INTRODUCTION

This dissertation is comprised of three chapters that independently present and discuss the sedimentological/ stratigraphic features within Upper Mississippian strata of southern West Virginia.

Chapter 1 presents an interpretation of the depositional environments and the high-resolution sequence stratigraphic history of the Hinton, Princeton, and Bluestone Formations. These three formations collectively comprise the bulk of the Mauch Chunk Group, a thick (up to 100 meters) series of strata that filled the central Appalachian basin during the Late Mississippian. High-frequency (~400 k.y.) and low-frequency (2-4 million year) sequences identified within the study interval are discussed with regard to glacio-eustatic and tectono-eustatic mechanisms as possible controls on sequence development.

Chapter 2 presents a detailed analysis of the hierarchy of very high-frequency (semi-diurnal to decadal scale) cycles that are recorded in the finely-laminated Pride Shale Member of the Bluestone Formation. The Pride Shale is interpreted as a prodeltaic highstand succession internal to one of the fourth-order (~400 k.y.) sequences that comprise the study interval. The very high-frequency cycles are discussed in regard to periodic tidal and climatic controls on sedimentation.

Chapter 3 discusses evidence for fluctuating paleoclimatic conditions that accompanied fourth-order (~400 k.y.) sequence development in the Appalachian Basin. Linked eustatic and paleoclimatic change is discussed in regard to the effect of glacial-interglacial cycles upon the relative strength and latitudinal range of global-scale monsoonal circulation. The Upper Mississippian record of seasonal conditions is compared with the Lower Pennsylvanian record that preserves evidence of everwet climatic conditions.

Chapter 1:
Sequence Stratigraphy of Upper Mississippian Strata in the Central Appalachians: A Record of Glacioeustasy and Tectonoeustasy in a Foreland Basin Setting

ABSTRACT

The Upper Mississippian Hinton, Princeton, and Bluestone formations of southern West Virginia constitute a westwardly thinning wedge of strata that filled the central Appalachian basin over a ~7 million year time interval. Up to 17 unconformity bounded, transgressive-regressive sequences comprise the study interval in the basin depocenter. Five sequence types are recognized based upon the degree of basal incision, the overall thickness, and the character of the dominant facies. These are: 1) major incised valley-fill to coastal plain, 2) major incised valley-fill to deltaic, 3) minor incised valley-fill, 4) coastal plain, and 5) marine-dominated sequences. Sequence development is ascribed to fourth-order (~400 k.y.) glacioeustatic cycles that reflect climatic change associated with variations in orbital eccentricity during the early stages of the Permo-Carboniferous Gondwanan glaciation. The cyclic character of the Upper Mississippian stratigraphic record is comparable to Pennsylvanian cyclothems in the Appalachian and Interior basins.

The fourth-order sequences stack into two third-order (2-4 million year) composite sequences that are bounded by regional unconformities. Each composite sequence consists of: 1) a basal retrogradational interval comprised of a major paleovalley-fill sequence overlain by a thick aggradational sequence set of fluvial/ coastal plain sequences that together constitute the transgressive systems tract; 2) a marine interval that demarcates maximum flooding; and 3)(where preserved) a progradational sequence set consisting of minor incised valley-fill sequences that comprise the highstand systems tract. The third-order sequences document long-term changes in accommodation that are ascribed to tectonically driven eustasy.

INTRODUCTION

The origin of cyclothemic Carboniferous (Mississippian-Pennsylvanian) strata in North America is controversial because of uncertainties in discriminating the effects of eustasy, climate, and tectonics on sequence development (Wanless and Shepard, 1936; Veevers and Powell, 1987; Klein and Willard, 1989; Cecil, 1990; Klein and Kupperman, 1992). Most workers favor a glacioeustatic mechanism to explain the repetitive record of fourth-order (~400 k.y.) sequences in Pennsylvanian rocks of the midcontinent (Crowell, 1978; Heckel, 1986; Boardman and Heckel, 1989; Archer et al., 1994). In the Appalachians, recognition of a periodic eustatic signal that may have influenced sequence development is more difficult because the proximal section is largely non-marine (Busch and Rollins, 1984). Whereas cyclothem-scale sequences are recognized in Pennsylvanian strata of Kentucky (Aitken and Flint, 1995), comparatively few studies have documented high-frequency sequences within Mississippian rocks of the foreland basin depocenter.

This study of Upper Mississippian strata in southern West Virginia focuses on predominantly terrestrial, siliciclastic units that filled the central Appalachian basin during the early stages of Alleghenian orogenesis (Englund and Thomas, 1990). The study interval is characterized by multiple depositional packages that correlate in outcrop and in the subsurface throughout southern West Virginia. Newly discovered tidal rhythmites within many of these packages provided a basis for interpreting transgressive episodes of deposition that followed valley-incision into terrestrial facies. We interpret the depositional packages within a sequence stratigraphic framework that relates their development to short term (~400 k.y.) fluctuations in

base level over approximately 7 million years. The periodicity and likely forcing mechanisms of cyclic base-level change is discussed with respect to the tectonic setting and the Milankovitch-band climatic change that forced coeval episodes of Gondwanan glaciation. Lastly, we discuss possible mechanisms for long-term accommodation change that controlled the preservation of individual sequences within two retrogradational-aggradational to progradational composite sequences.

PALEOGEOGRAPHIC AND STRATIGRAPHIC SETTING

This study focuses on the Upper Mississippian Hinton, Princeton, and Bluestone formations of the Mauch Chunk Group in southern West Virginia (Figs. 1.1, 1.2). These formations outcrop along the southeastern margin of the Allegheny Plateau, and dip 1- to 2-degrees west into the subsurface where they can be traced in well-logs throughout an
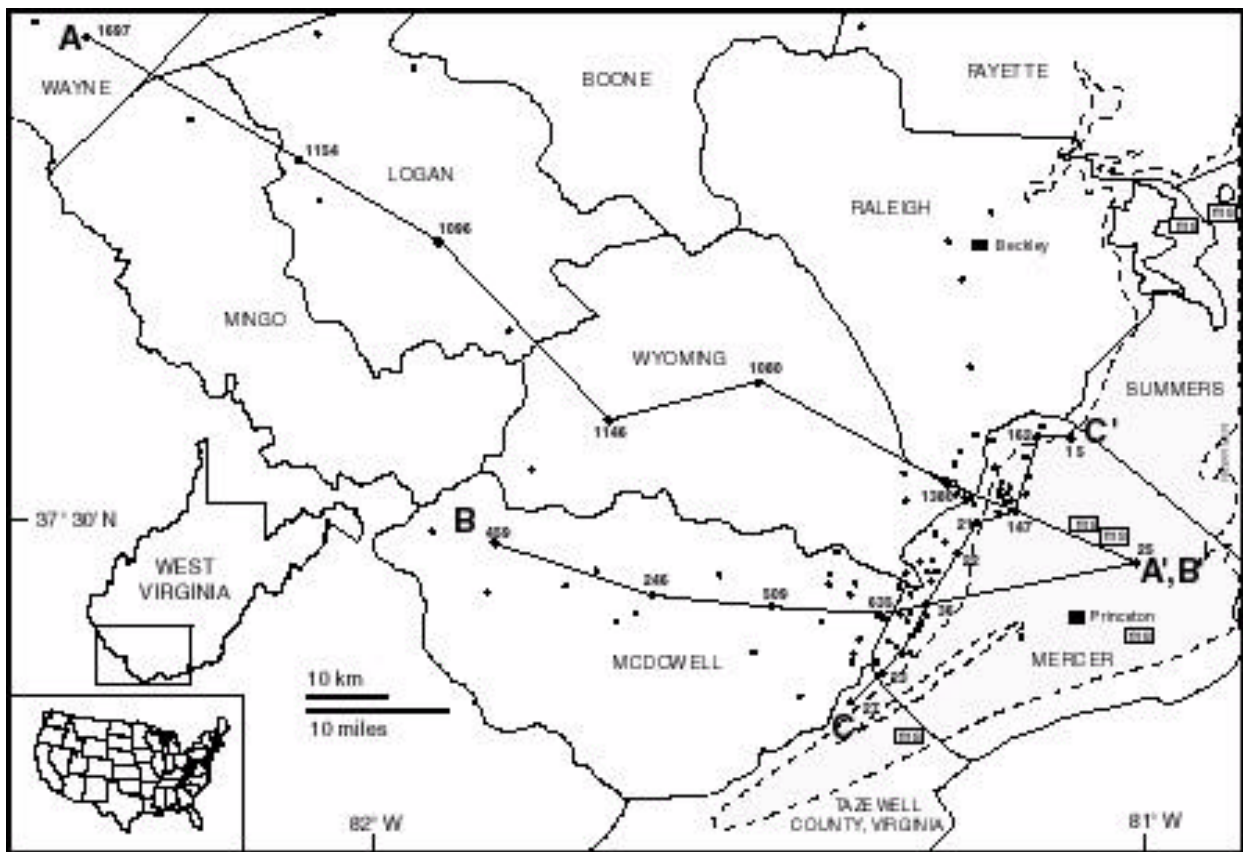

Figure 1.1- Map of the study area in southern West Virginia showing the outcrop belt of Upper Mississippian strata (shaded) and the locations of measured sections (ms) and well-logs used in this study.
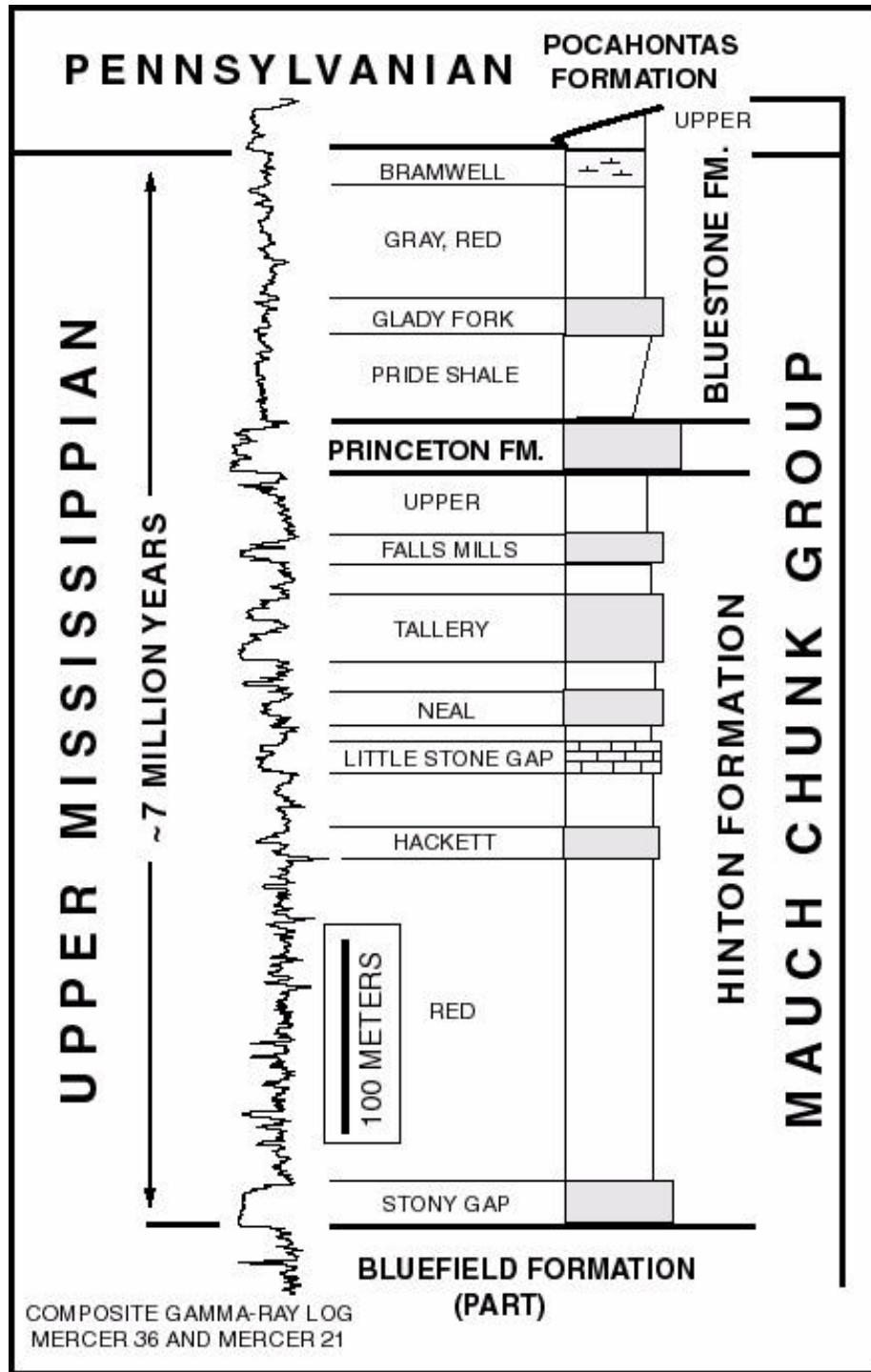
Figure 1.2- Stratigraphic column of the upper part of the Mauch Chunk Group in southern West Virginia showing typical gamma ray log signature from the west central part of Mercer County. General age constraints based on Manger and Sutherland (1984) and Jones (1996).

approximately 5000 km^2 area in the southernmost West Virginia counties. In adjacent parts of southwestern Virginia and Kentucky, the Upper Mississippian units that comprise the study interval are known as the Pennington Formation (Englund and Henry, 1984).

