New York this deepening is evident in the upper parts of the Mahantango Formation. The Tully Formation and its equivalents, the Sparrow Bush Formation and the Laurens Sandstone stratigraphically overlie the Mahantango. The Upper Devonian section reflects siliciclastic infilling of the basin from the east-northeast with terrestrial clastics prograding westward and southward across the underlying shallow water and basinal deposits. Concomitant westward migration of the basin axis is recorded by black shales

that step progressively south-westward across the Appalachian Basin (Ettensohn 1985, his figure 5, p.46).

CARBONIFEROUS AND PERMIAN

The general paleogeography and depositional framework established during Late Devonian time characterized the Appalachian basin throughout most of the early Mississippian (figure 13, this report, Milici and deWitt 1988). In eastern Pennsylvania, the Specty Kopf Formation is succeeded stratigraphically by the fluviatile Pocono Formation of Early Mississippian age. A transgression during Middle Mississippian (Meramecian) time resulted in deposition of carbonate sediments over most of the basin, except for its easternmost margin. There the Mauch Chunk rocks record subaerial fluvial deposition. In south-eastern Pennsylvania this transgression is marked by several tongues of fossiliferous calcarenite in the lower part of the Mauch Chunk Formation. Thus, in many places along the eastern margin of the basin, deposition was continuous, and therefore the boundary between the Mississippian and Pennsylvanian is difficult to place.

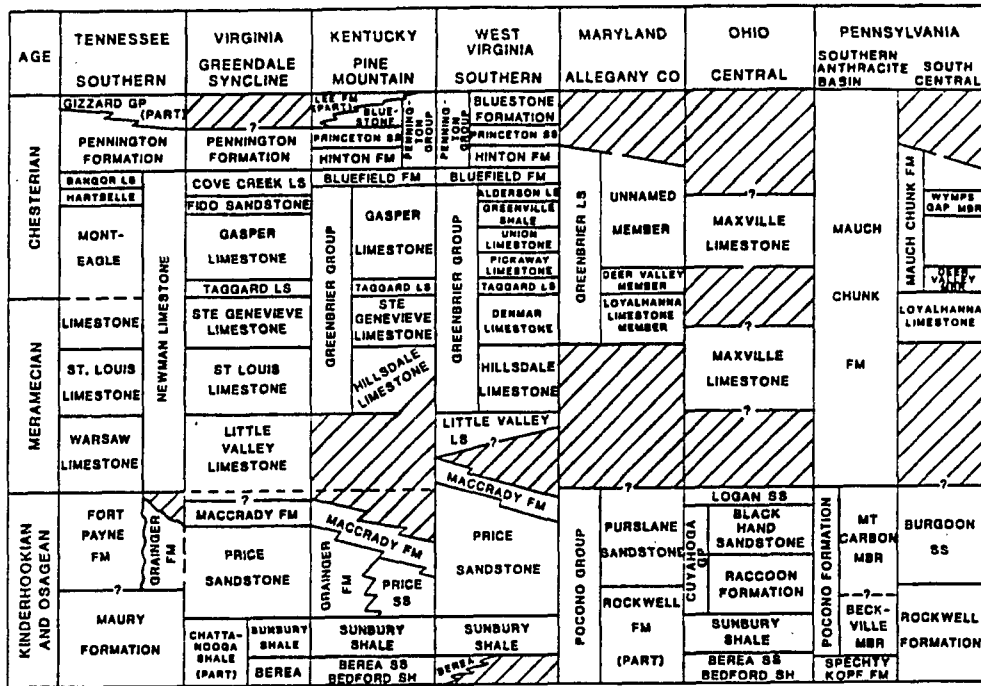


Figure 13. Mississippian stratigraphy of the Appalachian basin (From Milici, R.C. and deWitt W. Jr. 1988)

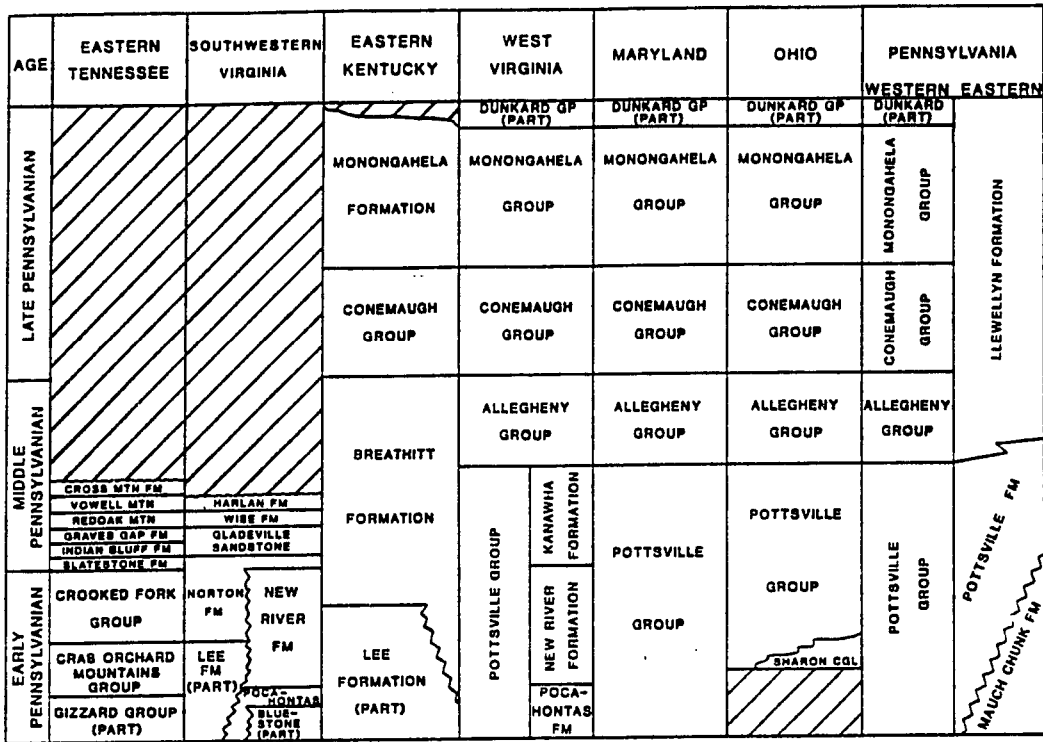


Figure 14. Pennsylvanian stratigraphy of the Appalachian basin. (From Milici, R.C. and deWitt W. Jr. 1988)

However, erosion along the northern and western margins produced topography that was filled by Pennsylvanian rocks. The relationship between the Mauch Chunk and the overlying Pottsville Formation in Pennsylvania (Fig. 14) has been variously interpreted as gradational or erosional. The unconformity between the Mauch Chunk and Pottsville Formations is the boundary between the Kaskaskia and Absaroka sequences of Sloss (Fig. 2). This unconformity is related to the Alleghanian orogeny. This pulse of orogenic activity has been linked by Van Der Pluijm (1987) to transpressional emplacement of the Meguma terrane in Late Devonian/ Carboniferous time. The Llewellen Formation, a non-marine coal bearing unit that overlies the Pottsville Formation is the youngest rock in the study area. In the western part of the basin deposition continued with the Permian Dunkard Group until continental collision and deformation of the Alleghanian orogeny ended the sedimentary history of the Appalachian foreland basin.

THE HAMILTON GROUP

The Middle Devonian Hamilton Group of Pennsylvania and New York was deposited in a shallow (Vogel et al. 1987), tropical (Van der Voo 1988), epeiric sea (Fig. 15) at or near the apex of the Acadian orogeny. Figure 16 shows the general paleogeography and facies belts of New York and Pennsylvania during Middle Devonian time.

The Hamilton Group is a detrital sedimentary package representing the first coarse clastic rocks shed from Acadian uplifts in New England and the Middle Atlantic area into the subsiding Appalachian foreland basin (Brett 1986). Hamilton Group rocks have been studied for over one hundred and fifty years. The term Hamilton was introduced into the literature by Vanuxem (1840,1842), based on exposures in the Chenango Valley near Hamilton, New York, and included all the rocks between the Marcellus black shale and the Tully Formation. Subsequent workers include Smith (1916), Chadwick (1933), and Cooper (1930, 1933, 1934).

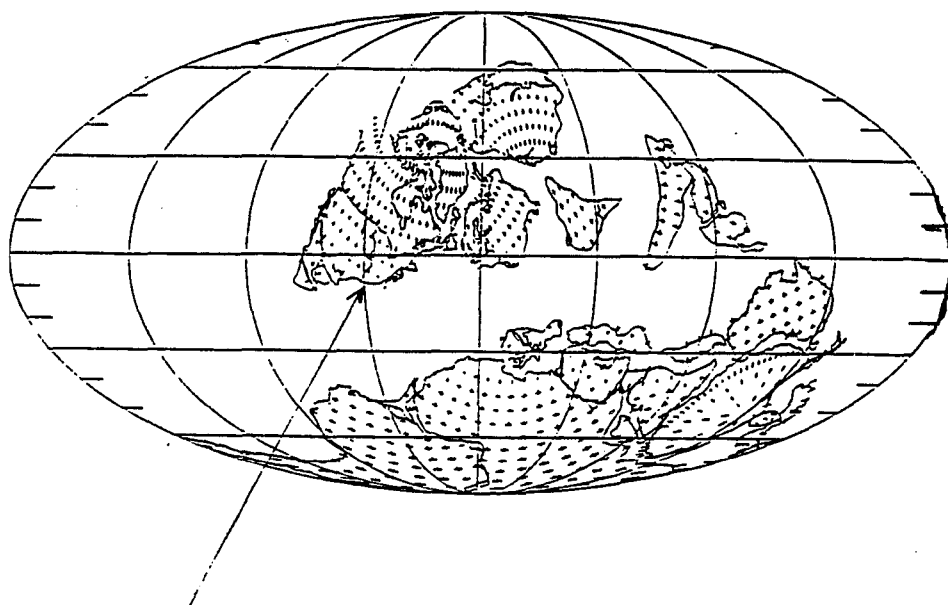


Figure 15. Paleogeographic reconstruction of North America during the Devonian. Arrow points to eastern margin of Laurentia (From Scotese et al., 1979). Black indicates position of the study area

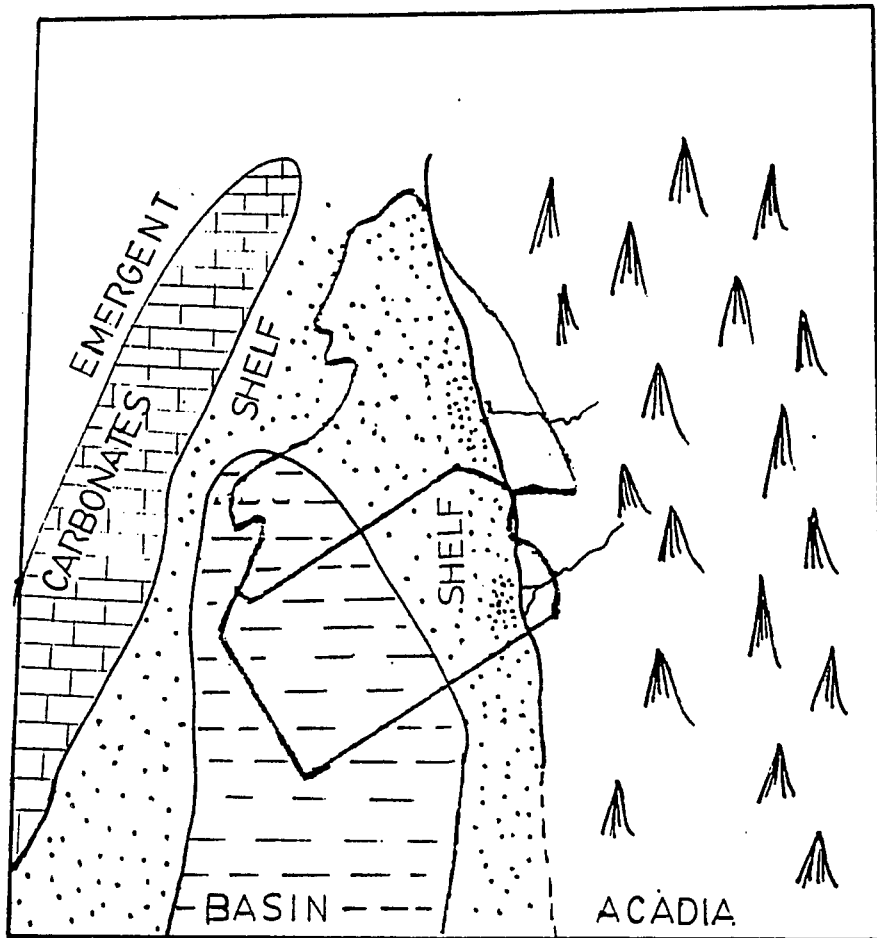


Figure 16. Paleoreconstruction of facies belts in the Devonian Catskill Sea (From Brett C.E., 1986)

The fossiliferous, medium to dark gray mudrocks that comprise the Hamilton Group of western and central New York were subdivided by Cooper into four formations largely on the basis of thin limestone marker beds. Cooper's formations in ascending order are the Marcellus, Skaneateles, Ludlowville and Moscow Formations (Fig.17). The limestone marker beds are the Stafford limestone at the base of the Skaneateles, the Centerfield limestone at the base of the Ludlowville, and the Tichenor-Menteth limestone at the base of the Moscow Formation. Limestones such as these have been interpreted as being deposited at times of transgression and sediment starvation (Johnson and Friedman 1969, McCave 1969, 1973), but recent work by Brett and Baird (1985) has interpreted these thin, tabular limestones as biogenic accumulations deposited during maximum regression of sea level. Although the Middle Devonian Givetian stage is not readily zoned biostratigraphically (Brett 1986), Brett and Baird have used conodont biostratigraphy (Klapper 1981, p.61) to demonstrate that the Tichenor limestone is slightly older in the Finger Lakes area relative to exposures further west near Buffalo, New York.

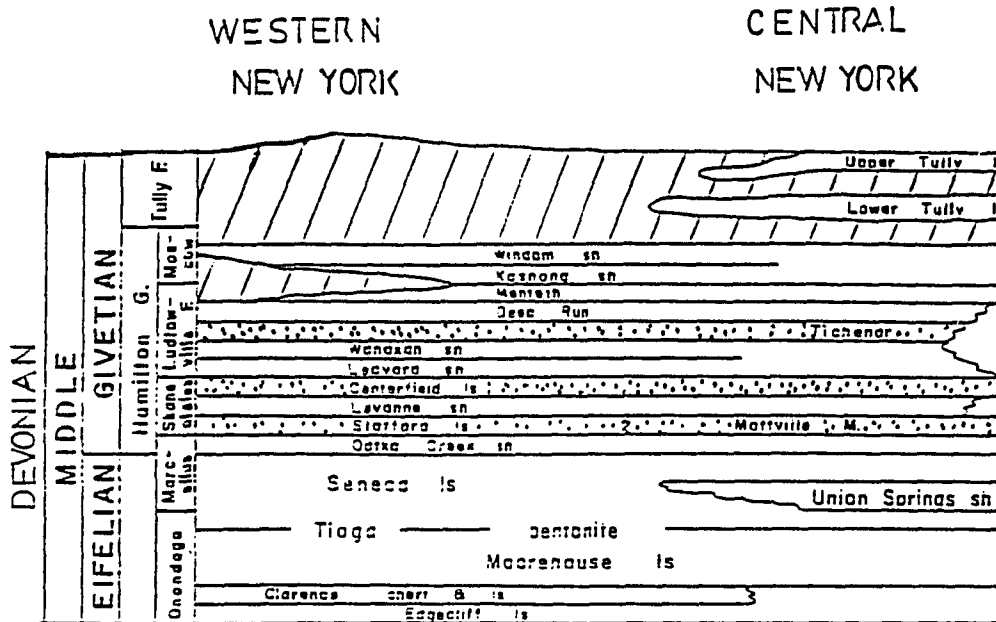


Figure 17. Hamilton Group stratigraphy of central and western New York. Limestone marker beds stippled (Modified from Sevon and Woodrow, 1985)

They view the thin limestones as regressive maxima of cycles (shale-limestone-shale). The cyclicity of the Hamilton Group in New York is also evident in Pennsylvania and was noted by the earliest workers there. White (1883, 1885) divided the Hamilton Group along the Susquehanna River into five parts (three shales separated by the Selinsgrove upper and lower sandstones), and Claypole (1885) defined a tripartite division (shale-Montebello Sandstone-shale) in central Pennsylvania. Because the Hamilton Group rocks in Pennsylvania were clearly different from those in New York, local names were developed for the Pennsylvania exposures. Work by Butts (1918), Willard, (1935) and Ellison, (1965) extended the stratigraphic nomenclature and Fail et al. (1978) made the most recent revision of the Hamilton Group in central Pennsylvania. The stratigraphically lowest formation in their revision is the Onondaga Formation, that contains two members, the Needmore Shale at the base and the Selinsgrove Limestone at the top. The base of the Middle Devonian is placed at the contact between the two members. The Tioga bentonite lies between the Onondaga Formation and the overlying unit, the Marcellus Formation. The Marcellus is a black to

dark gray fissile shale although minor sandstones and limestones are present within it. The Marcellus Formation is overlain by the Mahantango Formation everywhere in the study area.