Terrigenous units in the Mauch Chunk Group, including those in the basal Bluefield Formation, are part of a thick (up to 1000 m) clastic wedge that was sourced by tectonic highlands along the eastern margin of the Appalachian basin (Englund and Thomas, 1990). Basin infilling was coincident with northward drift of the study area through sub-equatorial latitudes (15 to 5 degrees south) during the collisional assembly of Pangea (Scotese, 1986; DiVenere and Opdyke, 1991; Golonka et al., 1994). Deposition also coincided with the early phases of the Permo-Carboniferous Gondwanan glaciation (Crowell, 1978; Veevers and Powell, 1987; Gastaldo et al., 1996). The thick section in the eastern outcrop belt represents one of the most complete records of Upper Mississippian sedimentation in North America (Englund and Randall, 1981; Englund and Thomas, 1990); the section thins dramatically westward into the distal parts of the basin (Fig. 1.3).
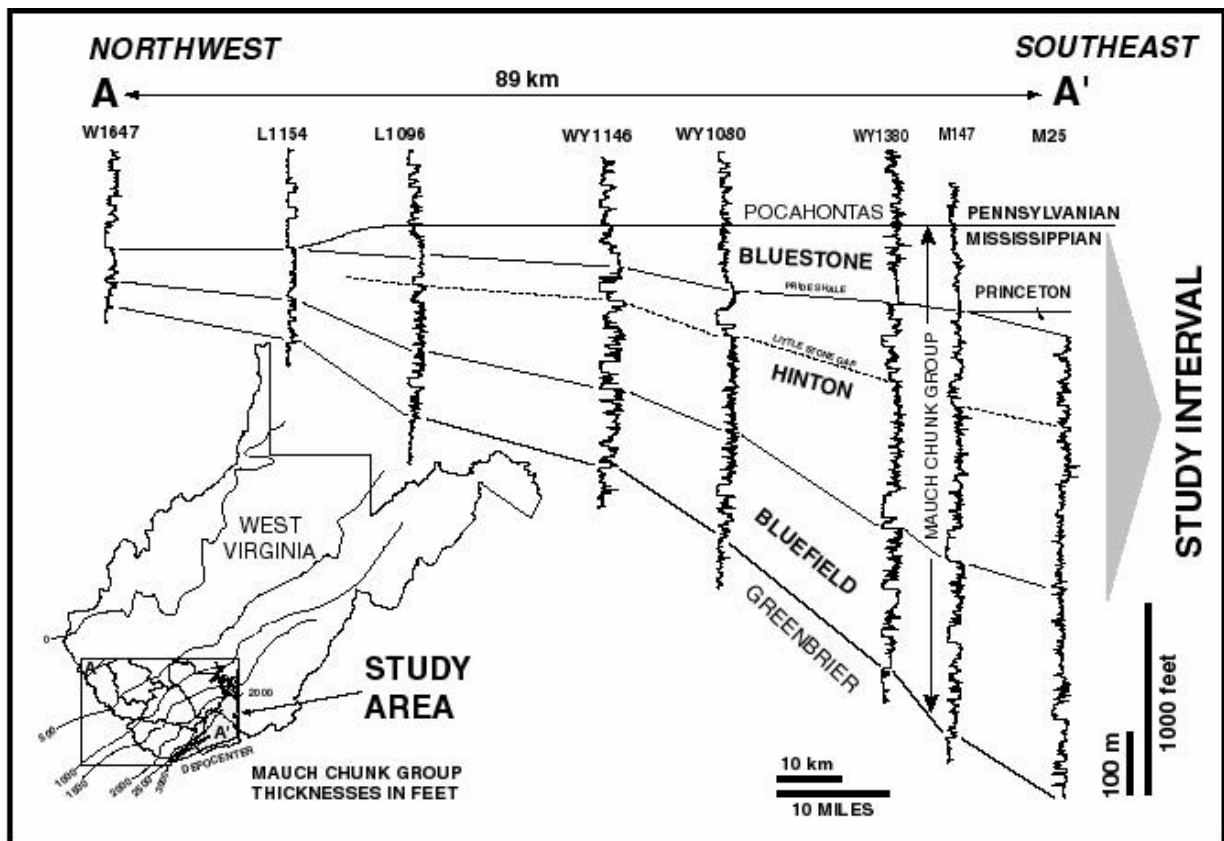

Figure 1.3- Regional cross-section showing northwestward thinning of the Mauch Chunk strata across southern West Virginia. Upper datum is the Mississippian-Pennsylvanian boundary. The Little Stone Gap Limestone and the Pride Shale are distinctive stratigraphic markers in the eastern part of the central Appalachian Basin. Isopach map for the Mauch Chunk interval adapted from Arkle (1974). See Figure 1.1 for location of cross section.

The study interval is dominated by red, terrestrial mudstone, but includes several thick sandstone units (the Upper Maxon and Ravencliff Sandstones of driller's terminology) that are the principal Upper Mississippian gas-producing zones in Wyoming and McDowell counties (Wrightstone, 1984; 1985). Barlow (1996) estimates that 1.4 TCF of gas was originally in place within these and other Mauch Chunk sandstones in the subsurface of West Virginia, Virginia, and Kentucky. Also within the interval are marine units (e.g. Little Stone Gap Limestone Member of the Hinton Formation, and the Pride Shale Member of the Bluestone Formation) that are distinctive, regional stratigraphic markers (Fig. 1.3). The Mississippian-Pennsylvanian

boundary at the top of the study interval is defined in outcrop where the fossiliferous Bramwell Member (Bluestone Formation) is overlain by either the heterolithic Upper Member or a quartzose sandstone (Pocahontas Formation) that contain the plant megafossil *Neuropteris Pocahontas* (Englund et al. 1981; Englund and Thomas, 1990; Hoare, 1993). Plant fossils from the Pocahontas Formation are of Early Pennsylvanian age (Pfefferkorn and Gillespie, 1981).

Brachiopod biostratigraphic data suggest that marine units in the Bluefield and Hinton formations correlate with the Glen Dean, Menard, and Clore Limestones of the midcontinental Chesterian type section (Fig. 1.4)(Gordon et al., 1983; Henry and Gordon, 1990). Ammonoid and conodont biostratigraphic data from the Bramwell Member (Bluestone Formation) suggest a
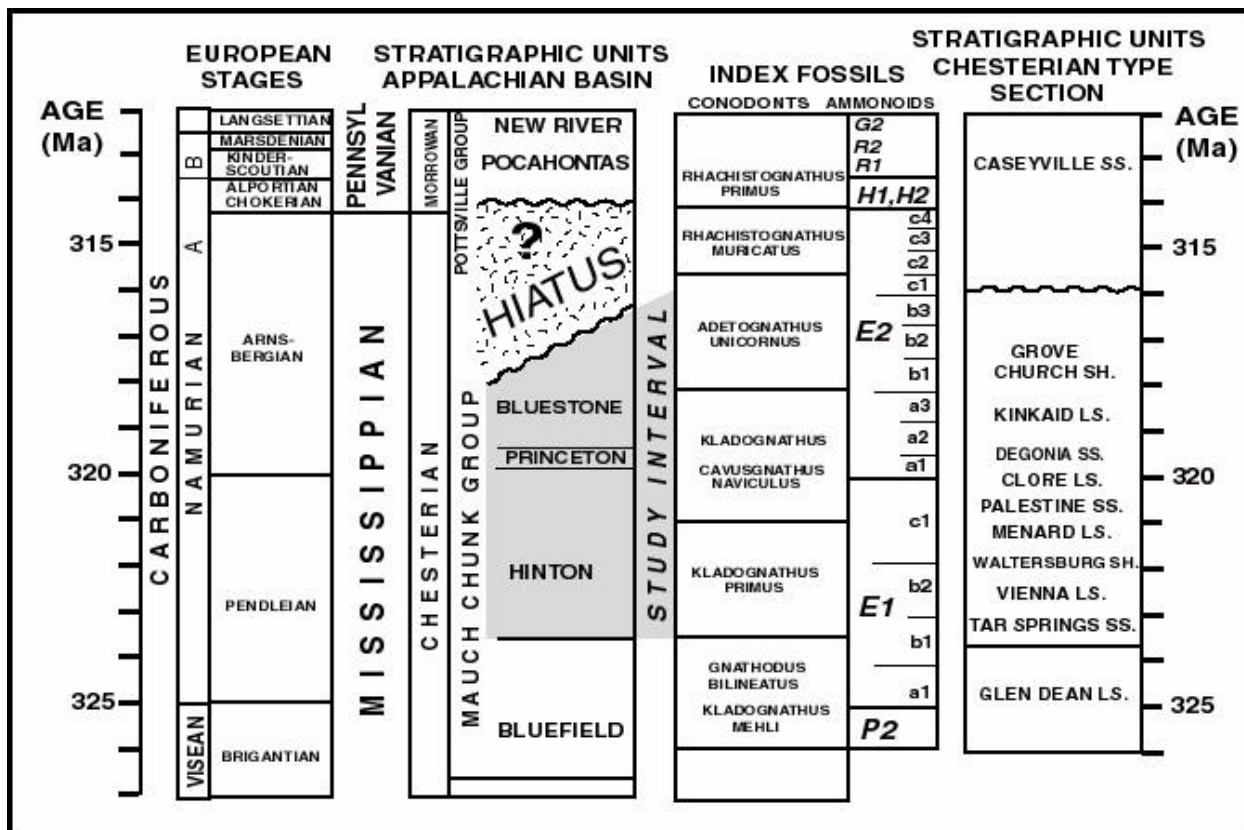

Figure 1.4- Correlation chart showing biostratigraphic and geochronologic constraints for the study interval (adapted from Jones, 1996). Marine invertebrate and plant megafossils suggest that the Missisppian-Pennsylvanian boundary in the Appalachian Basin represents a depositional hiatus that spanned the late Arnsbergian, Chokerian, and Alportian stages of Europe.

correlation with the Grove Church Shale of the Chesterian type section (Collinson et al, 1971; Gordon and Henry, 1981; Gordon et al., 1982; Sutherland and Manger, 1984; Repetski and Henry, 1983).