MAHANTANGO FORMATION

The Mahantango Formation was introduced into the literature by Willard (1935) as the equivalent of the Skaneateles, Ludlowville and Moscow Formations in New York. The name is taken from the north branch of Mahantango Creek in Snyder and Juniata Counties, Pennsylvania. The Mahantango Formation consists of shale, siltstone, and sandstone arranged in asymmetric, coarsening upward cycles (Kaiser 1972, Faill et al. 1978, Sarwar 1984). The contacts between successive cycles are sharp, coinciding with an abrupt decrease in grain size. Faill et al. (1978) note that in addition to the vertical variation in grain size, the maximum grain size within individual cycles decreases to the west and north away from central Pennsylvania. These rapid lateral and vertical facies changes force Mahantango Formation member definition and terminology to change from place to place, with correlations based on similarity of lithology. Below is a brief description of the Mahantango Formation

members in south-central Pennsylvania (Fig.18).

The Turkey Ridge Member is the stratigraphically lowest member of the Mahantango Formation. It consists of non-fossiliferous light to medium olive gray fine to medium sandstone. The next highest stratigraphic unit is the Dalmatia Member which contains the lowest coarsening upward cycle in the formation. Dark olive gray shales grade upward to a medium sandstone at the top. In the coarsest beds of this member, quartz and shale pebbles are present. The Fisher Ridge Member is a shale and siltstone unit that underlies the Montebello Member, the coarsest unit of the Mahantango Formation in south-central Pennsylvania. The Montebello contains mostly quartzitic sandstone and conglomerate with minor amounts of siltstone. The contact between the Montebello and the overlying Sherman Ridge Member is very sharp, with an abrupt decrease in grain size from coarse sand to shale. The overlying Sherman Ridge Member consists of two major cycles of fossiliferous silty claystone grading upward to a silty sandstone at the top of each cycle.

The contact between the Sherman Ridge and the overlying Tully Member in south-central Pennsylvania is gradational according to Faill et al. (1978), but a profound lithic break exists between the Tully carbonates and the overlying black shale of the Harrell Formation.

Northeast of this central area, the Mahantango Formation becomes finer grained and the lithological member framework of south-central Pennsylvania breaks down. At Milford, in north-eastern Pennsylvania the Mahantango has been informally subdivided by Fletcher and Woodrow (1970) into a lower shale unit, a middle sand rich unit with well developed cyclicity and an upper shale unit. In the Bowmanstown, Pennsylvania, area, the coarsest material in the entire Mahantango Formation is a thin (less than 20 meters thick) siltstone body named the Nis Hollow Siltstone. Epstein et al. (1974) noted that the Nis Hollow Siltstone occurs in the same stratigraphic position as the much thicker Montebello Sandstone in central Pennsylvania (Epstein et al. 1974, p.141) but physical correlation has not been attempted. To the west and south-west towards the Maryland border the Mahantango Formation is also fine-grained, with the thin, but persistent Chaneyville and Clearville

siltstone bodies present at the stratigraphic position of the coarser Montebello and Sherman Ridge Members. The westernmost extent of the Clearville Siltstone defines the lithostratigraphic limit of the Mahantango Formation where rocks above the Needmore Shale and below the Brailer Formation are called the Millboro Shale.

The rapid lateral and vertical facies changes and associated grain size variations both within the central Pennsylvania area and in all directions away from there make the lithostratigraphic correlation of the Mahantango Formation artificial and strained. Members defined on the basis of lithology can say little about time relationships or facies architecture. If stratigraphy is to be used as a tool for basin analysis, it must move beyond the descriptive cataloguing of lithologies, to a stratigraphy that reflects the dynamic processes at work in foreland basin evolution.

DESCRIPTION OF FACIES ASSOCIATIONS

Figure 1 shows the study area and the location of the measured sections incorporated into this study. A total of eighteen different facies defined on the basis of their sedimentological and paleontological characteristics were defined from these measured sections and many incomplete exposures. These facies have been grouped into seven Facies Associations. A complete description and interpretation of each follows.

Facies Association 1

Facies Association 1 (Figs. 19, 20) is mudstone dominated, locally fossiliferous or with thin siltstone or fine sandstone interbeds. All the facies in this Association exhibit some degree of bioturbation, mainly of the Cruziana ichnofacies of Frey and Pemberton (1984). Three facies have been recognized in this Facies Association.

Unfossiliferous Mudstone

This facies is commonly several meters to many tens of

meters thick, and consists of medium to dark gray mudstone with no evident sedimentary structures. This facies shows little lateral variation, but locally exhibits some vertical variation, with evidence of biogenic activity. Siltstone grains float in the mud matrix in some locations and lends an overall ropy texture to these intervals. The ropy intervals are slightly more resistant to weathering and therefore do not weather as recessively as intervals without siltstone.

Fossiliferous Mudstone

This facies also consists of medium to dark gray mudstones, commonly many meters to several tens of meters thick with no obvious sedimentary structures. Dispersed silt grains occur within some intervals as in the facies above but in addition this facies is sparsely to moderately fossiliferous, with a fauna of thin shelled brachiopods and pelecypods. The fossils may occur as isolated individuals or as scattered individuals along horizontal planes. Some planes may be traced for a few meters along outcrops. Shells in this facies are commonly unabraded but disarticulated.



Fig. 19. Photo of Facies Association 1, Mudstone with some fossiliferous horizons. Route 209, Milford Pa.

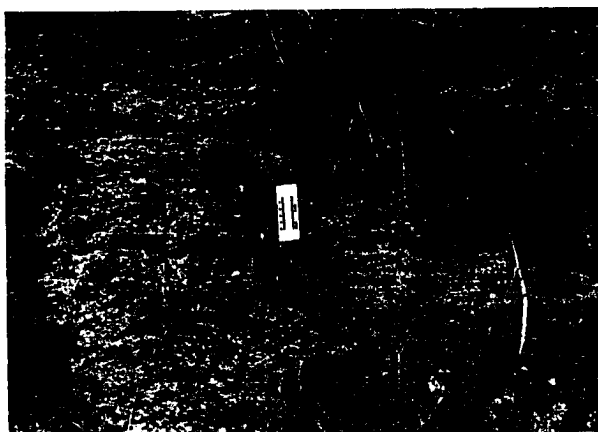


Fig. 20. Photo of Facies Association 1, Mudstone with thin interbeds. Route 209, Milford Pa.

Mudstone with Thin Siltstone Interbeds

This facies is mudstone dominated with thin (less than 15 centimeters thick) interbeds of siltstone or very fine sandstone. The mudstone lacks sedimentary structures, and silt grains float in some intervals that appear ropy and bioturbated. The siltstone interbeds have a sharp basal contact with the underlying mudstones and exhibit a slight fining upward in some beds. The upper contact is gradational into overlying mudstone and on rare occasions display symmetric ripples. The interbeds are usually laterally continuous across outcrops, but some beds are locally almost obliterated by bioturbation. More commonly, beds display a smaller measure of bioturbation, with burrowed tops and escape structures. Some slight pinching and swelling of beds laterally occurs in some locations.

Interpretation Facies Association 1

Based on the preponderance of mudstone, and the presence of trace fossils and scattered body fossils Facies Association 1 rocks are interpreted as hemipelagic sediments of the outer to mid shelf, generally emplaced below fair weather wave base. The thin siltstone

and sandstone interbeds are inferred to be storm emplaced deposits based on their sharp basal contacts and slight fining upward trends, and the local presence of oscillation ripples on the bed tops.

Facies Association 2

This Facies Association is composed of mudstone interbedded with sandstone (Figs. 21, 22). Mudstone dominates this Association, but sandstone locally comprises up to 50% of the rock. Occurrences vary in thickness from a few to several meters. The sandstones in this Facies Association are generally very fine to fine grained and from many centimeters up to 1.5 meters thick. The intervening mudstones are lithologically and paleontologically similar to those in Facies Association 1. Three facies have been defined in this Association.

Mudstone with Thin Hummocky Cross Stratified Sandstone Interbeds

This facies consists of mudstone interbedded with very fine to fine sandstone from 15 to 75 centimeters thick. The sandstone beds are sharp based and locally display slight fining upward trends. Disarticulated pelecypod and brachiopod fossils occur scattered on

the bottoms of some sandstone beds. Upper surfaces are locally bioturbated with forms from the Skolithos and Cruziana ichnofacies. The upper surfaces of the sandstone are capped with oscillation, asymmetric, or interference ripples in places. Beds are laterally continuous along outcrops and commonly pinch and swell with thickness variations of several centimeters, and spacings of one to a few meters. Internally the laminae display parallel to low angle truncations.

Mudstone with Thick Hummocky Cross Stratified Sandstone Interbeds

The distinction between this facies and the preceding is a matter of degree. This facies contains 75 to 150 centimeter thick sandstone interbeds. The overall sandstone component increases in abundance to approximately 50%. Beds are sharp based and locally display slight fining upward trends. Scattered body fossils are found at the bases of some beds. Internally, the sandstone contains sets of parallel to low angle laminae that truncate each other. Beds are continuous along outcrops and display centimeter to decimeter scale thickness variations due to pinching and swelling.



Fig.21. Facies Association 2 showing thin hummocky cross-stratified sandstone interbeds grading upward into thick hummocky cross-stratified sandstone interbeds. Watts section.



Fig.22. Mudstone with intensely bioturbated interbeds. Girtys Notch section.

The spacing from swell to swell is one to a few meters. Beds are capped in places with symmetrical, asymmetrical, or interference ripples. Bioturbation of the tops of beds by forms of the Cruziana and Skolithos ichnofacies is common locally. The thin and thick hummocky cross stratified sandstones of this Association commonly form a continuum, with less intervening shales and thicker sandstones higher in the Facies Association.

Mudstone with Bioturbated Sandstone Interbeds

This facies is characterized by intensely bioturbated intervals of mudstone and interbedded sandstone. Sandstone interbeds may be totally obliterated, giving the mudstone a silty or sandy appearance, or some part of the sandstone interbeds may be preserved. When preserved, sandstone interbeds are generally a few to several decimeters thick and riddled with "Teichichnus-like" burrows. The remnant sandstone may extend several meters along the outcrop. What internal structure remains in the remnant sandstone shows evidence of parallel to low angle laminae.

Interpretation Facies Association 2

The presence of scattered fossils at the bottom of some sandstone beds, internal parallel to low angle laminae, meter scale pinching and swelling and rippled tops are characteristic of storm emplaced hummocky cross stratified sandstone (Harms et al. 1975). The frequent occurrence of oscillation and interference ripples at the tops of beds, especially in the thick bedded facies implies that they were subject to waning storm waves or perhaps, a degree of fair weather wave agitation. However, the presence of intervening hemipelagic mudstones and colonization by opportunistic bioturbating organisms suggests that Facies Association 2 was deposited predominantly below mean fair weather wave base. This evidence therefore suggests that the rocks of Facies Association 2 were deposited on the storm dominated region of the mid to inner shelf.

Facies Association 3

Sandstone forms the bulk of rocks in this Facies Association, but interbeds of silty mudstone are present in the north-eastern part of the study area where mudstone forms a predominant part of

the section. In this area, the Facies Association is characterized by many centimeter to several decimeter thick amalgamated units whereas in the coarser clastic areas such as south-central Pennsylvania the amalgamated sandstone units are commonly several to many meters in thickness. Bioturbation is common and ranges from slight to intense in the sandstone beds and the centimeter thick intervening mudstone. Skolithos and Planolites forms dominate. Two facies have been recognized in Facies Association 3.

Thin Amalgamated Sandstones

This facies (Fig. 23) occurs primarily in the areas of north-eastern Pennsylvania and adjacent New York, where mudstone forms the predominant part of the section (Fig.1, locations A, B, C.), and ranges from a few to several meters in thickness. The individual sandstone bodies are commonly a few centimeters to a couple of decimeters thick and are composed of very fine to fine sandstone. Evidence of amalgamation exists in the form of scattered fossils along the base of individual beds and the presence of symmetric or interference ripples that locally cap individual beds within the

amalgamated sandstone body. Beds internally display flat to low angle laminations and little or no thickness variation laterally. Their lower contacts are sharp on the underlying sandstone or mudstone. The millimeter to centimeter thick silty mudstone interbeds display varying amounts of bioturbation, with Planolites the most commonly recognized traces. The thickness of the mudstone interbeds decreases upwards within this facies, and in places they are entirely absent.

Amalgamated Hummocky Cross Stratified Sandstone

This facies (Fig. 24) consists of amalgamated very fine to fine sandstone bodies a few to several meters thick. Individual beds are generally several decimeters thick and display flat to hummocky cross stratification. These beds are continuous along outcrops with variations of several centimeters in thickness, and may be capped locally with thin mudstone breaks. The sandstone bottom contacts are erosional and sometimes undulatory, trapping small amounts of shale within erosional scours. More commonly however, amalgamation surfaces are indicated by the presence of abraded fossil fragments, shale pebbles, or discrete quartz pebbles up to

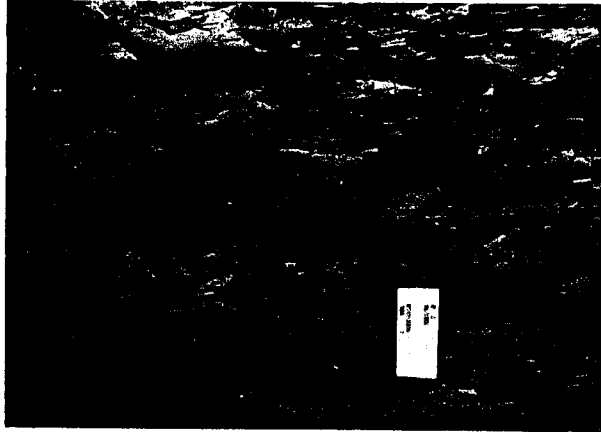


Fig. 23. Centimeter thick amalgamated sandstone beds of Facies Association 3 in mudstone-dominated section of the study area, Port Jervis, New York.



Fig. 24. Decimeter thick amalgamated sandstones of Facies Association 3 in sandstone-dominated area, Dalmatia Quarry, Pennsylvania.

one centimeter in diameter lying along sandstone laminae. Locally, intervals of this facies are intensely bioturbated and riddled with Planolites and "Teichichnus-like" burrows, obliterating internal stratification.

Interpretation Facies Association 3

Facies Association 3 is interpreted as storm emplaced deposits of the lowermost shoreface, straddling the storm-fairweather "mudline". Bioturbation increases with the amount of fine grained sediment present in the facies, reflecting deposition in less agitated environments just seaward of the mudline. The amalgamated hummocky cross stratified units probably represent deposition in areas landward of fairweather wave base, where the substrate was more or less continually agitated. Vertical coarsening upward trends are locally well developed, with grain size increasing upward.

Facies Association 4

Facies Association 4 contains the "cleanest" sandstones in the study area and displays sedimentary structures ranging from swaley

to massive sandstone in the sand-dominated areas of south-central Pennsylvania to soft-sediment features such as convoluted bedding and ball-and-pillow structures in the mudstone-dominated areas of north-eastern Pennsylvania and south-eastern New York. Three facies have been recognized in this association.

Ball-and-Pillow Sandstones

This facies (Figs. 25, 26) is present only in the mudstone-dominated sections of the study area of north-eastern Pennsylvania and adjacent New York. It consists of up to several meters of ball-and-pillowed very fine to fine sandstone. In places, convoluted bedding can be observed, but for the most part the intense soft sediment deformation has obscured all internal lamination. Although many beds appear massive, relict centimeter scale bedding, and scattered abraded and unabraded pelecypod and brachiopod shells indicate that these units are amalgamated. Ball-and-pillow structures extend a few to several decimeters into the underlying beds and are traceable for several meters along outcrops. Minor mudstone intervals of indeterminate thickness were present but are now concentrated as flame structures extending a few to several decimeters between the ball-and-pillow structures.



Fig.25. Ball-and-Pillow sandstone in mudstone-dominated section of the study area, Port Jervis, New York. Note that original internal structure of the beds has been retained.

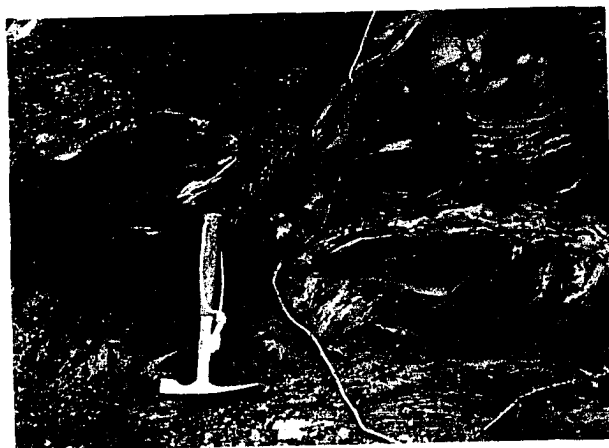


Fig. 26. Ball-and-Pillow sandstone, Port Jervis, New York. Note that all evidence of primary sedimentary structures have been obliterated by soft sediment deformation.

These flame structures usually are common lower in the facies, and decrease in abundance upward.

Swaley Sandstone

This facies (Fig. 27) is commonly a few meters thick and consists of clean, well-sorted fine to upper fine sandstone. Internally, laminae are rarely observed but the beds exhibit a meter-scale swaley geometry. Mudstone is preserved in places within the concave-upward swales (Fig. 28). The overall swaly geometry of the beds is present for distances of several meters along outcrops. Amalgamation of the sandstones is indicated by scattered articulated and disarticulated body fossils as well as rare occurrences of quartz pebbles a few millimeters in size along the basal surfaces of individual beds. The tops of the beds in this facies commonly are beveled, with the result that the convex-upward hummocks are not preserved.