The total time represented by the study interval is based on correlation with European successions for which geochronological constraints are available. The biostratigraphic data indicate that the base of the study interval correlates with the lower Namurian A (Pendleian) of Europe, and that the top of the study interval is no younger than the middle/ late Namurian A (middle Arnsbergian). Using the correlation chart of Jones (1996; Fig. 1.4) that scales Namurian biostratigraphic zones according to revised SHRIMP geochronological dates (Riley et al., 1995; Roberts et al., 1995), the study interval records ~7 million years of deposition (323.5 Ma to 316.5 Ma).

Although the study interval has been regarded as part of a continuous Mississippian-Pennsylvanian sedimentary succession (Englund and Randall, 1981), several lines of evidence suggest that the systemic boundary is correlative with the global Mid-Carboniferous boundary that marks a second-order eustatic lowstand (Vail et al., 1977; Saunders and Ramsbottom, 1986; Cleal and Thomas, 1996). Paleobotanical analyses suggest that Early Pennsylvanian megaflora are correlative with that of the European Namurian B (Kinderscoutian)(Kosanke, 1984; Gillespie and Pfefferkorn, 1979). The abrupt disappearance of Namurian A index flora and the appearance of species adapted to a swamp habitat is suggestive of a dramatic climate change (from seasonally dry to everwet tropical conditions) in the paleoequatorial belt (Phillips and Peppers, 1984; Van der Zwan; 1985; Jennings, 1986; Cecil et al., 1985; Cecil, 1990). This floral evidence is paralleled by a similar distinct break in ammonoid zones which suggest that basal Pennsylvanian strata in the Appalachian basin are significantly younger that the uppermost Mississpppian units (Fig. 1.4; Gordon et al., 1982; Pfefferkorn and Gillespie, 1982; Ramsbottom and Saunders, 1984). Lithologic evidence for a pre-Pennsylvanian erosional/ depositional hiatus includes the truncation of Mississippian strata as seen in subsurface data west of the basin depocenter (Fig. 1.3; cf. Wilson, 1984; Beuthin, 1994). In eastern outcrops, the top of the study interval is locally truncated by the Lower Sandstone Member of the Pocahontas Formation (Early Pennsylvanian; Englund, 1974). Where the Lower Sandstone Member is absent, a well-developed paleosol atop the uppermost Mississippian units has been interpreted as marking a extended depositional hiatus (Beuthin, 1997). This hiatus is likely equivalent to the European Mid-Carboniferous boundary (basal Chokerian stage) that demarcates a widespread eustatic fall at 314.3 Ma (Fig 1.4: Saunders and Ramsbottom, 1986; Ross and Ross, 1988; Riley et al., 1995; Cleal and Thomas, 1996).

METHODS

Sections were measured through the study interval in isolated exposures (large roadcuts) throughout the outcrop belt. By virtue of the dissected plateau topography, most historically named lithologic units (cf. Reger, 1926; Wilpolt and Marden, 1959; Miller, 1974; Englund and Randall, 1981) could be traced along strike to construct composite stratigraphic sections. Facies were described in regard to color, lithology, upper and lower contacts, fossils, sedimentary structures, and paleocurrent directions. Sandstone, shale, and limestone units that matched with distinctive gamma-ray and bulk density signatures in nearby well logs were correlated in over 100 gas well logs from Mercer, Raleigh, Wyoming, McDowell, and Logan counties (Fig. 1.1).

STRATIGRAPHIC FRAMEWORK

The base of the study interval is identified in outcrop and in well-logs by the abrupt appearance of the Stony Gap Member (Hinton Formation). The top of the study interval is less distinct because of patchy outcrop control and the lack of density signatures in cased-hole wireline logs. Nevertheless, regional cross-sections (Figs. 1.4, 1.5, 1.6) were leveled along a datum (the Mississippian-Pennsylvanian boundary) that preserves the continuity of inferred time-lines defined by underlying marine units (Little Stone Gap Limestone and Pride Shale members) and overlying Pennsylvanian markers (Lower Sandstone Member; Pocahontas #3 coal) (cf. Rehbein et al., 1981; Rice, 1985; Shepard et al., 1986).

Based upon surface and subsurface mapping, it is possible to subdivide the ~600 m-thick study interval into at least 17 stratigraphic packages that are bounded by unconformities of subregional to regional extent (Figs. 1.5, 1.6).
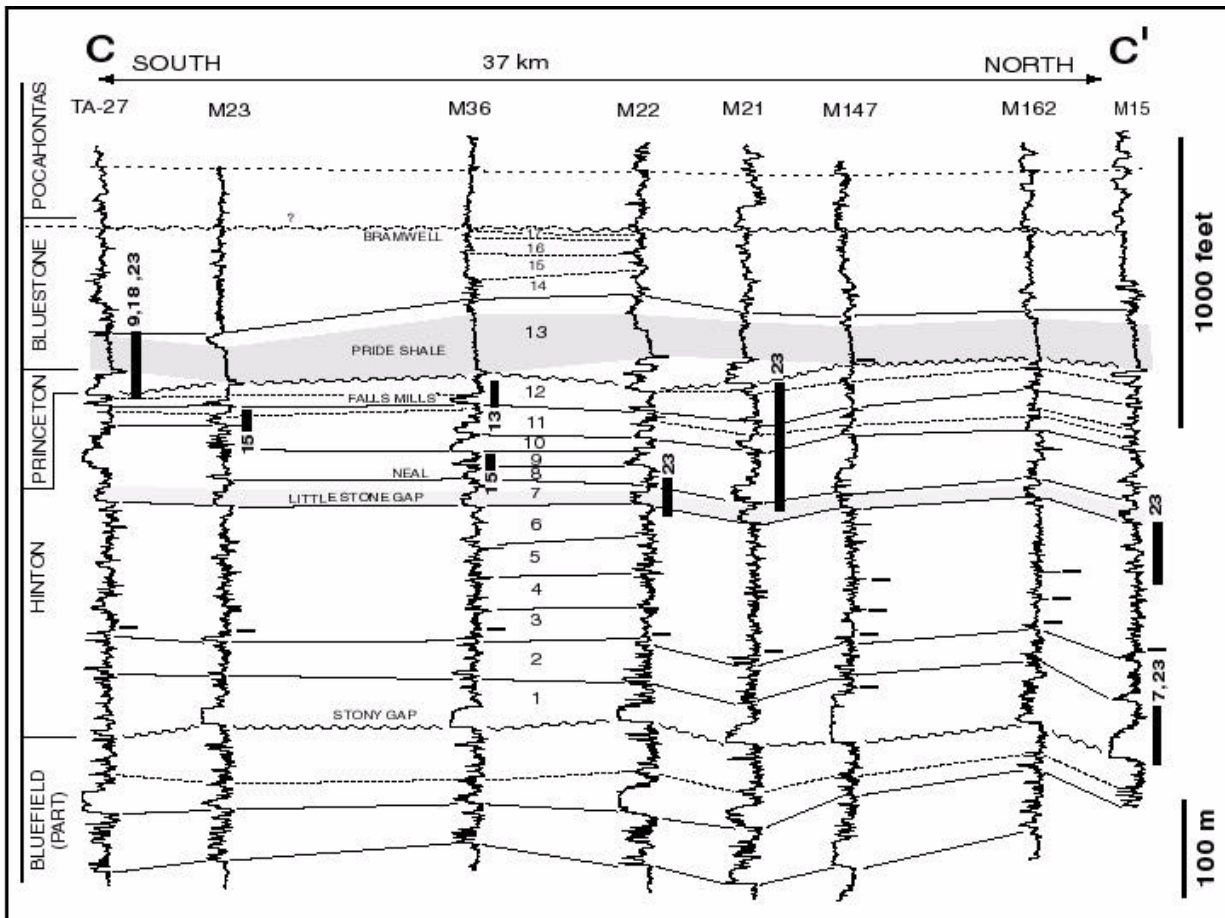
Figure 1.5- North-south (strike) cross-section based on gamma-ray logs from locations in the eastern part of the study area. Refer to Figure 1.1 for well locations and to Figure 6 for explanation of symbols. Major unconformities define two long-term sequences. Interpreted high-frequency sequences are numbered 1-17 with tentative correlations shown for cased-hole parts of wireline logs.

Each package can be subdivided into a suite of facies associations that are seen to recur throughout the study interval. The facies associations characterize four principal depositional systems: 1) incised fluvio-estuarine fills, 2) marine facies, 3) progradational prodeltaic facies, and 4) coastal plain/ floodplain facies.
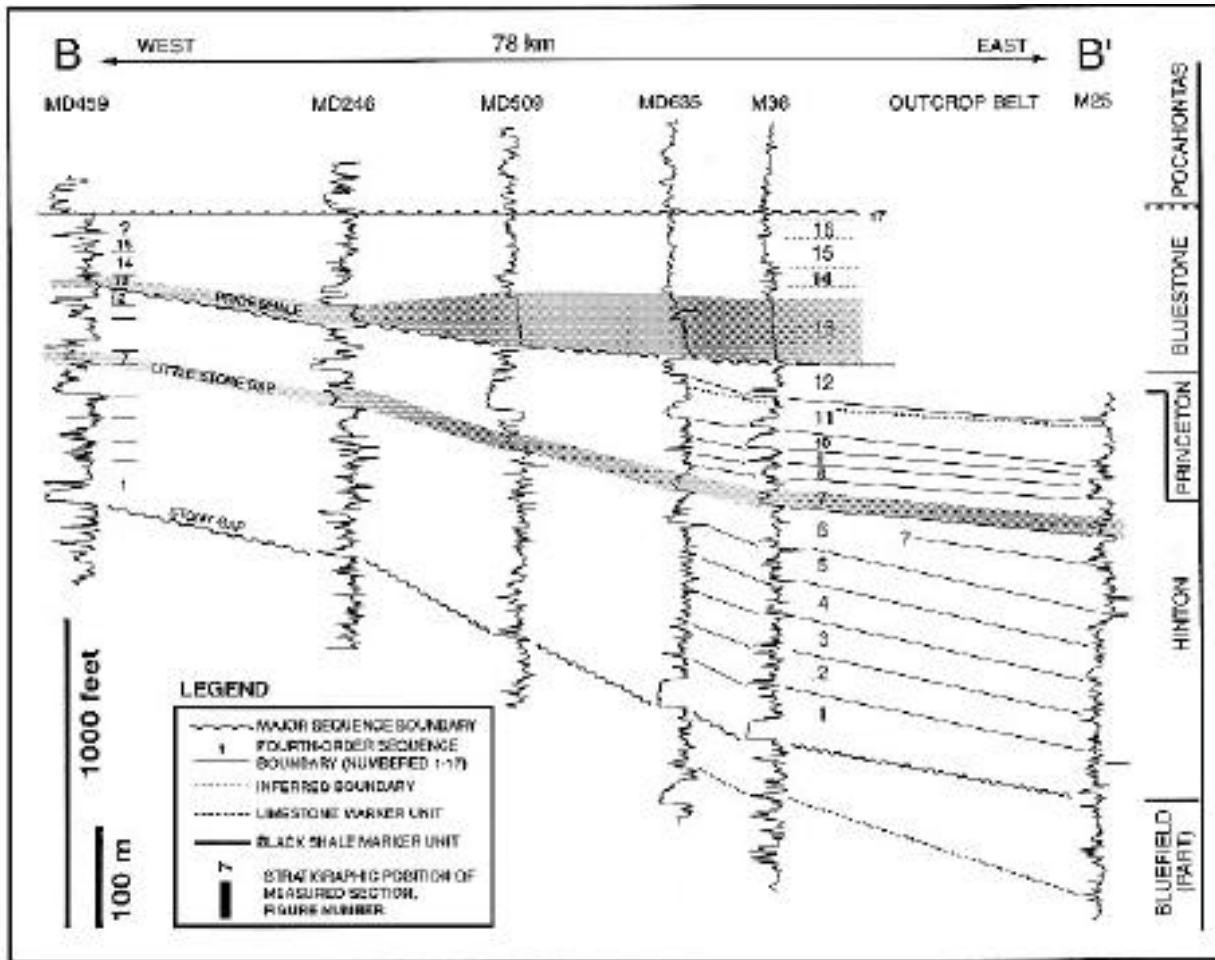
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Figure 1.6- East-west (dip) section though central Mercer and McDowell counties. Refer to Figure 1.1 for well locations. High-frequency sequences amalgamate westward; ~12 sequences are recognized in westernmost well-logs.

DEPOSITIONAL SYSTEMS

Incised Fluvio-Estuarine Valley Fills

Incised valley deposits comprise the nine major sandstone bodies in the study interval (Figs. 1.2, 1.5, 1.6). All sandstone bodies are unconformity-based and fine upwards. In outcrop, major and minor incised valley deposits are distinguished on the basis of their thickness, areal extent, and stratigraphic position with respect to regional unconformities.

Major Incised Fluvio-Estuarine Successions

Stony Gap Member : Description - The Stony Gap Member (Hinton Formation) that defines the base of the study interval ranges in thickness between 3 and 50 meters (Figs. 1.5, 1.6). This unit can be traced for ~50 km along strike in the outcrop belt and up to ~70 km to the west in the subsurface. Regionally, the Stony Gap Member pinches out northeast and southwest of the study

area (Huddle et al., 1956; McDowell and Schultz, 1990; Nolde, 1994). In outcrop, the Stony Gap Member lies unconformably upon pedogenically altered red mudstones of the
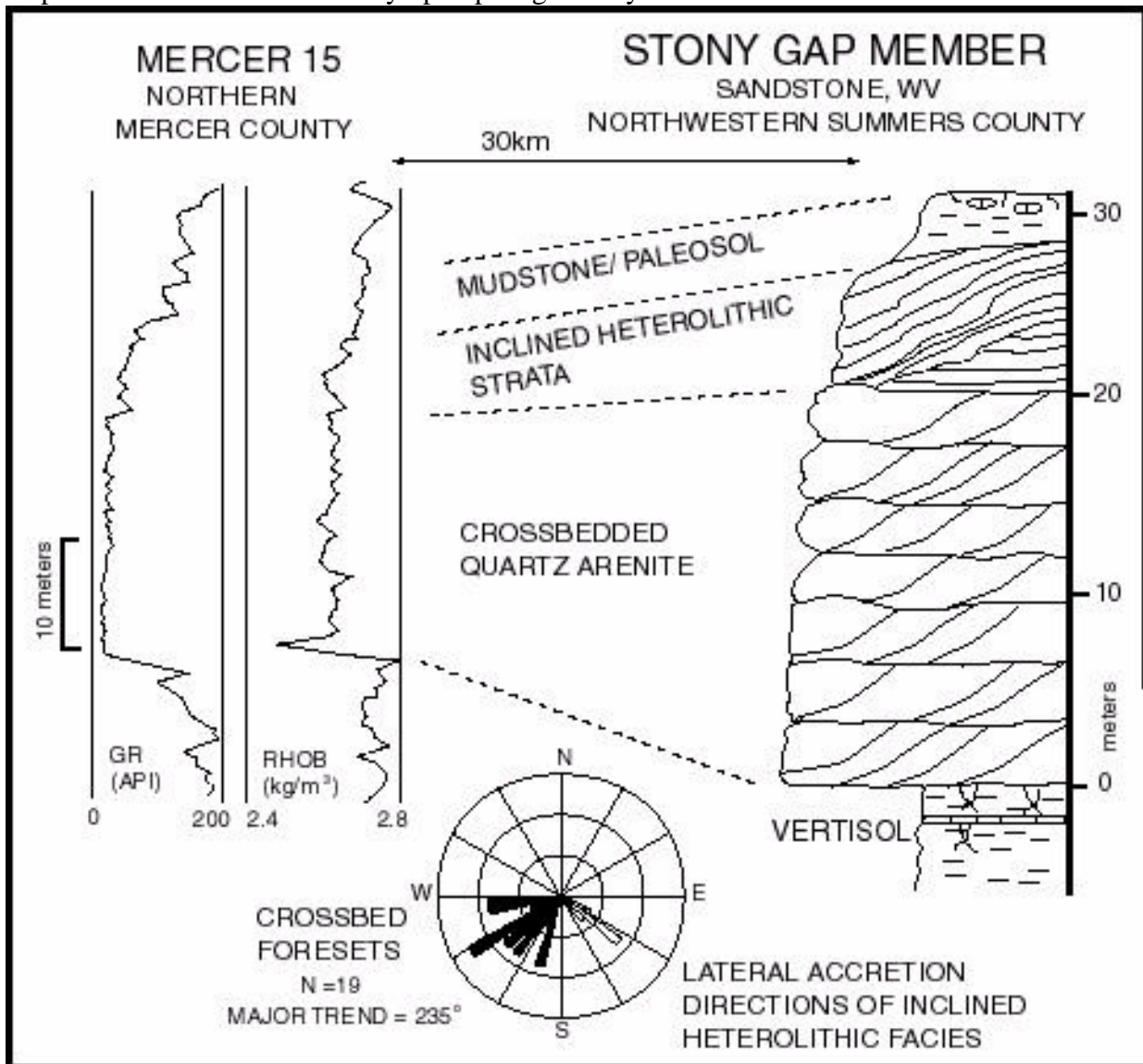

Figure 1.7- Measured section of the Stony Gap Sandstone Member exposed in the northwest part of Summers County at Sandstone, West Virginia. Also shown are paleocurrent data for arenite facies (crossbed foresets) and inclined heterolithic facies. Gamma ray (GR) and bulk density (RHOB) log signatures from Mercer 15 (30 km to southwest).

underlying Bluefield Formation. This sharp basal contact with older mudstones is seen in well logs throughout the study area, and suggests that the Stony Gap Member (or drillers' upper Maxon Sand) overlies a regional unconformity (Figs. 1.5, 1.6; cf. Reger, 1926; Nolde, 1994).

Outcrop observations indicate that the Stony Gap member is made up crossbedded quartzose arenite and heterolithic facies (Fig. 1.7). Log signatures confirm that these two facies associations are also present in the subsurface (Figs. 1.5, 1.6, 1.7). The quartzose arenite interval ranges in thickness from a few meters to 40 meters. The base of this interval is rarely conglomeratic with intrabasinal siltstone- and shale-clasts up to 10 cm in length. Overlying medium to coarse-grained quartzose arenites contain trough and tabular-planar crossbeds. Foresets in the upper part of the arenite interval commonly display a sigmoidal shape. Foreset

azimuths indicate a southwesterly paleoflow direction (Fig. 1.7). Such paleoflows are consistent with lateral thickness variations of large exposures along the New River (Fig. 1.1) that indicate generally southwesterly paleochannel trends through the outcrop belt.
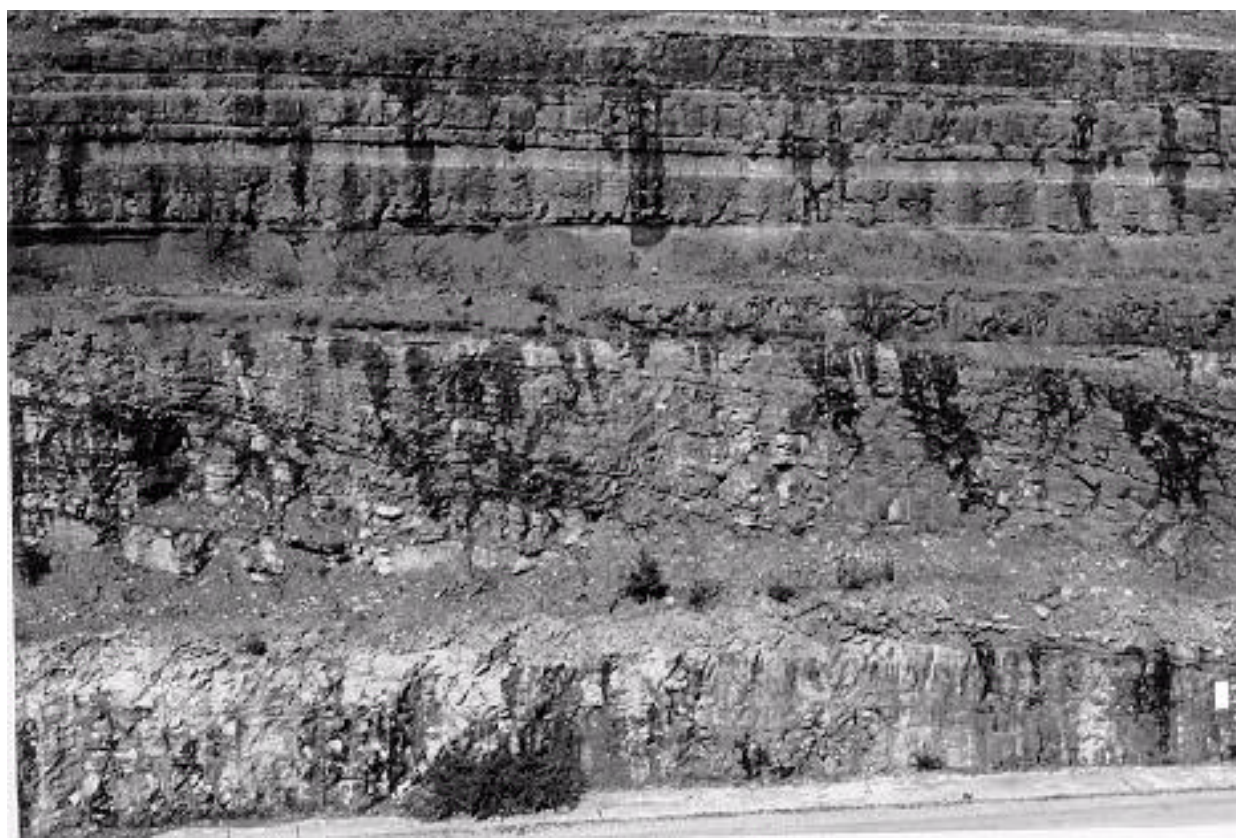

Figure 1.8- Inclined heterolithic facies in the upper part of the Stony Gap Member at Sandstone, WV. These units are interpreted as estuarine, point-bar deposits. Scale bar is 1 meter.

The overlying heterolithic facies association defines large-scale inclined strata (Fig. 1.8) that dip at angles of a few to 23 degrees and are up to 10 m in height. The heterolithic unit is comprised of a series of laterally accreted fills that lie upon a sharp or erosional contact with underlying crossbedded sandstones. In the few outcrops where three dimesional exposure is available, it can be demonstrated that accretion took place toward the southeast (Fig 1.7). Internally, the inclined strata are comprised of fine-grained sandstones and mudstones that are interlayered on a millimeter to meter scale. Locally, the older inclined strata are sandstone dominated, with successive fills becoming finer grained. Despite the rhythmic appearance of the sandstone-mudstone couplets, no conclusive systematic (cyclic) thickness variations (i.e. bundles) are noted. The heterolithic successions grade upward into a dark silty mudstone and a regionally-developed, gray paleosol that contains calcareous nodules and defines the top of the Stony Gap Member (Fig. 1.7).

Interpretation - Because the Stony Gap Member has a significant areal extent and quartz-rich lithology, Englund et al. (1979) proposed that this unit represents a marine deposit of coalesced offshore bars. However, the erosional basal contact and the association of facies in the Stony Gap Member point instead to a fluvio-estuarine origin.

The basal contact that truncates terrestrial facies in the outcrop belt is interpreted to define a surface of subaerial exposure. The areal extent of this unconformity and the lateral thickness variation of the Stony Gap Member suggests that several sub-valleys were incised

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within a regional, southwesterly-oriented paleovalley system. Local incision up to tens of meters into the underlying units indicates that the unconformity developed during a time of lowered relative sea level.