Trough Cross-Bedded Sandstone

The thickness of this facies is commonly one to a few meters. Medium grained "clean" sandstone displaying trough cross beds (Fig. 29) in sets one or two decimeters in thickness and one to a few



Fig.27. Swaley sandstone facies, Dalmatia Quarry, Pennsylvania. Facies Association 4.

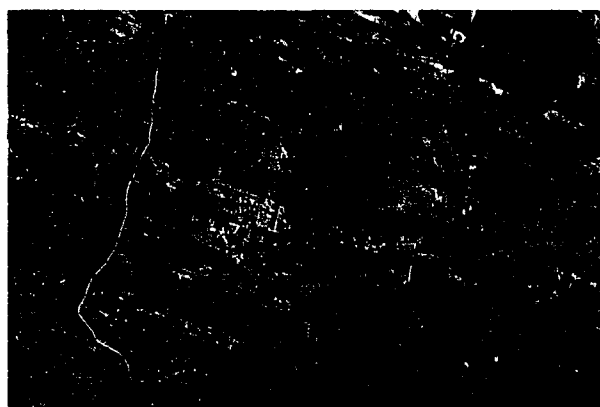


Fig. 28. Mudstone trapped within swales of Facies Association 4. Watts section.



Fig. 29. Trough cross-bedded sandstones of Facies Association 4. Dalmatia Quarry, Pennsylvania.

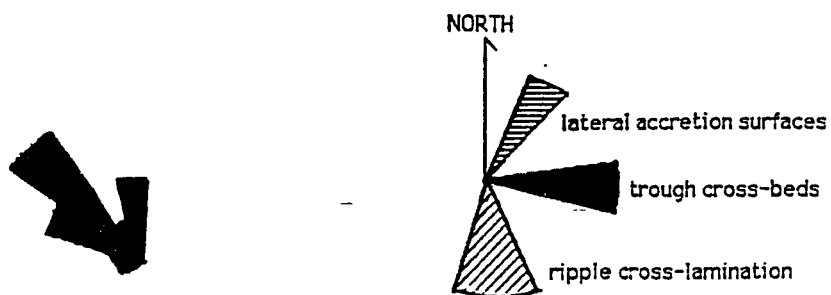


Fig. 30. Rose diagram of paleoflow in trough cross-bedded sandstones of Facies Association 4.

meters in lateral extent dominate the facies but structureless massive units are present as well. Centimeter size rounded shale pebble clasts are present in the bottom of troughs, although no mudstone appears in this facies as interbeds. Flow direction in the trough cross-bedded units is unimodal (Fig. 30). The troughs are filled with the upper fine to medium sandstone that comprises the bulk of this Facies Association. There is generally little change in the size of the trough cross-bedding either laterally or vertically within this facies

Interpretation Facies Association 4

Facies Association 4 is interpreted as being deposited on the mid to upper shoreface. The lack of a significant fraction of fine grained material and the larger grain size in this Facies Association suggest that wave agitation was more or less continuous, winnowing the finer grained sediments and truncating the tops of the swaley beds. The continuous agitation may have also destroyed primary sedimentary structures within the massive beds. Bioturbation is all but absent within this Facies Association suggesting that opportunistic organisms did not have time to

colonize and macerate the sediment, or that traces were subsequently obliterated by the agitation of the substrate by waves.

Facies Association 5

The facies of this association are thin, tabular fine to medium sandstone bodies with thin shale or mudstone partings. A diverse trace fossil assemblage of the Skolithos ichnofacies is present and brachiopods and ostracodes are locally abundant. Two facies have been recognized in this association.

Thin Sandstones with Mudstone Partings

This facies (Fig.31) consists of thin 10 to 20 centimeter thick, tabular, fine to medium sandstone. It forms a unit a few to several meters in thickness. The thickness of individual beds remains fairly constant both vertically and laterally. Sandstones are typically well sorted, and locally coarsening upward trends are developed within the facies. Beds in this facies exhibit a wide range of sedimentary structures, from faintly planar laminated, to structureless, to low angle laminations suggestive of hummocky cross-stratification. There is little variation laterally in terms of sedimentary

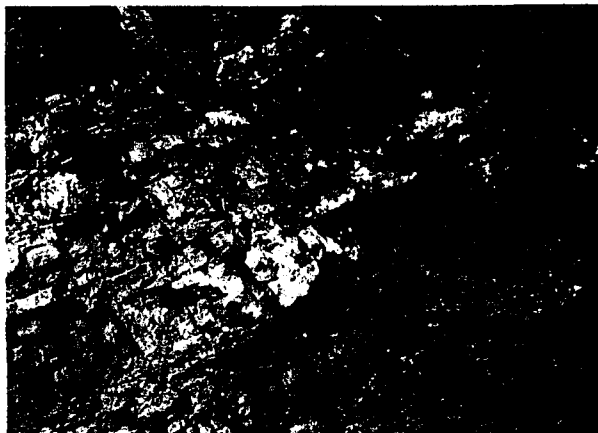


Fig. 31. Thin sandstones with mudstone partings of Facies Association 5. Girty's Notch section.

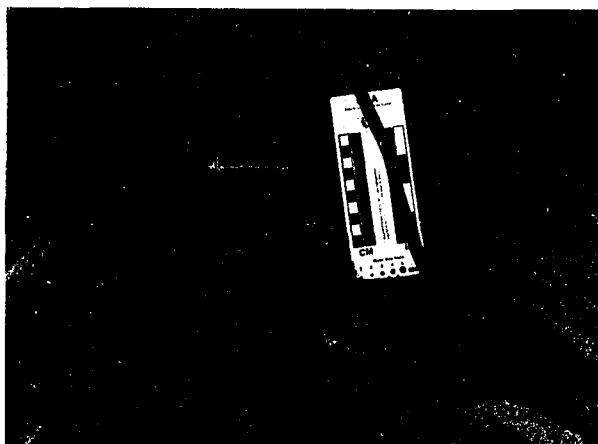


Fig. 32. Trough cross-bedded facies showing troughs lined with fossils and quartz granules. Dalmatia Quarry, Pennsylvania.

structures within individual beds. Symmetrical and asymmetrical ripples are found on the tops of beds in the upper parts of this facies, perhaps indicating a shallowing upward trend. Unabraded brachiopod shells are abundant locally along the basal surfaces of beds, as are ostracodes. Very thin millimeter partings of mudstone drape the top of most beds in this facies. Most are continuous along outcrops, but some pinch out due to erosion by overlying beds.

Trough Cross-Bedded Bioturbated Sandstone

This facies (Fig. 32) is dominated by trough cross-bedding and intense bioturbation. Individual beds are commonly 10 to 20 centimeters in thickness, with little variation either vertically or laterally. Grain size in this facies is consistently medium sandstone. This facies, where present, always overlies the thin sandstones with mudstone partings (Fig.33), and is always of a larger grain size. The trough cross-bedding is commonly 10-15 centimeters in thickness and troughs are generally one to a few meters in lateral extent. Abraded and unabraded, articulated and disarticulated shell material and quartz granules locally line the bottom of troughs, as well as rounded shale pebbles of centimeter

size. The paleoflow direction is bimodal in these trough cross-beds (Fig. 34). Beds of this facies are intensely bioturbated, with Teichichnus, Rosselia, Skolithos, and Planolites as common forms. Coquinite lenses several centimeters in thickness and several decimeters to a few meters in lateral extent are present in this facies. The coquinite lenses consist mainly of entire articulated shells.

Interpretation Facies Association 5

Several lines of evidence suggests that this Facies Association is of tidal origin. Bimodal paleoflow indicators in the trough cross-bedded units (Fig. 34) imply the ebb and flow of tidal currents. Additionally, the presence of ostracodes suggests a proximal offshore setting, possibly in or near brackish water. The variation in sedimentary structures indicate a variety of depositional conditions existed and that the flows were of variable intensity. Well developed vertical coarsening upward trends within this Facies Association suggest that the thin bedded sandstones with mudstone partings were deposited just seaward of tidal channels, and their thin mudstone partings imply that flow was

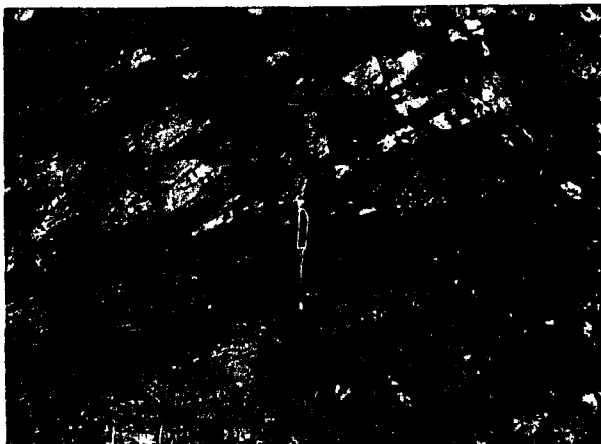


Fig. 33. Trough cross-bedded sandstone facies overlying thin sandstone with mudstone partings. Girty's Notch section.

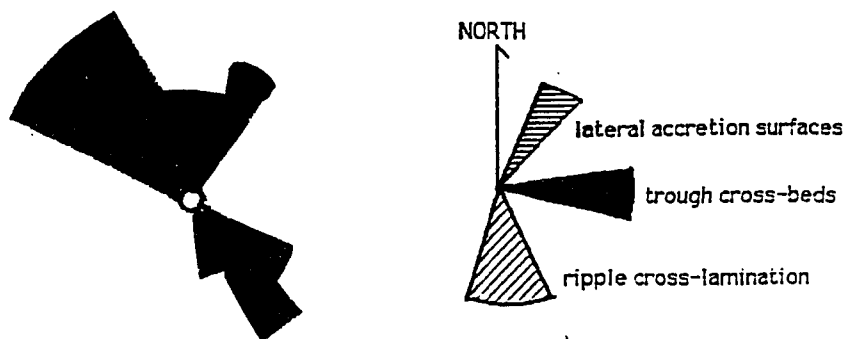


Fig.34. Rose diagram showing bimodal flow indicators in trough cross-bedded facies, Facies Association 5

slack long enough for fine grained sediment to settle, but not long enough for organisms to colonize and macerate the fine grained sediments. The overlying trough cross-bedded facies probably represent the seaward extent of tidal channels.

Facies Association 6

This Facies Association consists mainly of sandstones greater than one meter thick. Grain size in this Facies Association ranges from medium to pebbly sand. Trough cross-bedding is common with quartz pebbles, fossil hash and shale rip-up clasts lining the bottom of troughs in places. Some beds in this Facies Association display lateral accretion surfaces, having a restored dip of approximately five to fifteen degrees (Duke and Prave, 1991). These lateral accretion surfaces are a few to several meters in thickness (Fig. 35). Reversing flow directions are indicated by the opposing dips of superjacent bed sets. Thin sandstone beds may be flat laminated, structureless, or capped with ripples yielding bimodal paleoflow direction. Bioturbation is not extensive in this Facies Association but does occur in a few beds. Two facies have been defined in Facies Association 6.

Trough Cross-Bedded to Rippled Sandstone

This facies (Fig. 36) ranges from a few meters up to ten meters in thickness and consists of trough cross-bedded, poorly sorted sandstones. Trough sets are decimeters in size. Quartz pebbles are well rounded but some display angularity. The largest pebbles have long axes of up to five centimeters. Locally the boundaries between sets of trough cross-bedding are capped by asymmetrical ripples. The paleocurrent indicated by these capping ripples is opposite the sense of paleoflow in the trough cross-beds (Fig. 37). Mud drapes and reactivation surfaces are present on some cross-beds. Rounded shale pebbles up to a few centimeters in size locally line the basal surfaces of channels. Lateral accretion surfaces are common, and are a few to several meters in thickness. The bottom contacts are erosive, and conglomerates are locally present just above the erosive surfaces. Abraded brachiopod shells are also abundant in layers a few centimeters thick just above the erosion surfaces.



Fig. 35. Photo of Facies Association 6, showing lateral accretion surfaces. Girty's Notch section.

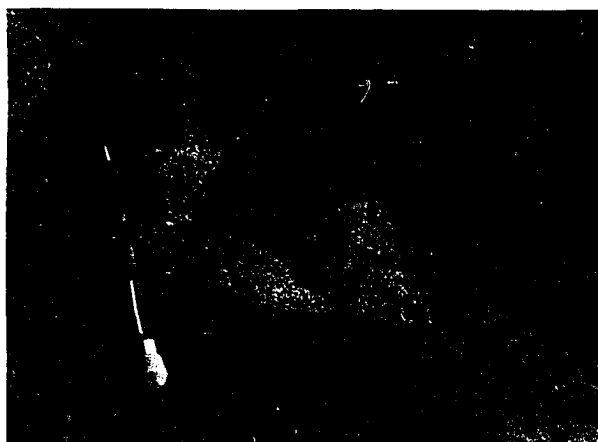


Fig. 36. Trough cross-bedded to rippled sandstone facies. Note shale pebbles and conglomerate lining troughs. Dalmatia Quarry, Pennsylvania.

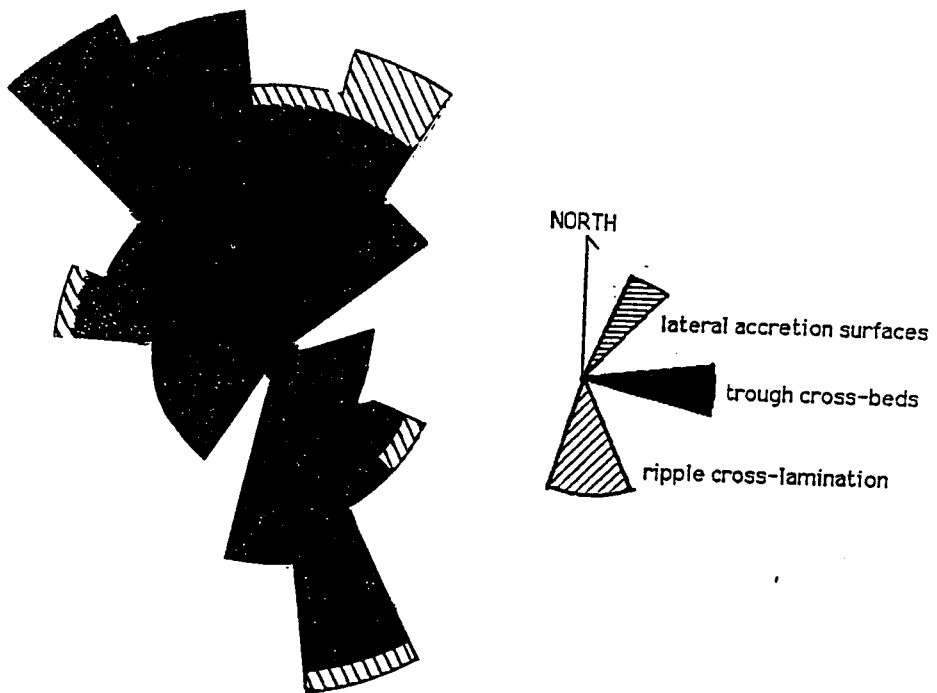


Fig. 37. Rose diagram of paleoflow, Facies Association 6 ripples, trough cross-bedding and lateral accretion surfaces.

Typically, fining upward trends are well developed. The dips of lateral accretion surfaces are at nearly 90 degree angles to the flow directions indicated by the trough cross-bedding (Fig. 37). Laterally, the epsilon cross-bedded lateral accretion surfaces truncate trough cross-bedded sandstone beds. Vertically, the lateral accretion surfaces have rippled tops, or are truncated by trough cross-bedded sandstone.

Poorly Sorted Pebbly Sandstone

This facies consists of poorly sorted and pebbly sandstone. The thickness of this facies ranges from a few to several meters. Sedimentary structures are generally lacking, but channels are incised in places along basal contacts of this facies. The channels are symmetrical in form and approximately several meters wide and roughly one meter deep. Shale rip-up clasts and quartz granules occur at the bottom of the channels.

Interpretation Facies Association 6

This Facies Association is interpreted as deposits of a subtidal to possibly intertidal flat based on the following criteria.

Reversing paleoflow directions (Fig. 37) are indicated by opposing dips of superjacent bed sets in this facies suggest the ebb and flow of tidal activity. Lateral accretion surfaces represent point bar migration of a channelized tidal distributary system several meters deep. The size and abundance of pebbles and poor sorting of the channel fill suggest direct connection with a fluvial system that debouched into the sea through this estuary or distributary system (Duke and Prave 1991).

Facies Association 7

This Facies Association consists of poorly sorted, pebbly, sandstone or coquinite beds that vary in thickness from a few centimeters to a few meters. The sandstones are generally poorly cemented and weather "vuggy". All occurrences are sharp based and undulatory erosional relief of several centimeters between this Facies Association and underlying rocks is common. This Facies Association occupies a distinctive stratigraphic position at the top of coarsening-up cycles, and most, but not all coarsening-up cycles are capped by it. Grain size sharply decreases and abrupt shifts to deeper water facies occur immediately above it (Fig.38).

Bioturbation is variable and some beds are crudely cross-bedded. Three facies have been recognized in Facies Association 7.