During the initial stages of relative sea-level rise, paleotopographic lows within the paleovalley system infilled with crossbedded quartzose arenites. Southwestwardly-dipping foresets that characterize the lower part of the Stony Gap Member are suggestive of downstream dune migration within unconfined channels of a braided fluvial system. This crossbedded sandstone facies may represent the deposits of shallow perennial braided sand-bed rivers as described by Miall (1985, 1996). The significant thickness of this facies suggests that sediment supply kept pace with accommodation during initial infilling of the paleovalley system. The upward transition into the finer-grained heterolithic facies marks a time of marine influence when relative sea-level rise outstripped sediment supply and the paleovalleys were transformed into estuaries. Clay drapes and sandstone/ mudstone couplets that characterize the inclined heterolithic units are suggestive of tidal influence (cf. Thomas et al., 1987). In the northeast outcrop belt, these units likely represent tidally-influenced point bar deposits that developed in the sinuous, upper reaches of the estuarine system (cf. Dalrymple, 1992; Ainsworth and Walker, 1994). Rooted horizons and paleosols at the top of the Stony Gap Member developed during times of exposure that likely relates to lateral migration or avulsion of estuarine channels. The total thickness of the Stony Gap Member suggests that accommodation space for the fluvial and estuarine fill was provided by a relative sea level rise of tens of meters.

Princeton Formation : Description - The Princeton Formation is another example of a major incised valley-fill succession. It is a distinctive unit that separates the finer-grained Hinton and Bluestone formations throughout most of the outcrop belt (Reger, 1926; Cooper, 1961; Weems and Windolph, 1986). In the northern part of the study area, the Princeton Formation is absent and basal marine units of the Bluestone Formation directly overlie redbeds of the Hinton Formation. Detailed correlation indicates that the Princeton Formation thins to the north and west; it is recognized only in selected well logs from the subcrop belt (Fig. 1.5). Similar relationships reported by Englund and Thomas (1990), Couzens et al. (1991), and Nolde (1994) indicate that the Princeton Formation pinches out to the southwest in accordance with regional thinning of Upper Mississippian strata (Fig. 1.3).

The Princeton Formation is comprised of two facies associations: conglomerate-lithic arenite; and dark mudstone with arenite lenses, thin coals, and heterolithic channel fills (Fig. 1.9). In central Mercer County (Fig. 1.1), basal polymictic conglomeratic facies overlie an erosional contact marked by several meters of relief into underlying fossiliferous shale and limestone.
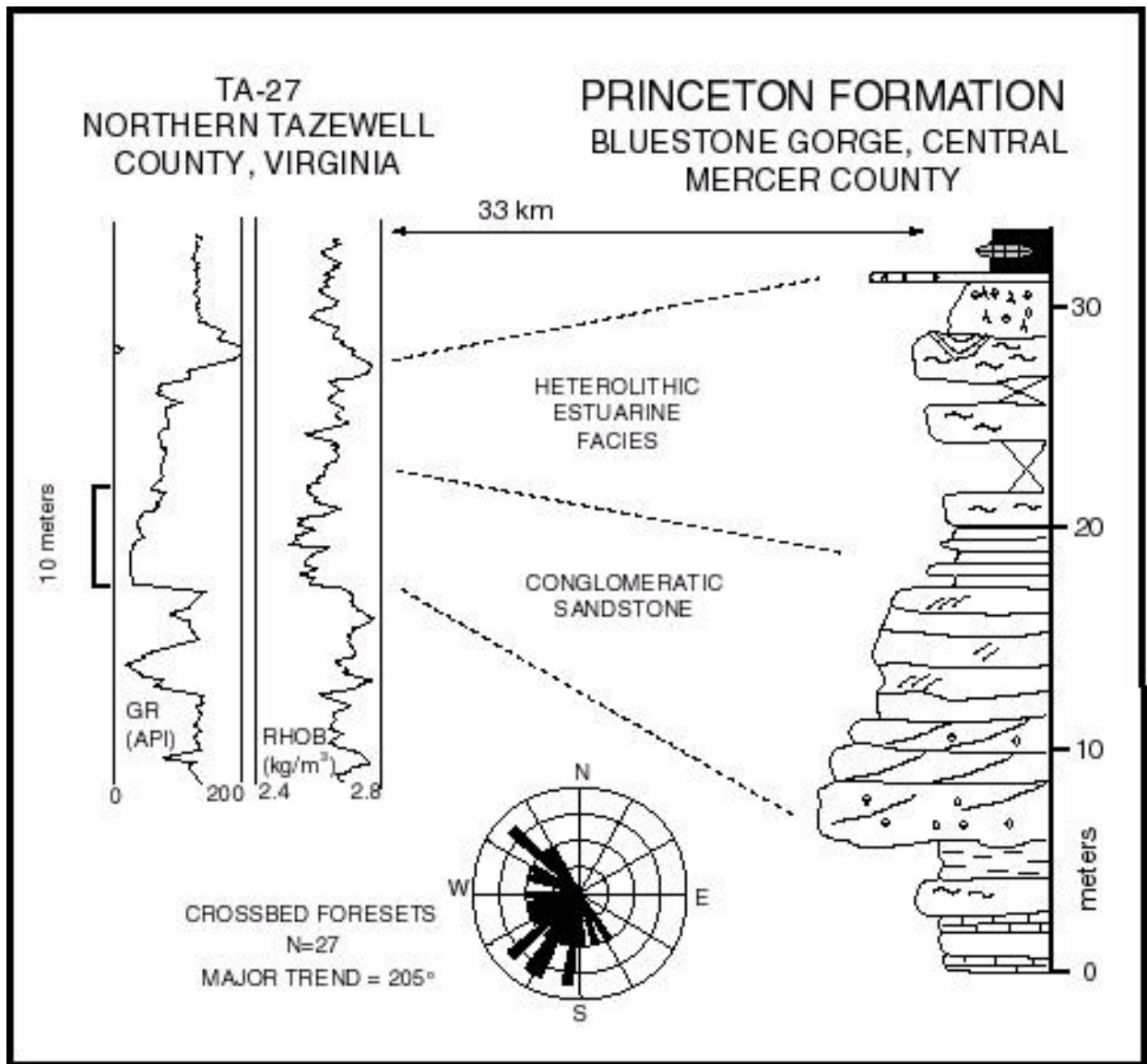
Figure 1.9- Measured section of the Princeton Formation based on outcrops along the Bluestone River in the central part of Mercer County. Also shown are paleocurrent data from basal, fluvial sandstones. Log signatures from TA-27 (33 km to the southwest).

Well logs confirm that this erosional surface truncates the uppermost Hinton Formation throughout the study area (Figs. 1.5, 1.6). Pebble and cobble sized clasts in the conglomerate facies include rounded quartz pebbles and lesser amounts of shale, sandstone, limestone, metamorphic lithics, and nodules of caliche and siderite. Plant fragments and molds of logs up to 30 cm in diameter are preserved locally. Overlying coarse- to medium-grained sandstones fill a series of nested channels. Tabular-planar and trough crossbeds indicate west to southwesterly paleoflow (Fig. 1.9).

The upper Princeton Formation is dominated by dark mudstones that contain plant fossils and siderite nodules. The mudstone encloses discontinous sandstone lenses up to 5 m thick and 100 m wide that consist of fine- to medium grained, flat-laminated to ripple-bedded sandstone. Along the margins of the sandstone lenses are heterolithic rhythmites that occupy shallow (few meter-deep) and narrow (2-5 m) channels (Fig. 1.10). The heterolithic channel fills contain root traces, abundant plant fragments and rare bivalves. Some heterolithic units show systematic

thickness variation of millimeter to centimeter-scale sandstone/ mudstone couplets (Fig. 1.11a, b). A few of the heterolithic channel fills directly overlie thin coals. Some horizons in the upper Princeton Formation have been pedogenically altered to leached, gray-white paleosols.

Uppermost Princeton facies are erosionally truncated along a regionally-developed surface that is marked by a sharp contact with overlying dark shales of the Bluestone Formation. This surface is defined by a fossil lag that includes marine bivalves, gastropods, and brachiopods. Locally, a thin (<1m) tabular bed of quartz-pebble conglomerate is associated with the fossil lag (Fig. 1.12).

Figure 1.10- Tidal creek fill with slump blocks in upper part of Princeton Formation along the Bluestone River Gorge, central Mercer County. Scale bar is 1 meter.
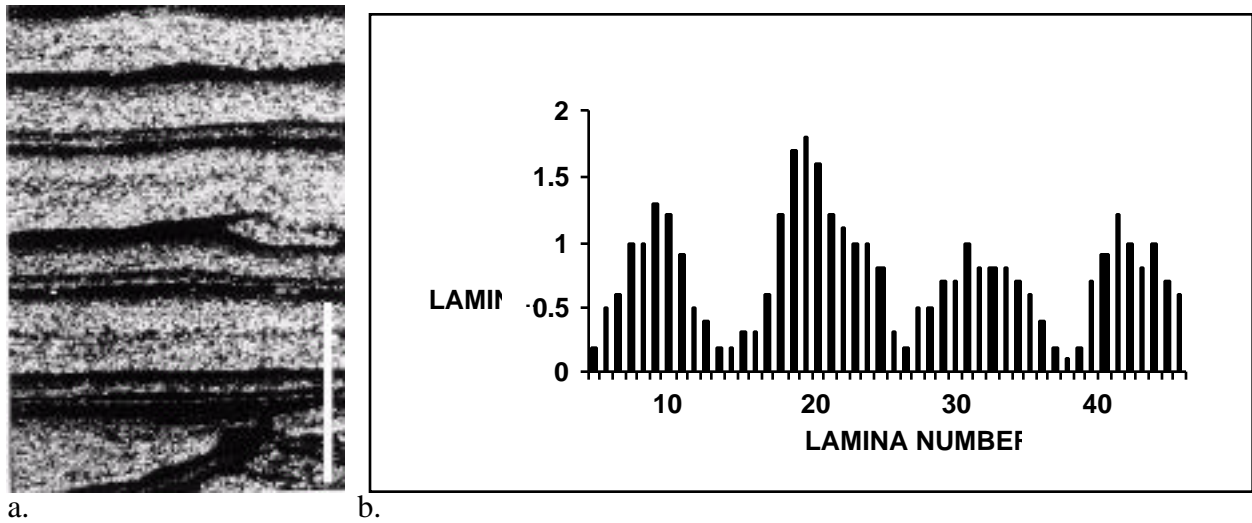
Figure 1.11- a.-Laminae bundling in thin section of tidal creek facies in the Princeton Formation. Successive thick-thin couplet pairs reflect deposition during unequal semi-diurnal tidal flows. Scale bar is 2 mm. b.- Histogram of laminae thicknesses showing 12-15 layer bundles that record neap-spring cyclicity.

Figure 1.12- Fossiliferous quartz pebble conglomerate overlying estuarine mudstones of the Princeton Formation and overlain by the dark Pride Shale Member of the Bluestone Formation. This fossiliferous, laterally discontinuous bed is interpreted as a ravinement bed that developed during transgression of the Princeton paleovalley. Scale bar is 1 meter.

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Interpretation - The Princeton Formation has been previously interpreted as a beach/ barrier bar deposit (Weems and Windolph, 1986), a deltaic distributary sandstone (Wrightstone, 1985), and a braided river deposit (Pinnix, 1992). These varied interpretations reflect lateral facies variation, but may also reflect miscorrelation of subsurface data.

Based upon numerous outcrop exposures and several well-logs in the type area, it is herein suggested that the Princeton Formation represents an incised valley-fill succession comprised of braided-fluvial and tidal-estuarine deposits. The irregular basal erosional surface is part of a regionally developed unconformity that incised into the Hinton Formation (cf. Reger, 1926). The fact that the conglomeratic facies outcrops in a ~20 km wide, northeast-southwest trending belt through Mercer County suggests that preservation was restricted to a deeply incised paleovalley in the eastern part of the study area. The variety of clasts indicates that the Princeton fluvial system eroded underlying units of the Hinton Formation, and served as a conduit for extrabasinal clasts from the east. The channel-form scours, nested channel-fills, and coarse nature of the conglomerate facies are consistent with deposition in a gravel-bed braided-fluvial system (cf. Miall, 1996). These facies are interpreted to mark initial fluvial aggradation in response to a rise in base level.

The vertical progression of facies in the Princeton Formation are consistent with that proposed for incised fluvio-estuarine settings under conditions of relative sea level rise (Demarest and Kraft, 1987, Dalrymple, 1992). The mudstone facies that dominate the upper part of the Princeton Formation were deposited in a less energetic environment, and are interpreted as inner estuarine and estuarine-margin marsh deposits. Such fine-grained deposits mark a time when the depositional site received little coarse sediment from either fluvial or marine inputs (Demarest and Kraft, 1987, Fletcher et al, 1990, Dalrymple, 1992). Siderite nodules and plant fragments in these mudstones are consistent with brackish to fresh water conditions (cf. Postma, 1982). The tabular sandstone bodies are interpreted as sandflat deposits or axial estuarine channel deposits that developed at sites of comparatively high-energy currents. Mud-layer couplets and wavy-bedding in the tabular sandstone bodies are suggestive of fluctuating energy conditions resulting from tidal influence. Convincing evidence for tides is provided by laminae bundling within the heterolithic channel fills that lie along margins of the tabular sandstone bodies. Laminae counts indicate that bundling is consistent with semidiurnal and neap-spring tidal cycles (Fig. 1.11b). Some of these tidal rhythmites preserve an uninterrupted short-term (< 0.5 yr) record of deposition that can be resolved into individual flood and ebb tidal flows (cf. Archer et al., 1995). The rhythmic channel fills are interpreted as tidal creek deposits. Thin coal beds and paleosols in the upper Princeton Formation mark relatively short-lived periods of exposure that may define times of minor sea-level stillstands when estuarine depositional systems shifted laterally or prograded during the transgressive history of the paleovalley (cf. Clifton, 1982; Reinson, 1992).

Depositional models suggest that the final stage of estuarine sedimentation is marked by sand deposition in elongate tidal ridge or barrier/ inlet facies (Dalrymple, 1992). The Princeton Formation, however, is not characterized by large sand bodies that may define such estuarine mouth facies. Instead, the regionally-developed erosional surface that truncates the estuarine units (Fig. 1.12) marks a regional contact with overlying marine facies. This abrupt boundary is a ravinement surface that developed during transgression of the Princeton paleovalley. A tidal origin for ravinement (cf. Dalrymple, 1992) is consistent with the fact that tidal signatures are seen in both the estuarine fill deposits and the overlying sediments (Miller and Eriksson, 1997). Such surfaces develop during transgression as high-energy, outer estuarine currents erode underlying facies in areas seaward of the highstand shoreline (Nummedal and Swift, 1987; Dalrymple et al., 1992; Reinson 1992). Although these currents may be restricted to channels that separate outer estuarine sand ridges, the amalgamation of the channel scours can produce an areally extensive ravinement surface (Demarest and Kraft, 1987). Estuarine-mouth sandstone facies may be removed except where they occur as incised tidal channel fills into underlying mudstone. It seems likely that coarse-grained estuarine mouth facies tracked eastward during transgression, and were not preserved within the outcrop belt. Evidence of such migration across the transgressed Princeton paleovalley includes the fossiliferous, quartz-pebble conglomeratic

bed that locally defines the ravinement surface. Ravinement can remove the entire estuarine succession such that the ravinement surface coincides with the sequence boundary on transgressed inland interfluves (Nummedal and Swift, 1987; Demarest and Kraft, 1987). This relationship is seen in the nothern part of the study area (southeastern Fayette County) where the Pride Shale directly overlies terrestrial red beds of the upper Hinton Formation.

Minor Incised Fluvio-Estuarine Successions

Minor incised fluvio-estuarine successions are represented by 7 fining-upward units between 3 and 20 m thick (Figs. 1.5, 1.6). These include the Hackett, Neal, Tallery, and Falls Mills members; two unnamed units of the Hinton Formation; and the Glady Fork Member of the Bluestone Formation (Fig. 1.2; cf. Englund and Randall, 1981). Where outcrop permits, it can be demonstrated that these units extend laterally for several kilometers, but vary in both thickness and character as mapped in isolated exposures. Regional correlation using well logs confirms that these units define discrete stratigraphic intervals that persist for tens of kilometers along strike in the eastern part of the study area (Fig. 1.5). Three types of successions are recognized based upon the dominant facies: conglomeratic, arenitic, and heterolithic. The minor incised valley fill successions display lateral variability in thickness. In several cases the conglomeratic, arenitic, and heterolithic successions reflect the local character of the same stratigraphic unit. In particular, the arenitic successions thin laterally into heterolithic successions.

Conglomeratic Minor Incised Valley-Fills : Description - The conglomeratic Glady Fork Member (Bluestone Formation) is developed in southeastern Mercer County and in the northwest corner of Raleigh County (Fig. 1.1). In both localities, this unit occupies southwest-directed paleovalleys up to 1 km wide that were incised tens of meters into the underlying Pride Shale Member. In the thickest sections, the valley-fill facies are up to 20 m thick. Large slump blocks and load structures are locally present along channel margins below the channel fill succession (cf. McColloch, 1986). The basal conglomeratic sandstone includes pebble to cobble-sized clasts of limestone, siderite, shale, and quartz pebbles. Plant fragments up to several centimeters in diameter are common. The overlying medium- to coarse- grained sandstone is trough crossbedded within a series of nested channels separated by lenses of shale. Trough axes and crossbed foresets within the nested channels suggest a westerly to southwesterly paleoflow direction that is consistent with the large-scale channel trend and isolated flute casts at the base. The crossbedded sandstones are overlain by red paleosols.

Interpretation - The restriction of the conglomeratic Glady Fork Member to deeply incised channels within a thick succession of dark shales suggests that deposition was preceded by a significant lowering of sea level. The presence of slump blocks along channel margins and the nested character of the basal valley-fill facies are suggestive deposition in a high-energy, sandy, braided-fluvial system that combed across the confines of the incised valley (cf. Miall, 1996). Fining upward into crossbedded sandstones and shale lenses with plant fragment is consistent with deposition in a braided fluvial system. The crossbedded sandstones suggest variable, yet generally westward paleoflow. Upward transition into a thick paleosol suggests that this conglomeratic succession marks the fluvial infilling of an incised valley that did not evolve into an estuary.

Arenitic and Heterolithic Incised Valley Fills : Description - The few meter to 20 m-thick arenitic successions are best exemplified by the Neal and Falls Mills Sandstones (upper part of the Hinton Formation) (Fig. 1.13) as developed in the southern and central parts of Mercer County (Figs. 1.1, 1.2) These units overlie red-green paleosols along a smooth erosional contact. A basal thin (< 0.5 m) conglomeratic deposit is found locally at the base of successions.
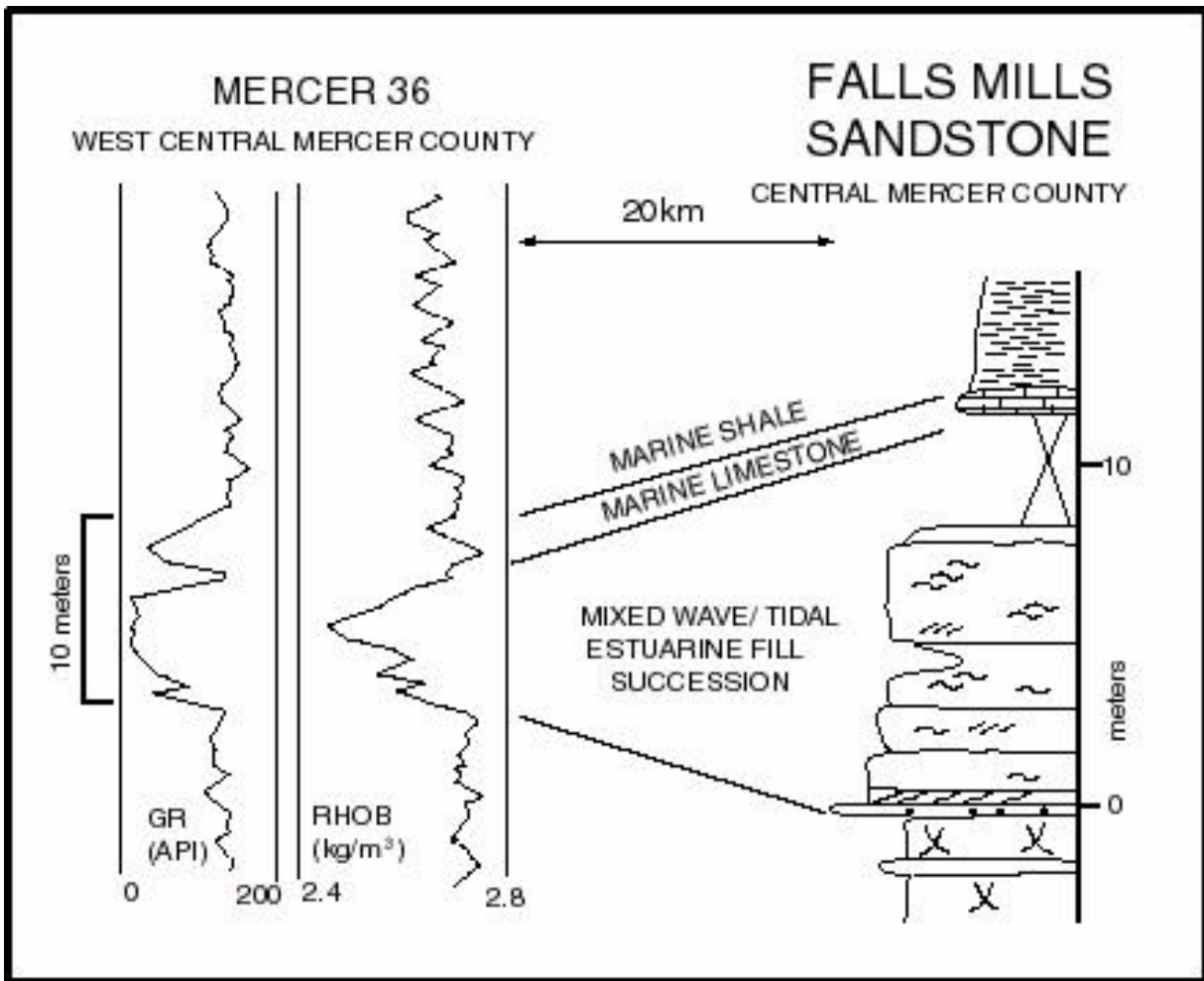
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Figure 1.13- Measured section of arenitic, wavy-bedded, incised valley-fill succession (Falls Mills Member of the Hinton Formation) exposed in central part of Mercer County. Gamma ray (GR) and bulk density (RHOB) log signatures from Mercer 36 (20 km to the southwest).