Single-Storey "Lag"

This facies (Fig. 39) ranges from a few centimeters to as much as a meter in thickness and consists of a single erosive based deposit of pebbly coarse sandstone or fossil hash. Erosion of the underlying bed is marked by an undulatory (centimeter to decimeter scale) surface traceable along the length of outcrops. The sandstone is poorly sorted with pebbles of multi-colored vein quartz and rounded shale rip-up clasts locally abundant. This sandstone is usually not well cemented and the facies weathers to a "vuggy" reddish condition due to abundant hematitic cement. Iron oolites are also locally present and these Mahantango ironstones have been noted by Kaiser (1972) and Prave and Duke (1990). Phosphate nodules may be present locally. Fossils are predominantly abraded brachiopod shells but crinoid ossicles and coral fragments up to a few centimeters in diameter are locally abundant. At some locations abraded fossils comprise the largest fraction of this facies (Fig. 40), with poorly sorted sandstone and shale pebbles subordinate.

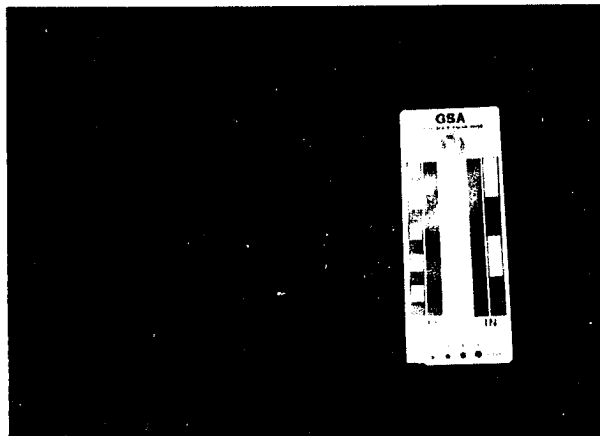


Fig. 38. Single-storey lag of Facies Association 7 showing overlying mudstone. Note bioturbation of underlying sandstone. Watts section.

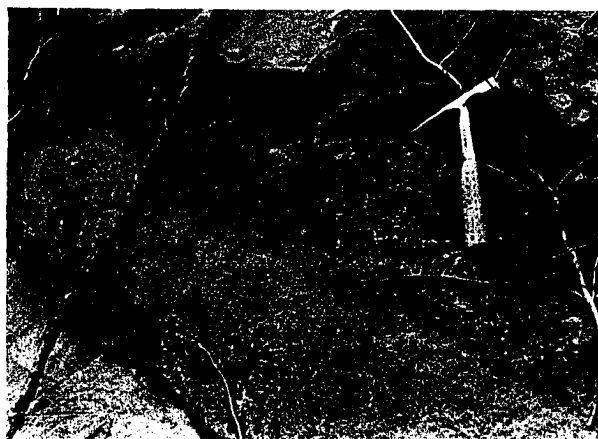


Fig. 39. Single-storey lag, Facies Association 7 Dalmatia Quarry, Pennsylvania.



Fig.40. Single-storey coquinite lag, Facies Association 7
Port Jervis, New York.



Fig. 41. Multi-storey lag of Facies Association 7 with siltstone
interbed. Port Jervis, New York

Locally, the basal portion of this facies is intensely bioturbated with decimeter scale burrows extending down into the underlying rocks (Fig. 39). The burrow fill is usually the same size and composition as the overlying "lag" and clearly connects the bioturbation features to depositional conditions associated with Facies Association 7.

Multi-Storey "Lag"

This facies (Fig. 41) occurs as two (or more) distinct beds, separated by decimeter to meter thick beds of mudstone or siltstone. The lithologic composition, position in the stratigraphic succession, and erosional relationship with underlying units are identical to those of the preceding facies, and these descriptions will not be repeated. The intervening mudstone or siltstone may be poorly bedded and possess varying degrees of bioturbation.

Cross-Bedded "Lag"

This facies ranges in thickness from 1 to 4 meters and consists of coarse pebbly sandstone and fossils identical to single-storey lags described above. However, beds display crude cross-

bedding at a low angle to underlying beds.

Interpretation Facies Association 7

Based on the anomalously coarse grain size, coquinite beds, and distinctive stratigraphic position, this Facies Association is interpreted as a hydrodynamic lag concentrate associated with retreat of the shoreface during relative sea level rise. Transition to deeper water facies immediately above Facies Association 7 indicate that relative sea-level rose and facies backstepped landward above each occurrence of this Facies Association. Evidence for the nature of the relative sea-level rise may be present in the multi-storey lag facies. This facies has been interpreted as transgressive "shoals" that migrated across the shoreface, trapping finer grained sediments (Duke and Prave 1991). Another possible explanation is that Facies Association 7 was regionally continuous along the paleoshoreface, and that relative sea-level rise was not monatonic but took place initially in a series of small amplitude rises and falls. Subsequent relatively rapid sea-level rise resulted in abrupt transition to deeper water facies immediately above Facies Association 7. Evidence for a possible step-wise rise in

relative sea-level exists in the mudstone-dominated sections of the study area of north-eastern Pennsylvania and south-eastern New York. There, the multi-storey lag of Facies Association 7 is locally overlain by several decimeters of Facies Associations 3 and 4 before an abrupt shift to deeper water facies occurs. The abraded fossil fragments abundant in the lags also indicate a period of reworking before burial by rapid relative sea level rise. Based on the large grain size in this Facies Association, considerable erosion and hydrodynamic sorting must have taken place during shoreface retreat and relative sea-level rise.

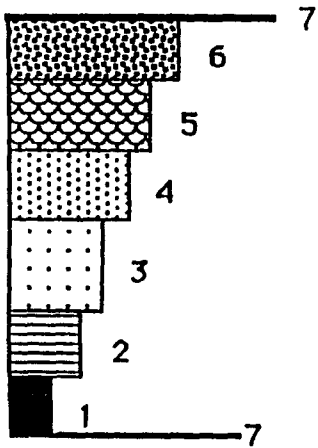
A GENERIC MAHANTANGO CYCLE

The suite of Facies Associations described above are interpreted as a continuum of depositional environments occurring along the eastern margin of the subsiding Appalachian foreland basin. Facies Associations recur stratigraphically in consistent patterns. Successive vertical cycles of a few to several tens of meters in thickness (average thickness 10 meters), are arrayed in coarsening-upward, shallowing-upward packages. The depositional

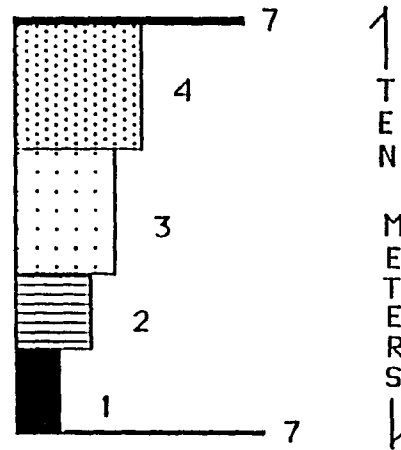
environments within each vertical cycle range from mid to outer shelf deposits, to more proximal wave-dominated shoreface or tide-dominated depositional settings. Facies Associations 1 and 2 are composed mainly of hemipelagic sediment, although storm beds occur in both Facies Associations. Hemipelagic sediment is subordinate to coarser grained sediments in Facies Associations 3 and 4. These Facies Associations are dominated by sedimentary structures suggesting they were emplaced by storms and more or less continually agitated by fair weather waves. Tide-dominated sedimentary structures make their appearance in Facies Associations 5 and 6. Bimodal trough cross-bedding (Figs. 34, 37) and the presence of lateral accretion surfaces suggest deposition in sinuous channels, while the variety of sedimentary structures such as planar cross-bedding, low-angle laminations, and structureless beds, argues for flows of varying intensity. These observations, in addition to the mudstone partings occurring in Facies Association 5 are indicative of flow expansion and slack water deposits and point to a tidal origin for Facies Associations 5 and 6. The full spectrum of Facies Associations are preserved in the sandstone-dominated units of south-central Pennsylvania. Although thicknesses of the

Facies Associations vary between outcrops, this probably reflects their paleolocation both with respect to the basin margin and fluvial sources. In the mudstone-dominated units of north-eastern Pennsylvania and Port Jervis, New York, tide-dominated Facies Associations 5 and 6 are not present (or are not preserved). The hydrodynamic lag deposits of Facies Association 7 are present in all parts of the study area, but exhibit a greater lithic variation in the sandstone-dominated areas, while in the mudstone-dominated areas they are mainly coquinites without the large pebbles and varied lithic components of south-central Pennsylvania (Fig. 40). Figure 42 shows the "ideal" Mahantango cycle in both the mudstone-dominated areas of north-eastern Pennsylvania and south-eastern New York, and sandstone-dominated regions of south-central Pennsylvania in terms of the Facies Associations present and the average thicknesses of those associations in the measured sections and incomplete exposures studied. Each cycle begins with Facies Association 7, and is overlain by deeper water facies, that grade upward to shallower water deposits. Grain size also increases upward in each cycle.

SANDSTONE DOMINATED



MUDSTONE DOMINATED



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Figure 42. An Ideal Mahantango facies association succession. note that tidal facies 5 and 6 are missing or not preserved in the mudstone-dominated sections.

Each "ideal" cycle is terminated by an abrupt return to deeper water facies immediately above the cycle bounding Facies Association 7 boundary.

DEPOSITIONAL CYCLICITY

Past workers have noted the cyclicity in the Mahantango Formation and have suggested various mechanisms for its origin. Kaiser (1972) and Faill et al. (1978) have suggested that the cycles developed as a result of the progradation of delta lobes. Goldring and Bridges (1973) suggested a prograding shoreline origin for the cycles. More recently, Sarwar and Smoot (1983), and Sarwar (1984) have attributed the genesis of cycles in the Mahantango Formation to the migration of storm-generated sand ridges. Duke and Prave (1991) have attributed the cyclicity to the progradation of a regionally straight shoreline. Given the large areal extent of the cyclicity in the Mahantango Formation, it is difficult to ascribe their genesis to migration of storm produced sand ridges. Such ridges would have to have migrated tens to perhaps hundreds of kilometers to produce the cyclicity (and pattern of cyclicity) extant

in the Mahantango Formation. The earlier interpretation that the cyclicity was produced by prograding shorelines is essentially substantiated by this study, although the progradational events were probably not simply the autocyclic progradation of deltaic lobes into the Catskill Sea. Duke and Prave (1991) utilized paleoflow indicators to argue that progradation of the shoreline was straight rather than lobate, and remained notably so through time and space. The observations made in conjunction with this study show that the sediments of the Mahantango Formation filling the Appalachian foreland basin during Givetian time were progradational and episodic, with gradual shallowing interrupted by abrupt marine flooding events. The pattern of cyclicity developed in superjacent cycles is remarkably consistent all along the eastern margin of the basin, as well as along the north-western margin of the basin near Buffalo, New York. This pattern of cyclicity developed in Hamilton Group rocks separated by hundreds of kilometers precludes the origin of most, if not all individual cycles by local autocyclic mechanisms. The processes that created them must therefore have been allocyclic in nature. What those may have been will be addressed later in this study.

SEQUENCE STRATIGRAPHY OF THE MAHANTANGO FORMATION

Figures 43 through 49 depict the measured sections detailed in this study. The Facies Associations within each cycle clearly start in deeper water, and shallow and coarsen upward in each cycle, defining parasequences. There are however, patterns of parasequence stacking that transcend grain size or facies in both the mudstone-dominated and sandstone-dominated regions of the study area. Figure 50 is a correlation of measured sections throughout the study area. Using a thick, regionally present lag of Facies Association seven as a datum (chosen for reasons stated below) the following parasequence stacking pattern is evident. Below the datum, parasequence sets are progradational, that is, each parasequence starts in "deeper" water and ends in "shallower" water, but each successive parasequence starts in "shallower" water than the parasequence below it. This is shown either by a reduction in the thickness of the "deeper" water Facies

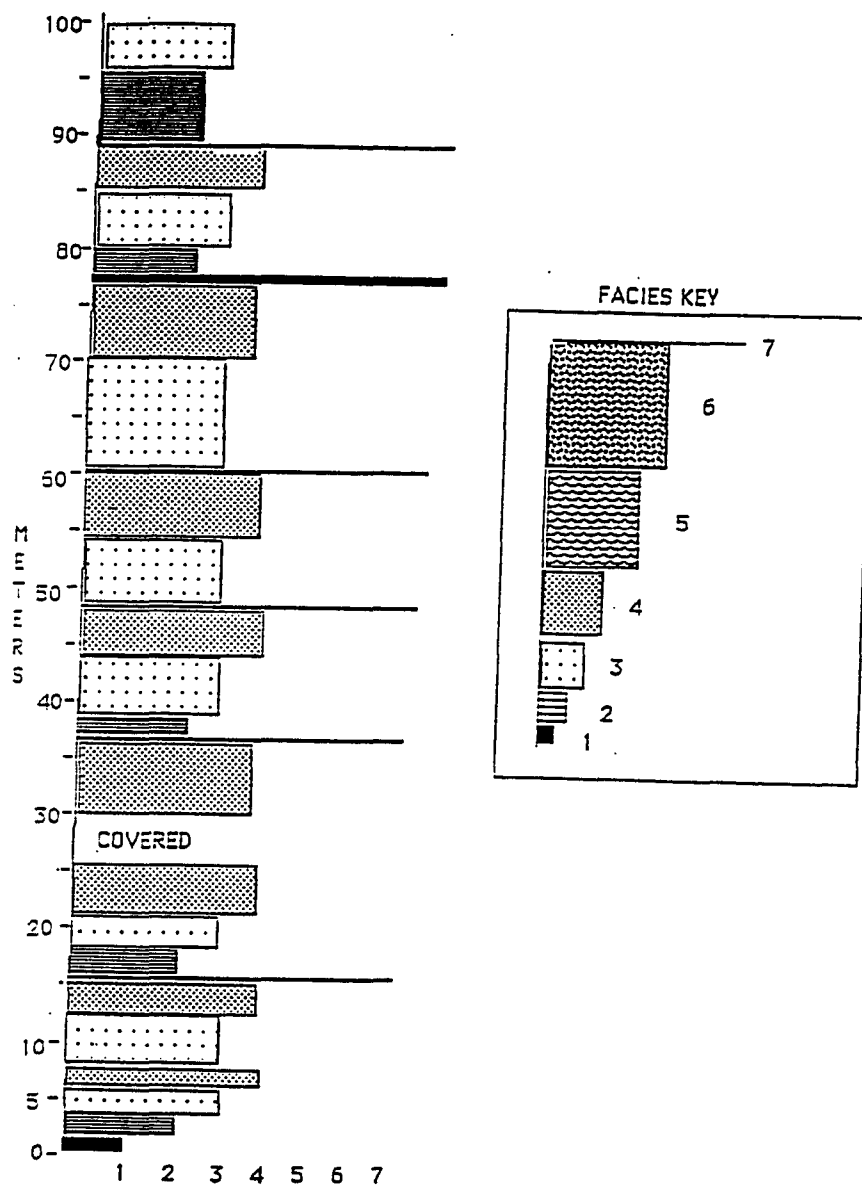


Fig. 43. Measured section Port Jervis, New York. Facies Associations present in section indicated as per above key.

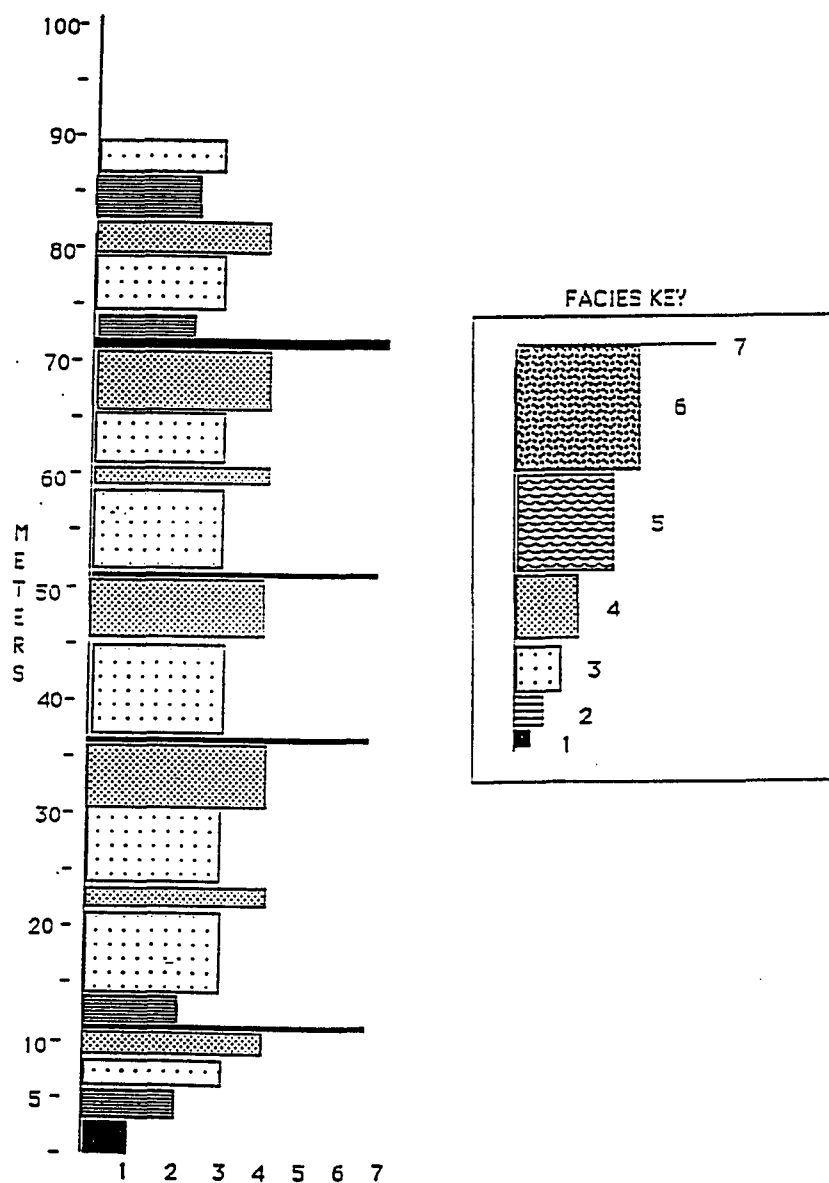


Fig.44. Measured section Sawkill Creek, Milford Pennsylvania. Facies Associations present in section indicated as per above key.

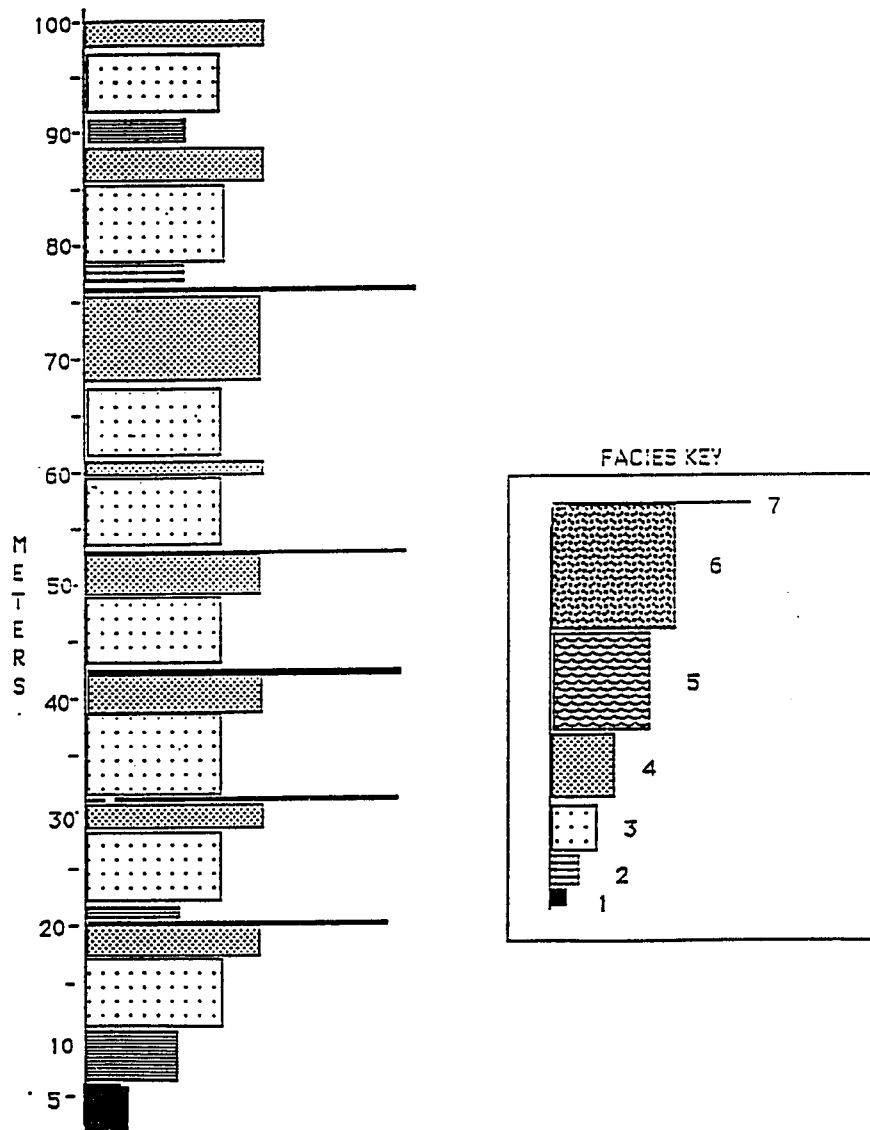


Fig.45. Measured Section Raymondskill Falls, Milford, Pennsylvania. Facies Associations present in section indicated as per above key.

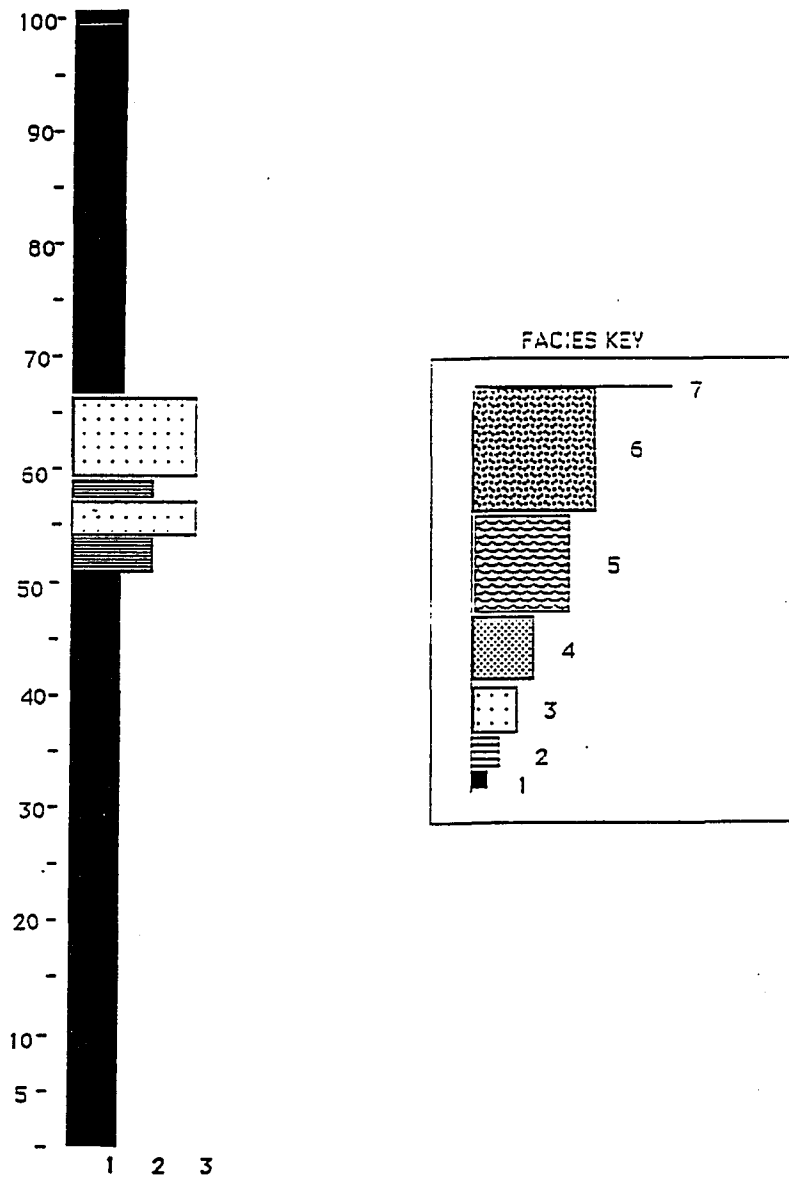


Fig.46. Measured Section Bowmanstown, Pennsylvania
Facies Associations present in section indicated as
per above key.

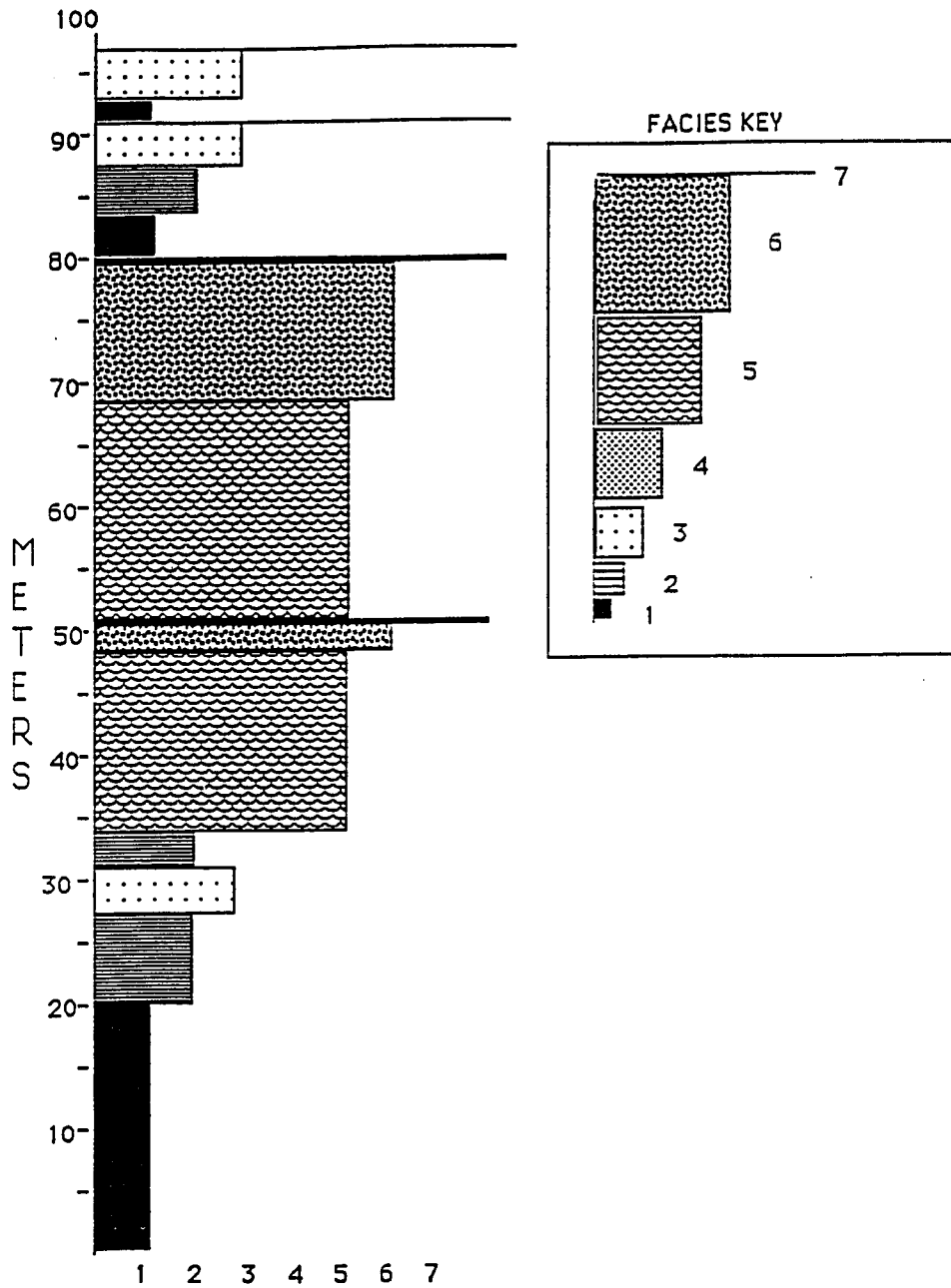


Fig.47a. Measured Section Rockville, south-central Pennsylvania. Facies Associations present in section indicated as per above key.

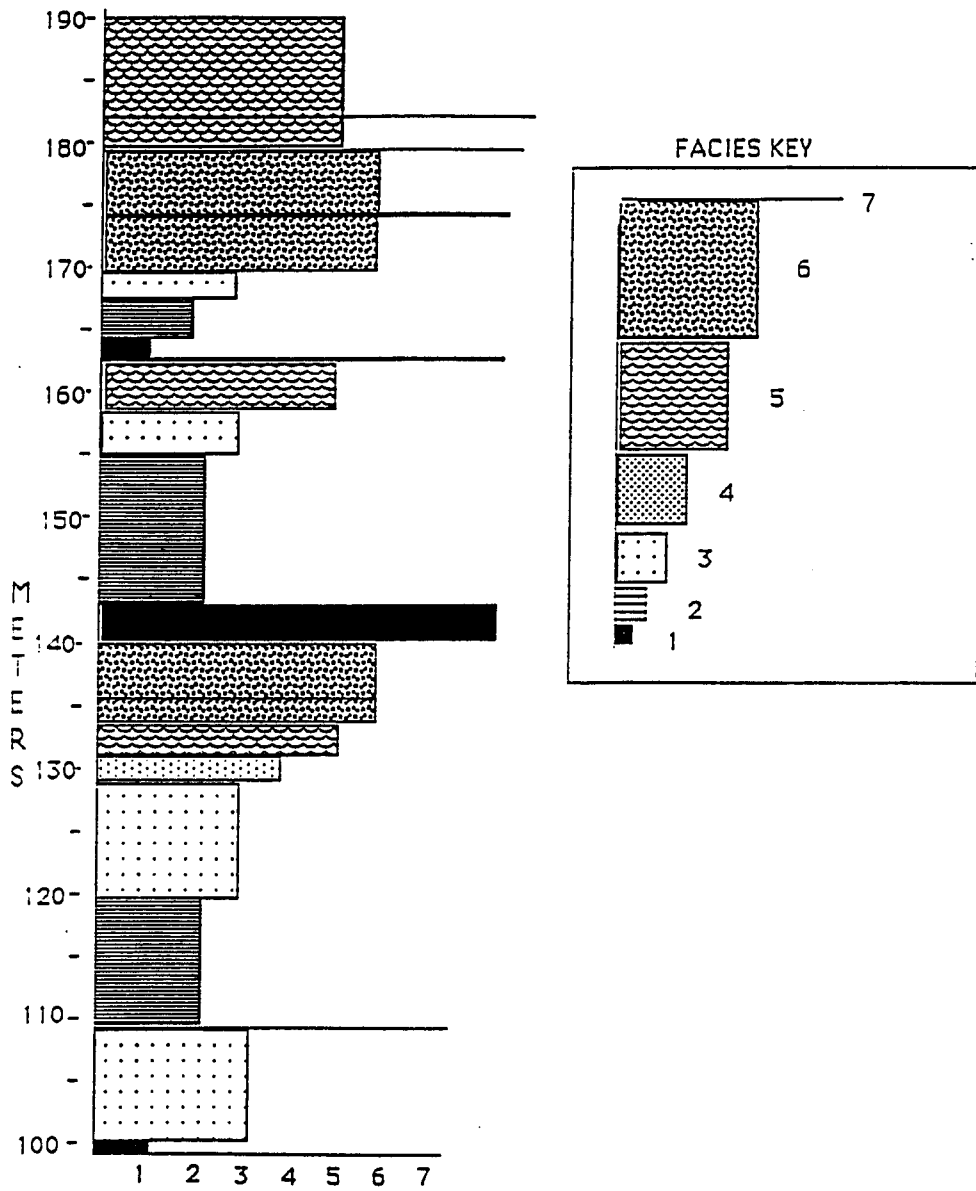


Fig.47b. Measured Section Rockville, south-central Pennsylvania. Facies Associations present in section indicated as per above key.

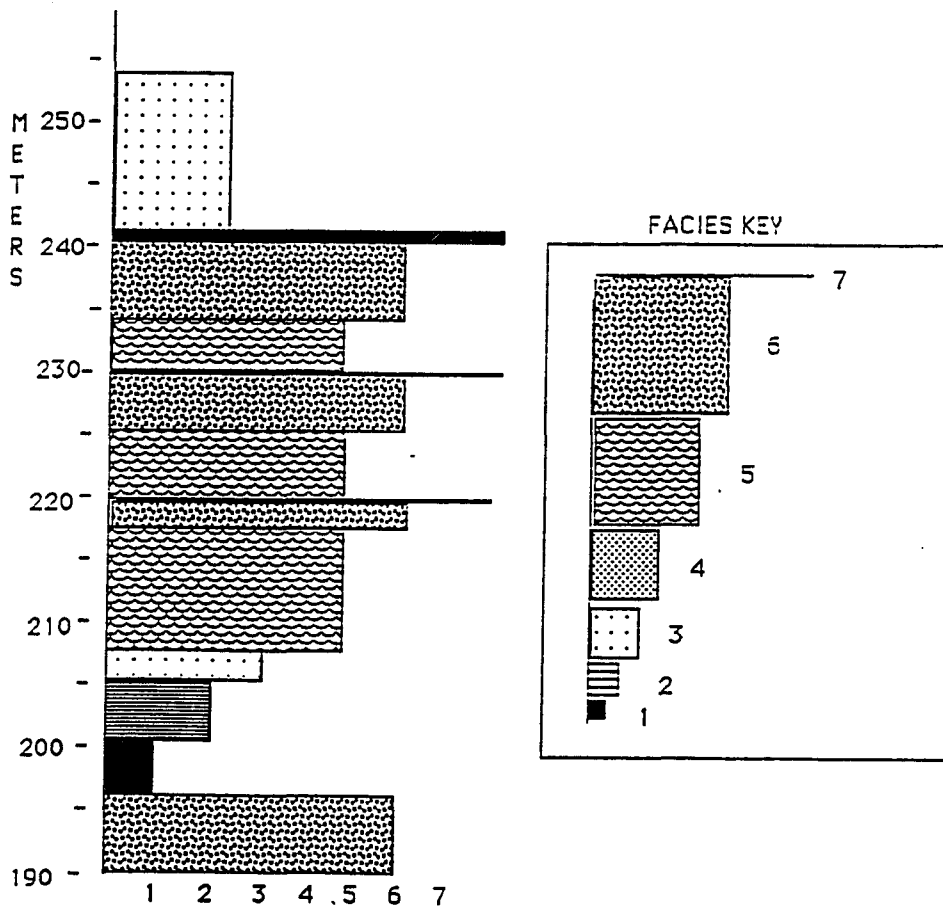


Fig.47c. Measured Section Rockville, south-central Pennsylvania. Facies Associations present in section indicated as per above key.