Clasts include intrabasinal limestone, red mudstone, siderite, and caliche nodules. The overlying fine- to medium-grained sandstones are rarely trough crossbedded, but more typically are thin-bedded (few to 10 cm) with plane laminations and cross laminations within symmetrical ripples (Fig. 1.14). Cross laminations indicate both east and west paleoflow. These successions grade upward into dark siltstone, and are capped by either marine red paleosols or fossiliferous limestone (Fig. 1.13).
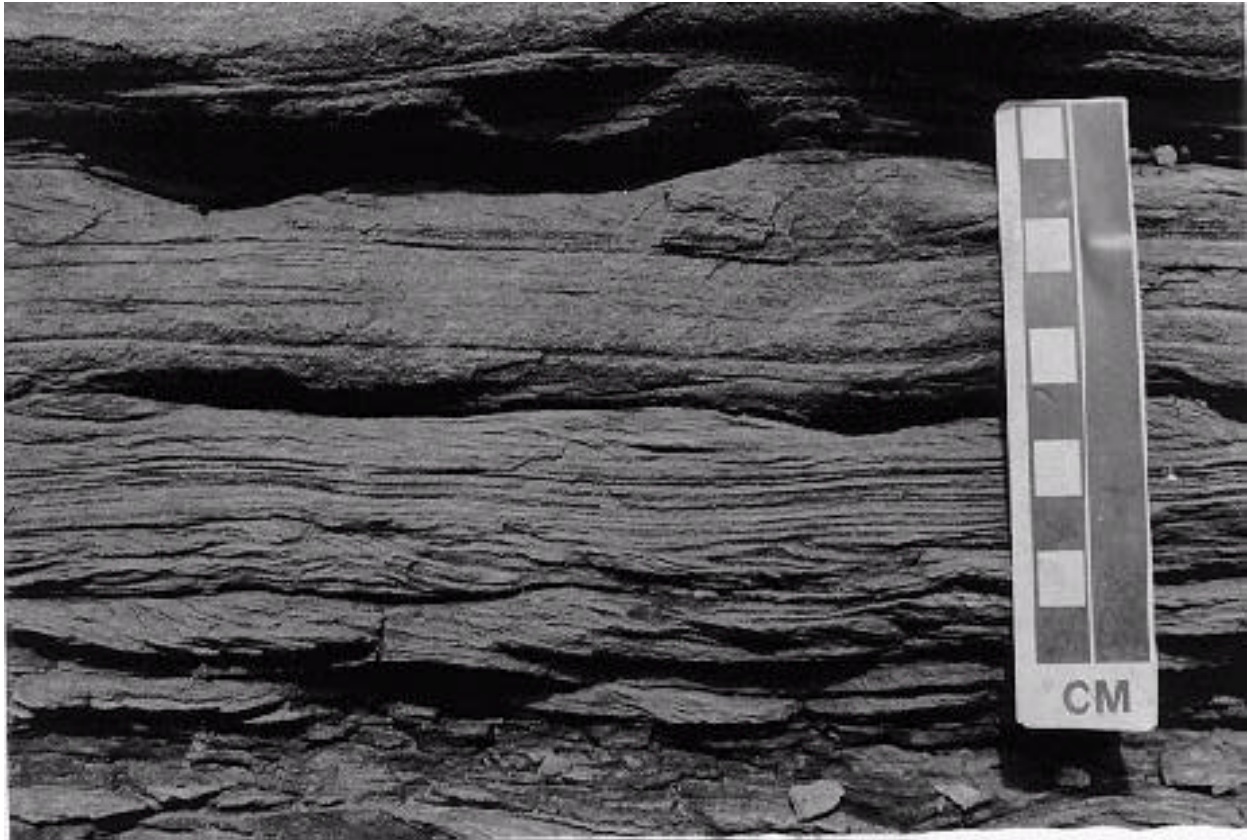
Figure 1.14- Plane laminations and symmetrical, form-discordant ripples in basal part of the Neal Sandstone Member, northern Tazewell County, Virginia. These structures are interpreted as the product of mixed tidal and wave energy within shallow incised paleovalleys. Scale in centimeters.

The thin (3-6 m) heterolithic successions occupy shallow, southerly-directed channels that show several meters of relief into underlying terrestrial (paleosol) facies (Fig. 1.15). Dark shales with abundant plant fossils or thin coals (up to 30 cm thick) locally underlie the channel-fill successions. The heterolithic fills consist of up to 4 m of inclined to horizontally-laminated fine-grained sandstone with mudstone drapes. In some successions, centimeter-scale thick-thin couplets (Fig. 1.16) comprise 18- to 28-laminae bundles similar to those described above for the upper Princeton Formation. Most of these successions contain multiple horizons of root traces. The successions typically fine upward into dark silty mudstone, but some examples are capped by a thin (1-3 cm) bivalve shell bed and an overlying unit of dark shale.
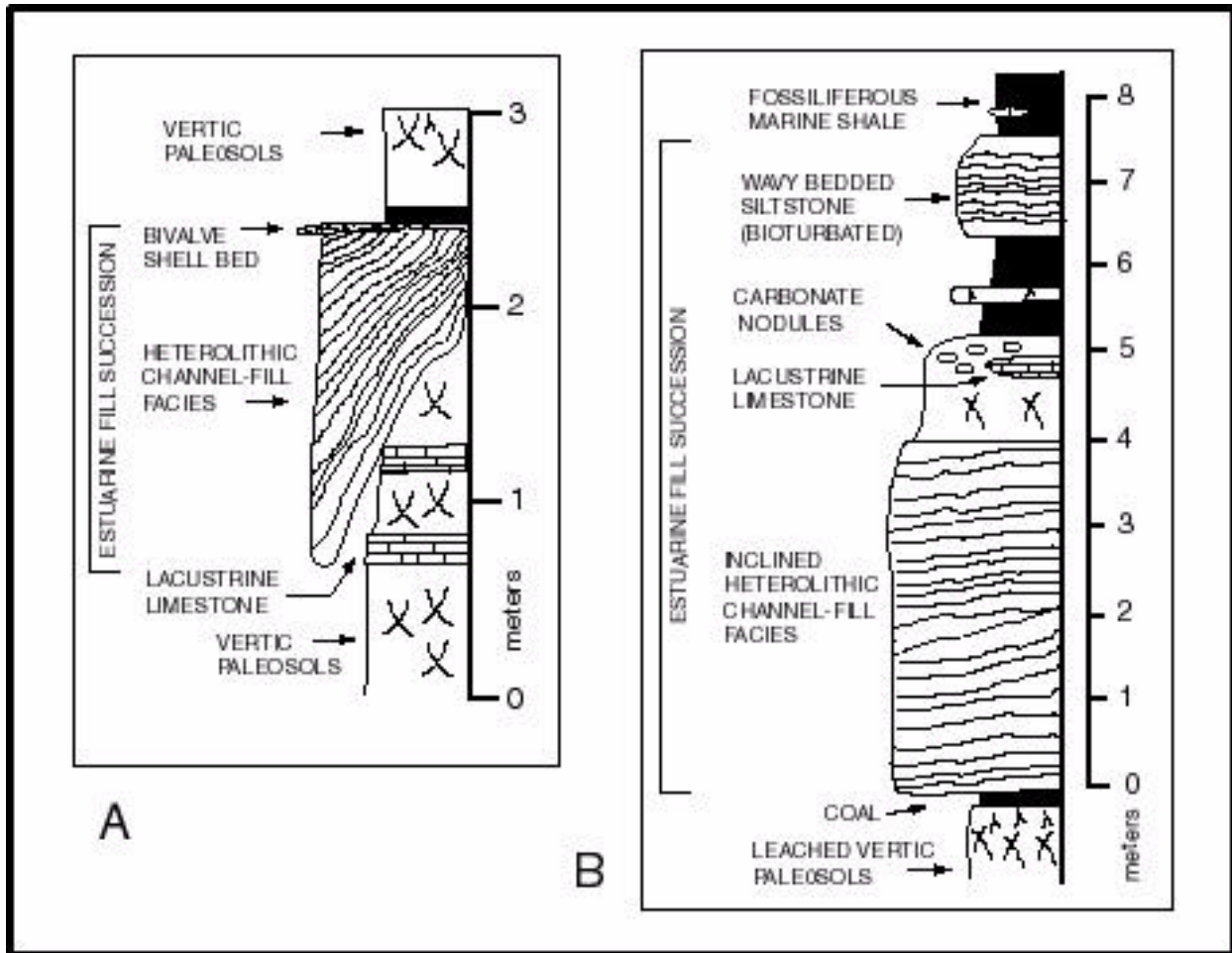
Figure 1.15- Measured sections of thin, heterolithic estuarine successions in the upper part of the Hinton Formation exposed in northeastern Raleigh county. a.- Thin succession is bounded by paleosols. b.- Coal-based heterolithic succession is overlain by fossiliferous marine shale.
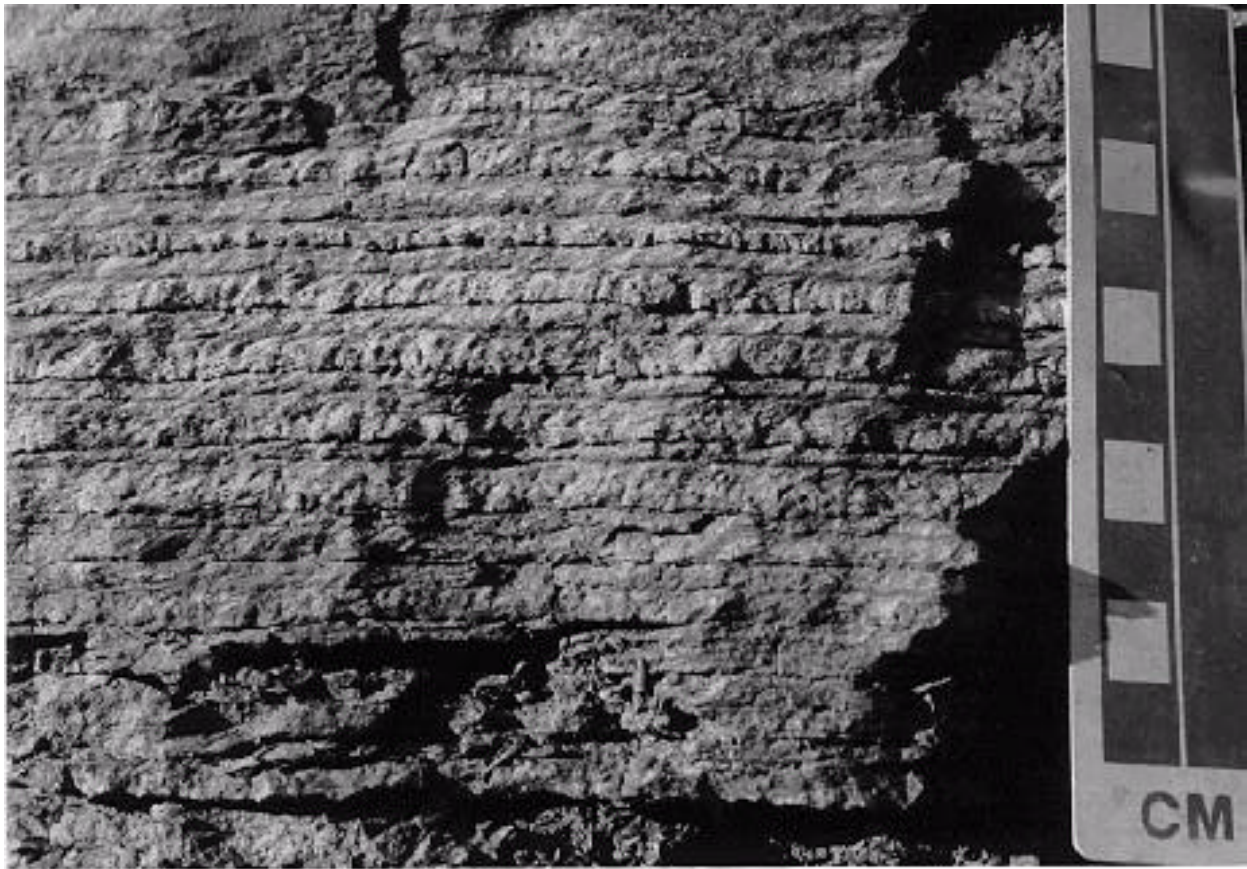
Figure 1.16- Tidal rhythmites within coal-based heterolithic incised valley-fill succession shown in Figure 1.15b.. Thick-thin pairing of sandstone-mudstone couplets indicates deposition by semi-diurnal tides.