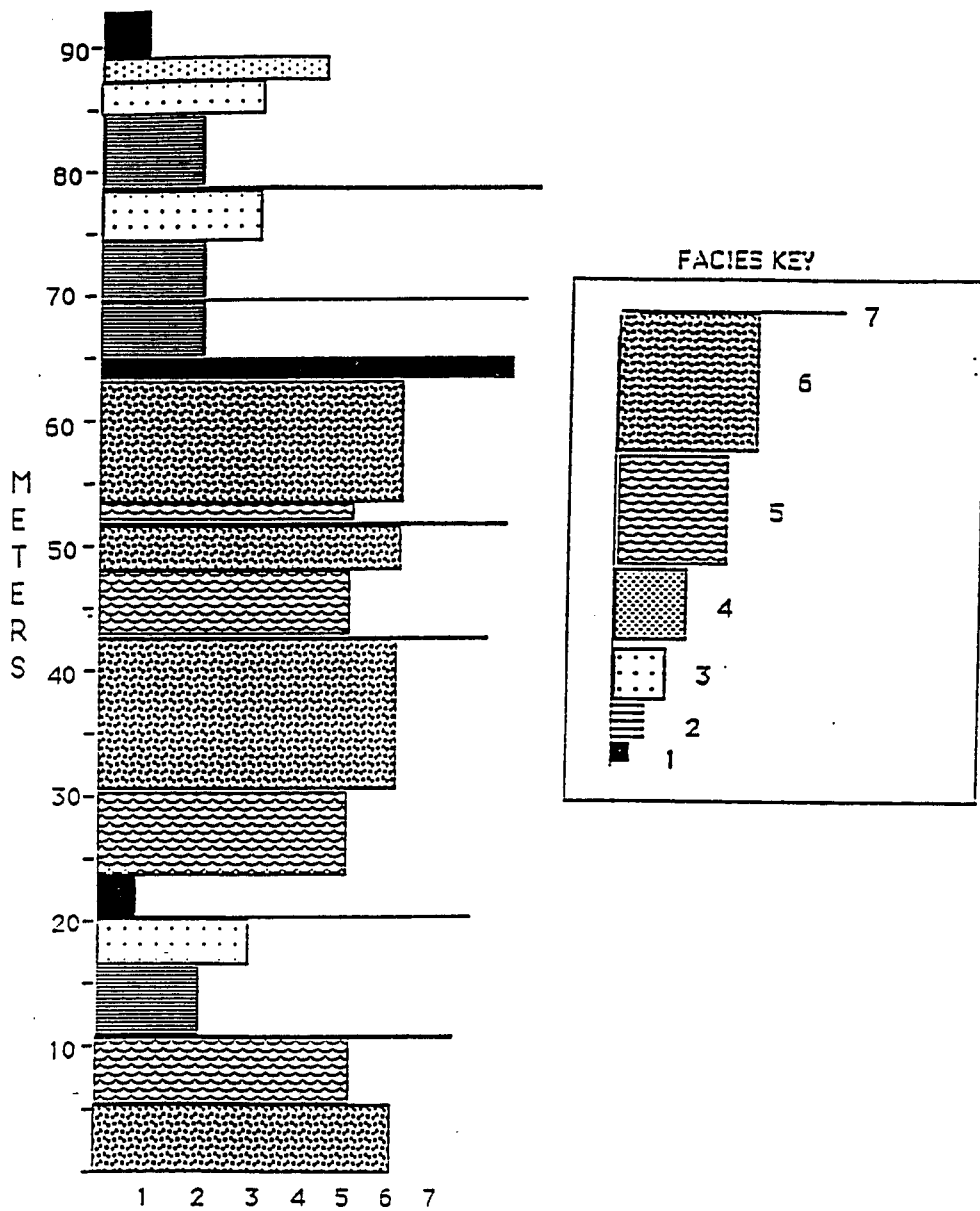


Fig.48. Measured Section Girty's Notch, south-central Pennsylvania. Facies Associations present in section indicated as per above key.

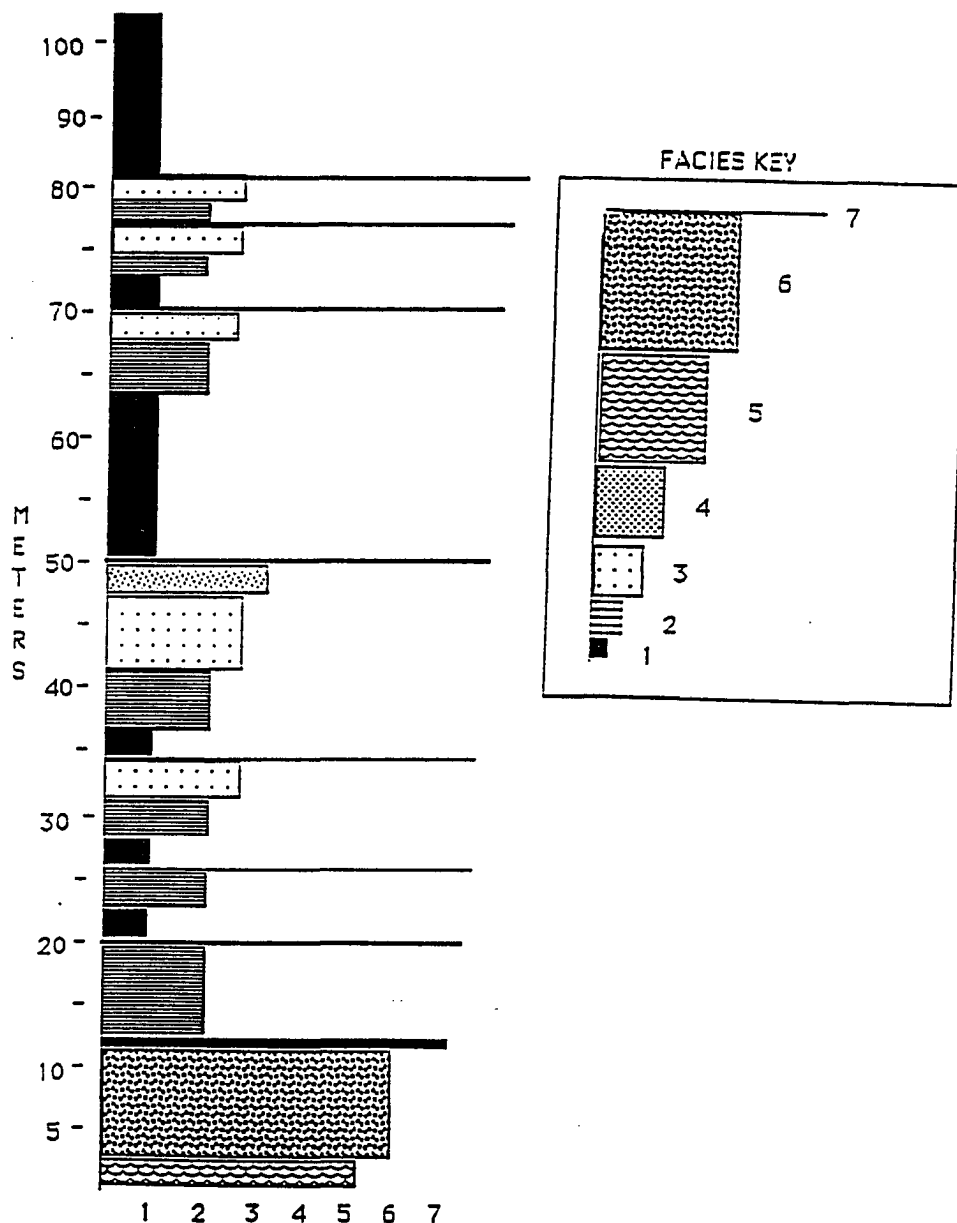


Fig.49. Measured Section Watts, south-central Pennsylvania. Facies Associations present in section indicated as per above key.

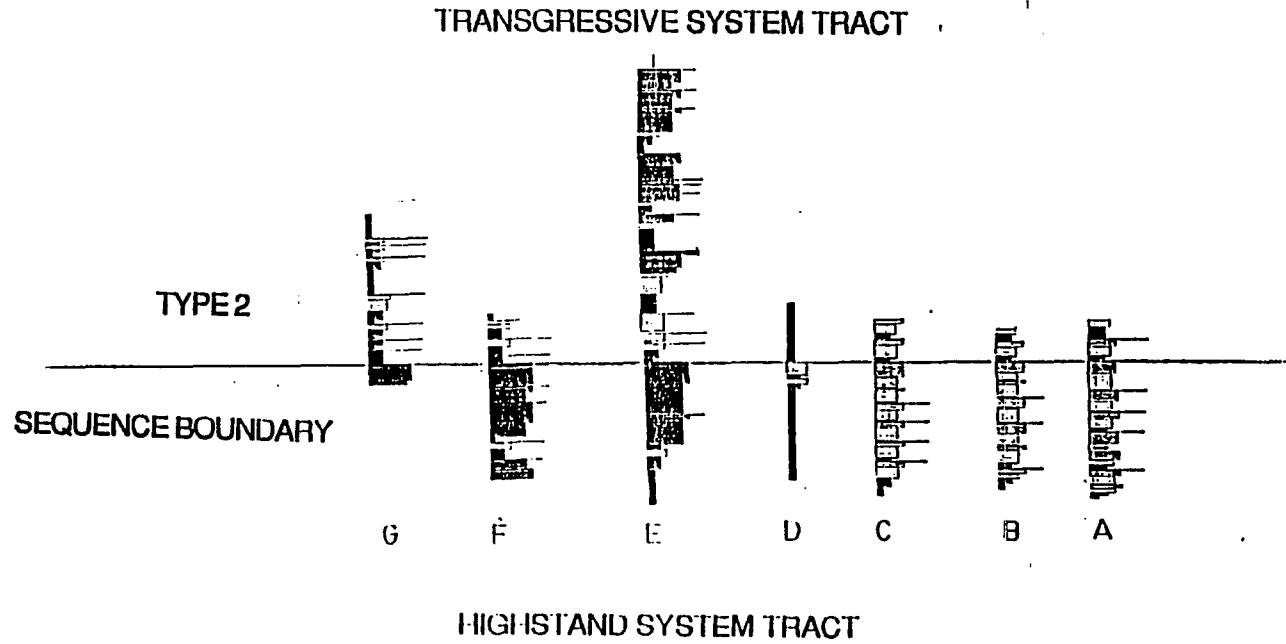


Fig. 50. Sequence stratigraphy of the Mahantango Formation. Datum is regionally extensive lag deposit. Locations of measured sections A-G given in Appendix 1.

Associations, or by the presence of "shallower" water Facies Associations at the start of cycles. This pattern is evident throughout the study area, both in the mudstone and sandstone-dominated areas. Above the datum, parasequences are retrogradational. Although they individually shallow and coarsen upward, each successive parasequence starts in "deeper" water than the one below, defining a retrogradational parasequence set. Therefore, below the datum, the parasequence stacking pattern in the Mahantango Formation defines a Highstand system tract, and above the datum the retrogradational parasequence stacking pattern defines a Transgressive system tract. Since there has been, to date, no evidence of fluvial incision or stream rejuvenation observed along the datum, it tentatively is considered a Type 2 sequence boundary. It should be noted at this point that the terms Highstand and Transgressive system tracts as originally defined considered eustasy as the driving mechanism for their genesis. However, with the Acadian Orogeny at or near its height in the upper Middle Devonian, and the proximity of the study area to the orogenic front, there could have been some tectonic overprinting of the eustatic component at work generating these system tracts. These variables

and others will be explored in subsequent sections. Whatever the fractional contributions are of the sedimentation rate, eustasy, and tectonics, the Mahantango Formation along the eastern edge of the Appalachian foreland basin in late Middle Devonian time defines two depositional sequences separated by a Type 2 sequence boundary. Brett and Baird (1990, their figure 3), independently derived exactly the same pattern of Highstand and Transgressive system tracts separated by a Type 2 sequence boundary for similar age strata along the northwestern margin of the Appalachian foreland basin. This relationship is shown in Fig. 51. Clearly, forces are at work that transcend local autocyclic mechanisms in defining the facies architecture and stratigraphic patterns within this foreland basin. Autocyclic mechanisms capable of producing patterns of cyclicity such as those observed in the study area include climate and sedimentation rate, the rheology of the lithosphere (including tectonics), and eustatic sea-level changes.

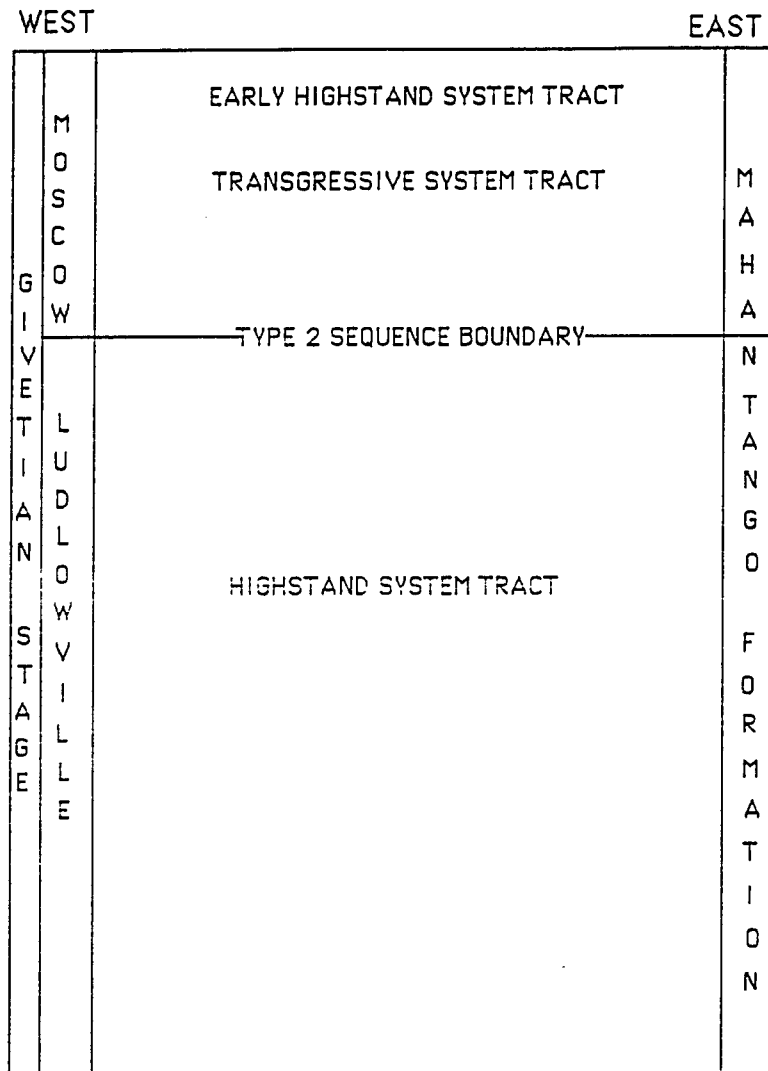


Fig. 51. Sequence Stratigraphic model of the Western and Eastern margins of the Appalachian Basin during Givetian time.

SEDIMENTATION RATES AND CYCLICITY

Climate and sedimentation are intimately related. Source areas respond to climatic influences in varying ways, including changes in runoff, discharge rate, and the weathering characteristics of the rocks in the source area. The Appalachian foreland basin was located between an orogenic peninsula and the craton of ancestral North America in low southern latitudes during the time of deposition of the Mahantango Formation and was probably subject to tropical climatic conditions with seasonal rainfall and intense tropical storms (Woodrow et al. 1988). Jervey (1988), suggested that the amount of sediment supplied to basin locations was a function of the general rate of sediment influx and proximity to zones of active transport. Flemings and Jordan (1990) attributed facies migration in foreland basins to the rate of thrust sheet propagation. Although the primary evidence needed to determine thrust sheet propagation rate and capacities of fluvial transport sources have been removed by erosion, it is possible to infer the effects of these variables on the sedimentation patterns and cyclicity observed in the study area. Assuming the

sedimentation model presented herein is essentially correct, sediment was delivered to the shoreline by fluvial systems and then redistributed by nearshore marine currents, waves, and tides. It is clear from measured sections at Rockville and Girty's Notch (Figs. 47, 48) that the effect of a net greater sediment influx to the proximal south-central sandstone-dominated Pennsylvania depocenter was the formation of more complex facies assemblages relative to those produced in the mudstone-dominated areas of north-eastern Pennsylvania and adjacent New York (Figs. 43-46). This more complex pattern of sedimentation in south-central Pennsylvania sections probably reflects the sedimentation rate approximately equalling accommodation rate, resulting in suites of shallow water facies that were sensitive to local autocyclic mechanisms such as the spatial distribution of Facies Association 4, and channel migration within Facies Associations 5 and 6. Perhaps aggradation to base level occurred in the sandstone-dominated areas of south-central Pennsylvania, however, such evidence for terrestrial fluvial deposition would have been removed by subsequent sea-level rise and shoreface retreat. Although it seems reasonable to conclude that increased sedimentation rates produced complex shallow water

facies suites, there must have been higher order controls driving the process of system tract and sequence generation within the Appalachian foreland basin during Middle Devonian time. This is clear from the overall similar pattern of parasequence set stacking patterns along the eastern rim of the basin. Irrespective of the proximity to the south-central Pennsylvania depocenter, facies along depositional strike in both sandstone-dominated and mudstone-dominated areas can be divided into the equivalents of a Highstand system tract and Transgressive system tract separated by a Type 2 sequence boundary. Similar age strata along the north-western rim of the basin display exactly the same sequence stratigraphic pattern (Brett and Baird, 1990). It is a safe assumption that the topographical relief and, therefore, erosion and sedimentation rates varied along the Acadian front as well as across the basin on its northwest rim, bordered as it was by cratonic lands. Perhaps there was some relief along the cratonic margin due to peripheral bulge uplift, but clearly orogenic scale relief was lacking there.

Brett and Baird (1985) traced the Centerfield limestone into its clastic equivalent across the basin axis and proposed the limestones correlated with the approximate tops of coarsening

upward cycles east of the basin axis. This relationship between limestone facies and coarsening upward clastic cycles has important implications for the correlation of beds in the Middle Devonian Appalachian foreland basin. Brett and Baird, (1985) suggested that limestone facies were developed immediately after periods of maximum regression of the shoreline during periods of subsequent sea-level rise, such as those associated with the Transgressive system tract developed immediately above the Type 2 boundary in Givetian time. Greater sedimentation rates along the eastern margin nearly, but not quite, swamped carbonate production as sea-level rose. The Clearville Siltstone member of the Mahantango Formation is the uppermost siltstone unit recognized in the Mahantango Formation in areas of Maryland, Virginia, West Virginia, and southern Pennsylvania. It occupies the approximate stratigraphic position as the Montebello sandstone member in south-central Pennsylvania (Hasson and Dennison, 1988, their figure 16), (Woodrow et al. 1988, their figure 4), and is probably its distal equivalent. Immediately above the Clearville Siltstone a nodular limestone, the Pokejoy Member, marks this rise in sea-level. Jolley (1982), suggested either delta lobe abandonment or eustatic sea-

level rise created the conditions necessary for the development of the Pokejoy member. Since the same overall sequence stratigraphic pattern is observed in areas with substantial differences in sedimentation rates and types, there must be a higher order variable controlling the observed higher order cyclicity, as discussed below.

RHEOLOGICAL MODELS

The Appalachian foreland basin formed on continental crust primarily in response to the loading of the lithosphere by orogenic thrust sheets, essentially involving a transfer of mass both horizontally and vertically. The observed thickness of several kilometers of accumulated sediments preserved within the basin clearly suggest that subsidence of the lithosphere occurred, but the rheological model that most accurately describes the lithospheric response to thrust sheet loading remains controversial. Proposed models include an elastic lithospheric response (Jordan 1981), visco-elastic response (Quinlan and Beaumont 1984), lithospheric lateral variation models (Stockmal et al. 1986), and thermal history models

of foreland basins (Komitz and Bond 1986). Although all of the above differ in detail, it is possible to generalize the response of the earth's lithosphere to loading. The emplacement of thrust sheets results in crustal thickening and a deviation from isostatic equilibrium. The lithosphere responds by subsiding under the load, but since the lithosphere has lateral strength, subsidence is not restricted to the loaded area but is distributed flexurally adjacent to the load. Surface expression of this flexural response is subsidence adjacent to the load, with the deflection of the lithosphere gradually diminishing with distance from the load, finally becoming a positive topographical feature (the peripheral bulge). Studies of post-glacial lithospheric rebound show that the lithosphere is capable of relaxing stresses by the mechanism of viscous flow over times of 10^3 - 10^4 years, yet is rebounding elastically. A substance that behaves elastically on short time scales but viscously over longer time scales is called a visco-elastic material. If there is no change in strain, a visco-elastic material relaxes stresses over its visco-elastic relaxation time. This parameter is a function of temperature, rheological parameters, initial stress, and stress history. The visco-elastic

relaxation time is not yet well constrained, and estimates vary from less than 10^6 years to 20-30 million years (Allen and Allen, 1990). If the visco-elastic relaxation time is short, than an elastic model is adequate to model the stratigraphy of the Appalachian foreland basin. However, if the visco-elastic relaxation time is long, then the visco-elastic response would have a profound effect on foreland basin stratigraphy. A temperature dependent visco-elastic model, in which stress relaxation propagates from the hotter, less viscous regions of the lithosphere to the colder, more viscous regions was used by Quinlan and Beaumont (1984) to explain the Paleozoic unconformities and stratigraphy in interior North America. They postulated a broad continuum of lithospheric relaxation times ranging from 0.1-200 million years for the lower regions of the lithosphere to time scales larger than the age of the earth for the upper regions. As stress is relaxed by the mechanism of creep (Slingerland et al. 1989) the lithosphere responds by evolving through a series of quasi-elastic states. Upon initial loading of the lithosphere by thrust sheets a flexural moat is formed adjacent to the orogenic front and a peripheral bulge migrates cratonward ahead of the thrusts, creating an unconformity as it progresses. As the

lithosphere relaxes stress, the peripheral bulge migrates towards the load, shortening and deepening the basin (Fig. 52a). The Quinlan and Beaumont model adequately explains the unconformities and stratigraphy above the intercratonic basin arches of the Paleozoic Appalachian foreland basin through the use of carefully chosen flexural parameters. However, this model did not consider the effects of non-equilibrium sedimentation and subsidence rates during the formation and filling of the Late Silurian-Early Devonian and Late Devonian-Early Mississippian deep basins, nor erosion and redeposition of thrust sheet mass.

Flemings and Jordan (1990) considered these variables. Their model is one in which the lithosphere, loaded by thrust sheets, responds elastically, creating a downwarping adjacent to the load. In contrast to the visco-elastic model, the peripheral bulge initially migrates toward the load and, as thrust sheets erode and sediment is redeposited within the basin, the redistributed mass drives the peripheral bulge away from the load, towards the craton (Fig. 52b). As thrusting episodes continue, the elastic model predicts that the basin axis will shift progressively cratonward.

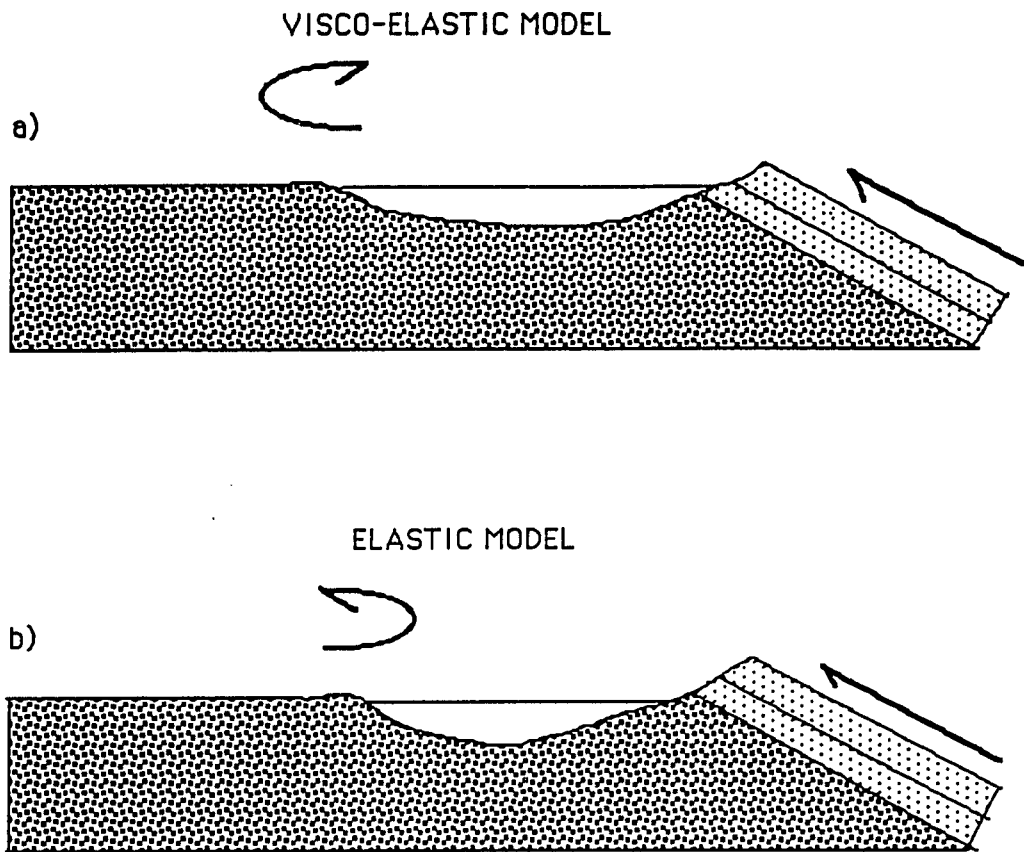


Fig. 52. Motion of the peripheral bulge during thrusting predicted by the Visco-elastic and Elastic Models of lithospheric flexure. (After Flemings and Jordan, 1990)

The viscoelastic and elastic models predict different facies architectures and stratal patterns. Flemings and Jordan (1990) note that in the viscoelastic model erosional unconformities associated with the peripheral bulge are cut during quiescence and a gradual retrogradation of facies occurs during this time. The reverse is predicted for the lithosphere in the elastic model. Unconformities are cut and retrogradation of facies occurs at the start of loading events and facies prograde during thrusting and quiescence.

Whether the lithosphere can be modelled visco-elastically or elastically over basin forming time scales is largely a function of the flexural parameters chosen. Temperature gradients within the lithosphere, which control the flexural rigidity of the lithosphere, are often invoked to determine visco-elastic relaxation time. However, not all lithospheric weakening is temperature related. The Appalachian foreland basin is a product of a complex thermal and mechanical history (see previous discussion of Appalachian foreland basin history). The lithosphere underlying the Appalachian foreland basin was subjected to at least one episode of collisional tectonics (the Grenville orogeny), subsequent rifting, and development of a passive margin. Since there is some suggestion that the flexural

rigidity of the lithosphere decreases as the duration of loading increases, and since Bouguer gravity anomalies (Karner and Watts 1983) and models proposed by Slingerland et al. (1989) suggest that a significant fraction of thrust sheet mass remains outboard of the basin, this accumulating mass may have weakened the lithosphere during the secular history of the Appalachian foreland basin. Hauser, (1993) reports that Grenville age thrusts and perhaps a remnant foreland basin of Grenvillian age remain under the United States midcontinent. There are also suggestions that the lithosphere may be weakened by the decoupling of elastic and visco-elastic regions, allowing each to flex independently. Other variables that could affect peripheral bulge migration and complicate stratigraphic architecture include inherited anisotropies in lithospheric strength and structure that may restrict the motion of the peripheral bulge (Waschbusch and Royden, 1992) and intraplate stresses proposed by Cloetingh (1988) to explain relative sea level fluctuations of up to 100 meters in amplitude. Cloetingh related these changes in sea-level to changes in regional stress regimes. In his model, forces generated by ridge push and slab pull are transmitted for long distances through lithospheric plates. Compression produces uplift

at basin margins and basin axis subsidence. Conversely, tensional forces produce subsidence in marginal areas and uplift of the basin axis. Intraplate stresses of a few kilobars have been documented in a study of earthquake focal mechanisms in the Bay of Bengal. Major changes in the regional stress regime of this area appear to have taken place 5-6 million years ago. These changes may have affected the Australian passive margin in that the Otway Bay area of southern Australia has experienced a regressive event despite the rise in global sea-level (Allen and Allen 1990, p.161). Lesser regional stress fields have been documented along the Java-Sumatra trench and within the Nazca plate.

Since crustal shortening played a major role in the development of the Appalachian foreland basin during the Acadian orogeny, horizontal forces within the lithosphere may have been responsible for a measure of the observed stratigraphy. Whatever the fractional contribution of each of the complex interdependent array of variables responsible for the observed cyclicity it is significant that parasequences, system tracts, and sequences are generated both in passive margin and foreland basin settings. Clastic parasequences display the same coarsening upward,

shallowing upward trends in both of these tectonic settings. This suggests that there is some mechanism of formation that transcends the characteristic tectonic conditions of foreland basin or passive margin settings.

EUSTASY

Global sea-level changes result from either a change in the amount of water in the ocean basins or in the volume of the ocean basins themselves. Allen and Allen (1990) suggest four possible causes of global sea-level change. These are (1) continuing differentiation of lithospheric material as a result of plate tectonic processes, (2) changes in the volumetric capacity of the ocean basins as a result of sediment accumulation or abstraction, (3) changes in the volumetric capacity of the ocean basins caused by expansion or contraction of the mid-ocean ridge system, and (4) reduction of available water as a result of ice cap formation or continental glaciation.

Continuing differentiation of lithospheric material may result in more juvenile water being added to the hydrosphere by volcanic

activity. However, this process is roughly balanced by the subtraction of water at subduction zones and hydrothermal alteration of material at spreading centers. Since there has been no appreciable change in the chemistry of sea water in the last 1.5 billion years (Allen and Allen, 1990 p.163), this suggests that no appreciable amount of water has been added to the oceans during that time. Other variables that may produce truly eustatic changes in sea-level are changes in basin volume caused by sediment accumulation or by changes in mid-ocean ridge volume. Sediment is added to ocean basins at varying rates and settles on the sea floor. Although the weight of the accumulating sediments will depress the sea floor slightly, there will be a net rise in sea-level as water is displaced by the volume of sediment added. This process is balanced by the subduction of sediments at trenches and the addition of material to continental margins in accretionary wedges. Given the short life of the oceanic lithosphere the amount of sediment that may be accumulated during its lifetime is not thought capable of producing the observed changes in sea-level. Changes in mid-ocean ridge volume and its effect on sea level have been noted by many workers, including Johnson (1971), Hayes and Pitman (1973), and

Pitman (1978, 1979). The present ridge system is approximately 10% of the volume of ocean water. Since ocean crust cools and sinks as a function of the square root of age, changes in spreading rates would increase or decrease the volume of the ridge system causing eustatic rises and falls in sea-level. Changes in the total length of spreading axes would also cause eustatic changes in sea-level.

Rapid glaciation and deglaciation has had a significant effect on sea-level in the recent past. Sea level curves based on oxygen isotopes and benthic foraminifera show that sea-level has undergone eight high frequency fluctuations in the past 120,000 years ranging from 20-180 meters in height. These fluctuations have a primary period of 10^5 years and shorter secondary periods of 40,000 and 20,000 years. This periodicity is what is expected from Milankovitch orbital forcing parameters. Based on the paleogeography of the Middle Devonian, glacio-eustasy was probably not a factor during that time (Johnson et al. 1985), and so the most reasonable cause of eustatic changes during deposition of the Hamilton Group would be fluctuations in spreading rates and resulting ridge volume changes, and/or changes in the lengths of ridge spreading axes.

If a eustatic component is found, it may not be synchronous in widely separated basins with differing rates of subsidence. Parkinson and Summerhayes (1985) use an example of different basins undergoing subsidence at different rates during a sea-level fall. If the rate of sea-level fall is greater than the subsidence rate, and an adequate supply of sediment is available, facies will prograde in each of the basins simultaneously. If at some later time the rate of subsidence in one basin exceeds the rate of sea-level fall, then that basin would experience a transgression while the other basins with smaller subsidence rates continued to produce progradational sediment packages. Although this is the case for eustatic sea-level falls, eustatic sea-level rises would be recorded in each of the hypothetical basins as a transgression.

ORIGIN OF THE TYPE 2 SEQUENCE BOUNDARY

Brett and Baird's (1985) tracing of carbonate units into the tops of coarsening upward clastic units across the basin axis, and their observation that the erosional surface under the Tichenor Limestone represents the largest magnitude fall in sea-level during

Givetian time may allow an approximate determination of tectonic versus eustatic control on the generation of the Type 2 sequence boundary. Brett and Baird (1990) indicated that an erosional surface underlying the Tichenor Limestone is evident both east and west of the basin axis, consistent with their interpretation of this event as caused by a drop in relative sea-level. Along the eastern rim of the basin, progradational coarsening upward units reach their maximum basinward extent during the late Middle Devonian in the Montebello Sandstone and its equivalents, the Clearville Siltstone to the south, and the informal "Middle" member of the Mahantango Formation in north-eastern Pennsylvania and adjacent New York. The top of the Montebello Sandstone and its equivalents along the eastern rim of the basin is marked by a lag of regional extent and a shift to retrogradational parasequence sets above the Type 2 sequence boundary. In distal areas to the south and west, the approximate stratigraphic equivalent of the Montebello Sandstone, the Clearville Siltstone, is capped by a nodular limestone, the Pokejoy member. The Pokejoy member is analogous to the Tichenor Limestone both in terms of their common mechanism of formation and their inferred stratigraphic position immediately above a Type 2 sequence

boundary. This suggests that the Type 2 sequence boundary along the eastern and northwestern margins of the Appalachian foreland basin was caused by a sea-level fall of at least regional extent. But because peripheral bulge migration can cause unconformities in foreland basin settings, the key to determining if the Type 2 boundary was created by regional sea-level fall or global eustasy is the stratigraphy away from the effects of peripheral bulge migration. Lash (1987), suggested that along strike variations in stratigraphy could be used to discriminate between unconformities caused by eustatic drops in sea level and those caused by peripheral bulge migration. Taking that tack, strata of similar age both south and west of the study area show unconformities of similar age as the Type 2 boundary. West of the study area in Ashtabula County, Ohio, the Tichenor Limestone sits unconformably on the Wanakah Shale (Woodrow et al. 1988, their figure 6). Further west, cratonward of the Cincinnati Arch in the southern Illinois basin, the upper Givetian contact between the Lingle and Alto Formations is marked by a phosphatic lag deposit. South of the study area in Lincoln County, West Virginia, an unconformity is mapped at the top of the Clearville Siltstone (Woodrow et al. 1988, their figure 7). The

top of the Clearville Siltstone thus marks a maximum regression and probably defines the Type 2 boundary in southern Pennsylvania, Maryland, Virginia, and West Virginia. Well south of the study area, Millici and deWitt (1988) state that the lateral equivalent of the Mahantango Formation, the Millboro Shale, is truncated by a late Middle Devonian unconformity in south-west Virginia and eastern Tennessee. Sea-level fall of upper Givetian age has been documented by Johnson et al. (1985) in the western United States, Canada, Belgium, and Germany (Johnson et al. 1985, their Fig.2). Since this fall in sea-level occurred penecontemporaneously both within and outside of the Appalachian foreland basin, it is suggested that the Type 2 boundary separating the Highstand and Transgressive system tracts in the Appalachian foreland basin during Givetian time was formed by a eustatic fall in sea-level. It should be noted that the unconformity marking the Type 2 boundary does not underlie the entire basin. Deep basinal areas experienced continuous deposition throughout Givetian time, although detailed studies may show that the Type 2 boundary in those areas is marked by a submarine erosion surface, as it is in the basinal areas of western New York. If rates of subsidence were greater proximal to the load

then the timing of the relative sea-level fall and Type 2 sequence boundary generation on the eastern rim of the basin would have been slightly younger than the Type 2 sequence boundary along the western margin. The data available to date indicate that this is not the case.

EVIDENCE FOR PERIPHERAL BULGE MIGRATION

As stated previously, if the visco-elastic relaxation time is long enough, the migration of the peripheral bulge will play a major role in producing the patterns of stratigraphy and unconformities observed within the Appalachian foreland basin during Middle Devonian time. In this model, erosional unconformities caused by migration of the peripheral bulge are cut during tectonic quiescence and retrogradation of facies proximal to the thrusts occurs gradually (Flemings and Jordan 1990). If the visco-elastic relaxation time is short, as in the elastic model, the distal unconformity is cut and facies proximal to the load retrograde rapidly at the start of a thrust cycle. Based on the results of this study, and the work of Brett and Baird along the northwestern rim of the basin, a tentative test of the rheological response of the

lithosphere and the contributions of tectonic versus eustatic controls on stratigraphic patterns may be undertaken. However, certain assumptions must be made. Given the current state of knowledge regarding the biostratigraphical zonation of the upper Middle Devonian along the eastern margin of the Appalachian foreland basin, it is not possible to offer independent biostratigraphical evidence as to the depositional synchronicity of units along the eastern and north-western rims of the basin. Similar age of these units may be inferred by the choice of measured sections parallel to strike on the eastern margin and by the stratigraphic position of the Hamilton Group below a widespread marker unit, the Tully Formation and its equivalents. The peripheral bulge is assumed to remain at constant width, and consequently unconformities are assumed to be due to motion, rather than width changes. Also, the relative rapidity of facies retrogradation or progradation is inferred from the magnitude of paleobathymetric change of facies superposed above and below flooding surfaces of parasequences.

Because erosion has removed the cross-cutting relationships required to independently constrain the timing of thrust sheet

motions, the details of stratigraphy along the eastern and north-western rims of the basin may provide evidence for lithospheric rheology. Differentiation between tectonic and eustatic influences must be obtained from cyclicity. Figure 51 shows that the Windom Shale lies disconformably on the top of the Tichenor limestone near Buffalo, New York. However, further east into the basin, the Tichenor Limestone is overlain conformably by the Menteth Limestone and the Kashong Shale members of the Moscow Formation. Minor hiatuses intervene between the Tichenor Limestone and the Menteth Limestone and the Menteth Limestone and the Kashong Shale (Brett and Baird 1990, their Fig.3). These members have been interpreted as a Transgressive system tract by Brett and Baird. The Menteth and Kashong Members of the Moscow Formation occupy a stratigraphic position below the Tully Formation and its equivalents, and above the underlying Type 2 boundary. This is a direct correspondence to the Transgressive system tract represented in the study area by the Sherman Ridge Member and the Harrell/Burket Shales in sandstone-dominated areas and the unnamed upper shale member in mudstone-dominated areas. The Windom Shale member of the Moscow Formation in western New York has been interpreted by Brett and

Baird (1990) as early Highstand deposits truncated by erosion and resting in angular unconformity below the Tully Formation. In the mudstone-dominated areas of north-eastern Pennsylvania and adjacent New York, the upper shale member of the Mahantango Formation also begins to grade upward from shale to distinct coarsening upward cycles, consistent with the development of the equivalent of an early Highstand system tract (Slattery et al. in prep.). In this area, as similar to the north-western rim of the basin, the Tully equivalent Sparrow Bush Formation rests on an erosional surface beveled into the uppermost Mahantango Formation. It is evident from the relationships between the study area on the eastern rim of the basin and the north-western margin, that each system tract in the study area has its counterpart along the north-western rim.

Many previous studies (Jacobi 1981, Quinlan and Beaumont 1984, Etensohn 1981, among others) have attempted to infer peripheral bulge migration by noting stratigraphic patterns and unconformities. Etensohn (1981) in particular used patterns of unconformities to infer peripheral bulge motion within Mississippian age strata in Kentucky. The pattern of unconformities

observed along the northwestern margin of the basin are shown in Fig. 53. The unconformity above the Tichenor Limestone merges into younger unconformities between the Menteth and Kashong members of the Moscow Formation, and the Kashong and Windom Shale members of the Moscow Formation until all are merged in the west into the angular unconformity developed below the Tully Formation. The pattern of unconformities is consistent with uplift and "oscillatory" motion of the peripheral bulge, with the principal motion of the peripheral bulge being towards the craton. Given the contrasting lithospheric responses in the visco-elastic and elastic models, this pattern of unconformities is best explained by an elastic response to loading. In this model, Acadian thrusting loaded the lithosphere causing the peripheral bulge to migrate toward the load. As sediment from the eroding thrust sheets was flushed into the basin, further loading the lithosphere, the peripheral bulge migrated towards the craton. Subsequent episodes of thrusting and erosion had the effect of driving the basin axis slightly westward over time producing the observed pattern of unconformities during Givetian time. This is the general pattern of unconformities throughout Middle Devonian time on the western flank of the

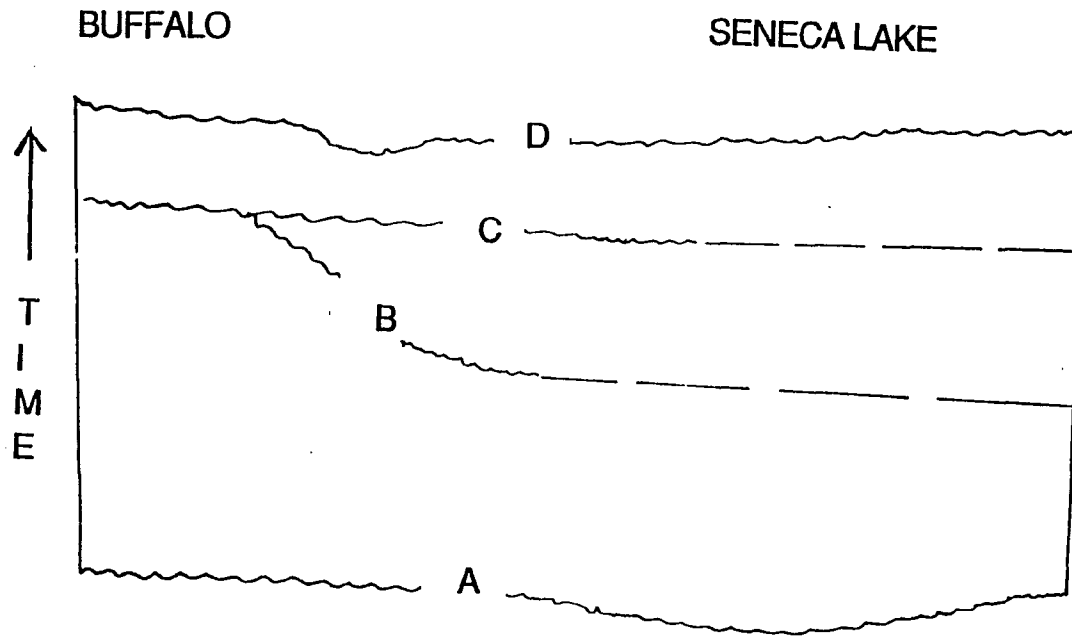


Fig. 53. Merging unconformities of the Moscow Formation on the western rim of the Appalachian foreland basin during the Middle Devonian. A= sub-Tichenor unconformity, B=sub-Kashong unconformity, C=sub-Windom unconformity, D= unconformity at top of Windom Shale. Dashed lines are conformable contacts.

Appalachian foreland basin. Unconformities "vee" to the west, omitting more time in cratonward sections. Millici and deWitt (1988) state that this erosional hiatus separates progressively younger rocks from older Paleozoic strata west and south of the study area. For example, in north-central Ohio, Upper Devonian rocks sit disconformably on lower Middle Devonian strata, in southern Kentucky late Upper Devonian black shales overlie Upper Ordovician deposits, and in eastern Tennessee the Upper Devonian Chattanooga Shale rests disconformably on Middle Ordovician black shales. If one assumes a constant peripheral bulge height and width then the motion of the peripheral bulge is consistent with a short visco-elastic relaxation time, and the rheology of the lithosphere underlying the Appalachian foreland basin during Middle Devonian time may be modeled as elastic. This interpretation is preliminary, and further stratigraphic testing is needed to bring additional evidence to bear on this question. Clearly, the same pattern of merging unconformities could have been produced by eustatic or relative changes in sea-level.

A DEPOSITIONAL MODEL

It should be stressed that an elastic lithospheric response to loading by thrust sheets of the Acadian orogeny has only been inferred during the Middle Devonian. This may or may not have been the response of the lithosphere during the entire history of the basin. However, given that the lithospheric response was elastic in the Middle Devonian, the following sequence of events may be inferred. A major thrusting event occurred at the beginning of Cazenovian time with the deposition of the Marcellus Formation. Correlative with basin subsidence and Marcellus deposition in the east, an unconformity was cut in areas to the west of the study area (i.e. eastern Ohio) between the Delaware Limestone and the Marcellus Formation (Fig.12, this report). This unconformity "vees" to the west, marking the motion of the forebulge as it moved first towards the thrusts and then away from it as erosion and deposition of Marcellus clastics further loaded the lithosphere. Continued thrusting increased relief in the source area producing a net greater volume of clastics available to the basin. Quiescence and erosion due to isostatic rebound of the source area followed, leading to

progradation of facies into the basin and reflected in the prograding parasequence sets of the Highstand system tract. A second smaller scale thrusting event may have led to the deposition of the thick Montebello Sandstone Member. Eustatic sea-level fall is suggested as the mechanism that produced the Type 2 sequence boundary, for reasons outlined previously. Subsequent sea-level rise then produced the carbonates of the Tichenor Limestone in the west and the Pokejoy member in the south-east. The Tichenor Limestone is also bounded by a disconformity above (Fig.53). Another unconformity exists between the overlying Menteth Limestone and Kashong members, and the Kashong and Windom Shale members of the Moscow Formation. The motion of the forebulge may be inferred from the stratigraphy. If the greatest net movement of the peripheral bulge was towards the load at this time one would expect coarser forebulge derived clastics, not the fine grained Transgressive and early Highstand deposits of the Moscow Formation. The unconformities reflect uplift and motion of the peripheral bulge first towards, then away from the basin. The elastic model provides a "best fit" for field data on both sides of the Appalachian foreland basin. Renewed thrusting and basin subsidence is recorded by the

deposition of the condensed section portion of the Transgressive system tract along the eastern rim of the basin. Lacking independent biostratigraphic evidence, the condensed section along the eastern rim of the basin is tentatively placed within the Harrell/Burket shales and the Upper Shale Member of the Mahantango Formation. Concurrent with deposition in the east, the disconformity above the Tichenor Limestone and the unconformity between the Menteth Limestone and Kashong Shale marks the passage and retreat of the peripheral bulge, followed by the deposition of Transgressive and early Highstand system tract deposits of the Moscow Formation in the west (Brett and Baird 1990). This pattern of onlapping strata punctuated by unconformities is consistent with uplift and oscillation of the peripheral bulge. The observed pattern could also have been produced by relative or eustatic sea-level changes. However, lacking any evidence for glaciation in the Middle Devonian it is suggested that the mechanism producing the minor unconformities between the members of the Moscow Formation along the north-western rim of the Appalachian foreland basin during upper Middle Devonian time is tectonics. By the same line of reasoning, the fourth order cyclicity throughout the basin is

probably controlled by tectonics, while the mechanism controlling third order cyclicity in the study area during the upper Middle Devonian was eustasy. Therefore, the stratigraphy of the Appalachian foreland basin during Middle Devonian time is probably a product of the interplay of tectonics and eustasy. These two mechanisms may be found to be end members of the same process. Perhaps interplate stresses and changing ridge volumes account for the observed cyclicity. This is suggested because clastic parasequences in diverse tectonic settings are all characterized by the same shallowing upward, coarsening upward trends, terminated by marine flooding surfaces. This begs the question of a common mechanism of formation. Perhaps a component of intraplate compression is responsible for the observed cyclicity. In foreland basin settings, crustal shortening provides the compression and on passive margins, perhaps the driving mechanism is ridge push. Further research may show that the ultimate mechanism for stratigraphic cyclicity at all scales are driven by processes deep within the earth.

CONCLUSIONS

The Mahantango Formation is composed of shallowing upward, coarsening upward cycles bounded by marine flooding surfaces with cycle thicknesses ranging from several to many tens of meters. In sequence stratigraphic terms they are defined as parasequences. Parasequence stacking patterns in sandstone-dominated and mudstone-dominated sections of the study area are identical. Parasequences are progradational up to a regional lag deposit datum, and subsequently retrograde, defining a Highstand system tract, Type 2 sequence boundary, and a Transgressive system tract. Similar age strata along the northwestern rim of the basin display the same sequence stratigraphic stacking pattern. Sedimentation rates varied along strike and across the basin, and although higher sedimentation rates produced more complex facies suites in sandstone-dominated areas, the parasequence stacking pattern of progradation and retrogradation was unaffected. Limestones along the northwestern rim of the basin have previously been correlated with the tops of coarsening upward cycles, and this study extends this concept, linking the erosional surface below the Tichenor Limestone in the west and to the marine flooding surface above the Montebello

Sandstone and equivalents in the east as a Type 2 sequence boundary. This marks the maximum regression of a Vail third order cycle. This Type 2 boundary was probably formed by a eustatic drop in sea-level, inasmuch as sea-level falls of similar age have been noted in areas not thought to be linked to the Acadian Orogeny. This eustatic event was probably caused by changes in ridge spreading volume or plate reorganization during the Middle Devonian. The process responsible for the generation of Vail fourth order cycles (parasequences) is not clear, but given the lack of a suitable glacial mechanism, intra-plate stresses, including thrusting seems the most reasonable driving mechanism. The pattern of unconformities along the western rim of the basin suggests that motion of the peripheral bulge is consistent with an elastic lithospheric response to loading. The results of this study may be used to simplify the confusing lithostratigraphic assemblage of Mahantango Formation Members into a genetically linked depositional system. It therefore allows study of time equivalent facies within the Appalachian foreland basin during the Middle Devonian to be studied free of restrictions imposed by lithostratigraphy.

APPENDIX 1

LOCATIONS OF MEASURED SECTIONS

A. Port Jervis, New York. Roadcuts and hillside exposures along Skyline Drive, starting 0.1 miles after its intersection with U.S. Route 97. Port Jervis North Quad.

B. Sawkill Creek, Milford, Pennsylvania. Stream exposures starting at the lowest exposures in the stream bed of the Sawkill 0.2 miles north-west of the point where the Milford-Childs Park road crosses the stream. Milford Quad.

C. Raymondskill Creek, Milford, Pennsylvania. Stream exposures starting at the lowest exposures in the stream bed of Raymondskill Creek 0.25 miles west of the point where U.S. Route 209 crosses the stream. Milford Quad.

D. Bowmanstown, Pennsylvania. Roadcut along east side of entrance ramp to Pennsylvania Route 248. Palmerton Quad.

E. Rockville, Pennsylvania. Roadcut above power station on Roberts Valley Road, 0.2 miles east of its intersection with Front Street. Harrisburg West Quad.

F. Watts, Pennsylvania. Roadcut on U.S. Route 22-322 1.1 miles west of the Watts exit. Amity Hall Quad.

G. Girty's Notch, Pennsylvania. Roadcut at picnic area on U.S. Route 11-15, 1.5 miles southwest of Montgomery Ferry. Halifax Quad.

OTHER REFERENCE SECTIONS

Dalmatia Quarry, 2.5 miles south of Dalmatia, Pennsylvania along the east side of the Penn Central Railroad near milepost 121. The section is in the third of four quarries, counting from the south. Photographs.

Borrow Pits, 1.6 miles south of Milford, Pennsylvania along U.S. Route 209. Photographs.

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