Interpretation - The arenitic and heterolithic incised valley fills are interpreted as predominantly estuarine and subordinate fluvial deposits that accumulated in shallow incised paleovalleys. Basal contacts that show a few meters of relief into underlying terrestrial units indicate that incision during relative sea level lowstand was minor compared to the major and conglomeratic incised valley fill successions described above. The coals below the heterolithic successions are suggestive of locally wet conditions prior to transgressive estuarine sedimentation. Inclined heterolithic strata comprising most of the valley fill (Fig. 1.15a) record lateral migration of tidal channels (cf. Ainsworth and Walker, 1994). Well-developed, thick-thin laminae couplets indicate deposition by semi-diurnal tides (cf. deBoer et al., 1989). In heterolithic units where distinct cycles are absent, the presence of pin-stripe laminations, and crude laminae bundles are suggestive of tidal sedimentation. It is likely that tidal energy may have been sustained or amplified within funnel shaped embayments. Wave overprint is indicated by planar laminations, and form-discordant symmetrical, ripples that reflect east-west oscillatory flow (Fig. 1.14; cf. de Raaf et al., 1977; Greb and Archer, 1995). The thickest of these minor incised valley fills are interpreted to reflect transgression of dissected coastal lowlands following times of moderate (e.g. 10 m) relative sea level fall. The thin successions of limited areal extent may define minor episodes of relative sea level change, but may also represent the lateral/ updip equivalent of thicker successions (Fig. 1.5). These thin successions mark the landward fringe of marine influence upon the coastal plain during transgressive episodes. Subsequent inundation is marked by the marine units (shale and limestone) that overlie most of the estuarine units.

<u>Marine Facies</u>

Laterally persistent, fossiliferous limestone, shale, and siltstone units comprise a minor component of the study interval. These units are key horizons for both outcrop and subsurface correlations, and yield the fossils that permit general biostratigraphic correlation with midcontinental U.S. and European successions.

<u>Limestone Facies</u>

Description - The most significant carbonate unit in the study interval is the Little Stone Gap Member, a limestone and calcareous shale succession up to 18 m-thick in the middle part of the Hinton Formation (Fig. 1.2). With a relatively uniform thickness and a distinctive gamma ray signature (Fig. 1.5), the Little Stone Gap Member has long been utilized as a marker unit in the central Appalachian Basin (Shepard et al., 1986; Wrightstone, 1985). In outcrop, the Little Stone Gap Member overlies red paleosols along a sharp, flat basal contact. The Little Stone Gap carbonate facies include interbedded calcareous shale, lime mudstone, skeletal wackestone, packstone, and grainstone (Fig. 1.17). Fossils include brachiopods, bryozoans, trilobites, and bivalves (Gordon and Henry, 1981; Hoare, 1993). The upper contact of the Little Stone Gap member is marked by a calcareous shale that grades upward into a red paleosol. Thin marine limestone units are present in the Hinton Formation. These are typically argillaceous, skeletal packstones to wackestones that occur as thin (< 1 m-thick), discontinuous beds within shale facies.

Interpretation - The non-channelized basal contact of the Little Stone Gap Member with underlying paleosols suggests that coastal inundation occurred rapidly over most of the study area. The thickness of this limestone unit, together with the diversity of marine invertebrates are suggestive of an extended period of open-marine conditions. Regionally, this unit is regarded as the most widespread marine deposit within the predominantly terrestrial Upper Mississippian section (Englund and Randall, 1981). Neal (1989) and Place and Beuthin (1998) recognize shallowing-upward trends within the Little Stone Gap Member that define two, transgressive-regressive episodes of deposition within subtidal, lagoonal, and tidal flat/ beach environments. Fossiliferous limestone units of limited thickness in the Hinton Formation are interpreted to define comparatively short periods of marine deposition during which clastic influx was relatively low.
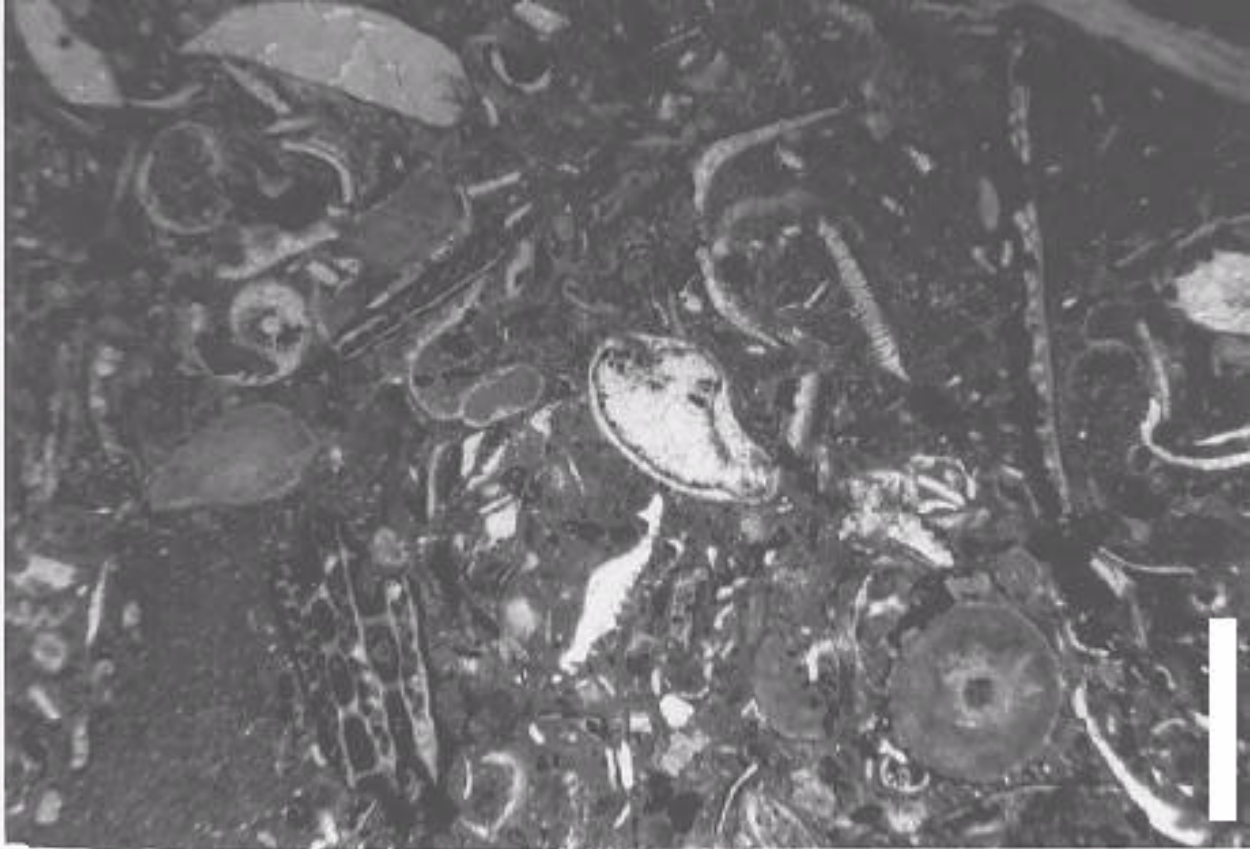
Figure 1.17- Photomicrograph of fossiliferous Little Stone Gap Member. Brachiopod, trilobite, crinoid, and bryozoan fauna is suggestive of open marine conditions. Scale bar is 2 mm.

Fossiliferous Shale Facies

Description - Fossiliferous dark shale units occur as thin (few cm to few m-thick) horizons that overlie incised-valley successions, or less commonly, overlie red or green (leached) paleosols (see Figs. 1.5, 1.6). Thin beds of argillaceous limestone are common within the black shales. Fossils include thin-valved bivalves, gastropods, trilobites, crinoids, rugose corals, and rarely, fish fossils (Weems and Windolph, 1986). Fossils rarely are preserved as articulated forms within calcareous beds or carbonate (siderite?) concretions but, more commonly, occur as impressions within the shale. These shale units grade upward into progradational deltaic successions described below, or are erosionally truncated by terrestrial sandstones and paleosols. The shale units correlate with the moderate to highly radioactive gamma ray signatures in well logs (Figs. 1.5, 1.6). Some of the shale units that overlie incised valley-fill successions correlate with those that overlie terrestrial units beyond the preserved extent of the paleovalley fills.

Interpretation - The presence marine fossils and the lateral extent of the shale units is suggestive of regional marine inundation. In some beds, filter feeding invertebrates (crinoids and corals) are suggestive of relatively open-marine conditions. More commonly, the shale units are characterized by low-diversity faunal assemblages that indicate restricted conditions. The abrupt occurrence of the shales within the paleosol intervals suggests that flooding occured as punctuated, short lived episodes in interfluvial areas of the coastal plain. The thick shale units that overlie the incised valley/ estuarine successions likely mark longer periods of inundation over infilled paleovalleys.

Fossiliferous Siltstone

Description - The Bramwell Member (Bluestone Formation) that defines the top of the study interval is an upward coarsening, siltstone dominated unit with lesser shale to fine-grained sandstone, and is up to 36 m thick (Fig. 1.2; Englund and Randall, 1981; Beuthin, 1997). Basal beds of dark, carbonaceous shale overlie red, terrestrial mudstone units of the Red Member (Bluestone Formation) along a flat, non-erosive contact. This shale contains freshwater or brackish water ostracodes (*Darwinula*), bivalves (*Naiadites, Anthraconaia*), and inarticulate brachiopods (*Lingula*)(Englund and Randall, 1981; Englund and Thomas 1990). Overlying beds of green-gray calcareous siltstone and fine-grained sandstone contain a diverse assemblage of marine fauna (bivalves, brachiopods, bryozoans, trilobites, ostracodes, crinoids, cephalopods) (Gordon and Henry, 1981; Hoare, 1993). Bedding is locally obscured as a result of bioturbation. Limestone concretions up to 0.5 m in diameter are present within the shale beds (Englund et al., 1986). At localities where the Bramwell Member is less fossiliferous, wavy-bedding is defined by interlaminated siltstone and shale. The top of this unit is locally overlain by a thick paleosol and nonmarine, heterolithic beds of the Upper Member (Bluestone Formation), but is more commonly truncated along an erosional contact with the Lower Sandstone Member of the Pocahontas Formation (Englund, 1974; Beuthin, 1997). The Bramwell Member thins to the northwest across the outcrop belt (Strickter, 1981), and is not present in most well logs because it is cut out along the erosional base of the Pocahontas Formation (cf. Figs. 1.3, 1.6).

Interpretation - The Bramwell Member is interpreted as marking a regional flooding event that, unlike the marine units that overlie the incised valley fill successions, was not preceded by a significant sea level fall. The flat contact of carbonaceous shale or coal upon underlying redbeds suggests that initial transgression transformed the coastal plain into a paludal or peat swamp environment. The calcareous lenses, limestone concretions, and diverse invertebrate fauna in the overlying siltsones are suggestive of open marine conditions following regional inundation.

Progradational Prodeltaic Deposits

Rhythmically-bedded Siltstone-Shale

Description - The Pride Shale Member is a thick (up to 60 m), rhythmically-bedded unit that defines the basal part of the Bluestone Formation (Fig. 1.2). By virtue of its thickness and distinctive gamma-ray signature, the Pride Shale is recognized throughout the study area except in the westernmost well-logs (Figs. 1.3, 1.5, 1.6). The base of this unit is marked by a few meters of dark marine shale described above. The bulk of the Pride Shale Member is a coarsening-upward succession of thinly interlaminated dark shale, siltstone, and fine-grained sandstone (Fig. 1.18). Submillimeter- to millimeter-thick siltstone/ shale couplets define laterally continuous laminae that are characteristically organized into a hierarchy of laminae bundles (Miller and Eriksson, 1997). Fossils include thin-valved bivalves, plant fragments, and carbonaceous imprints of shrimp-like arthropods. The southwestward dip of the rhythmic beds in large exposures is typically a few degrees greater than the regional dip of the enclosing units.
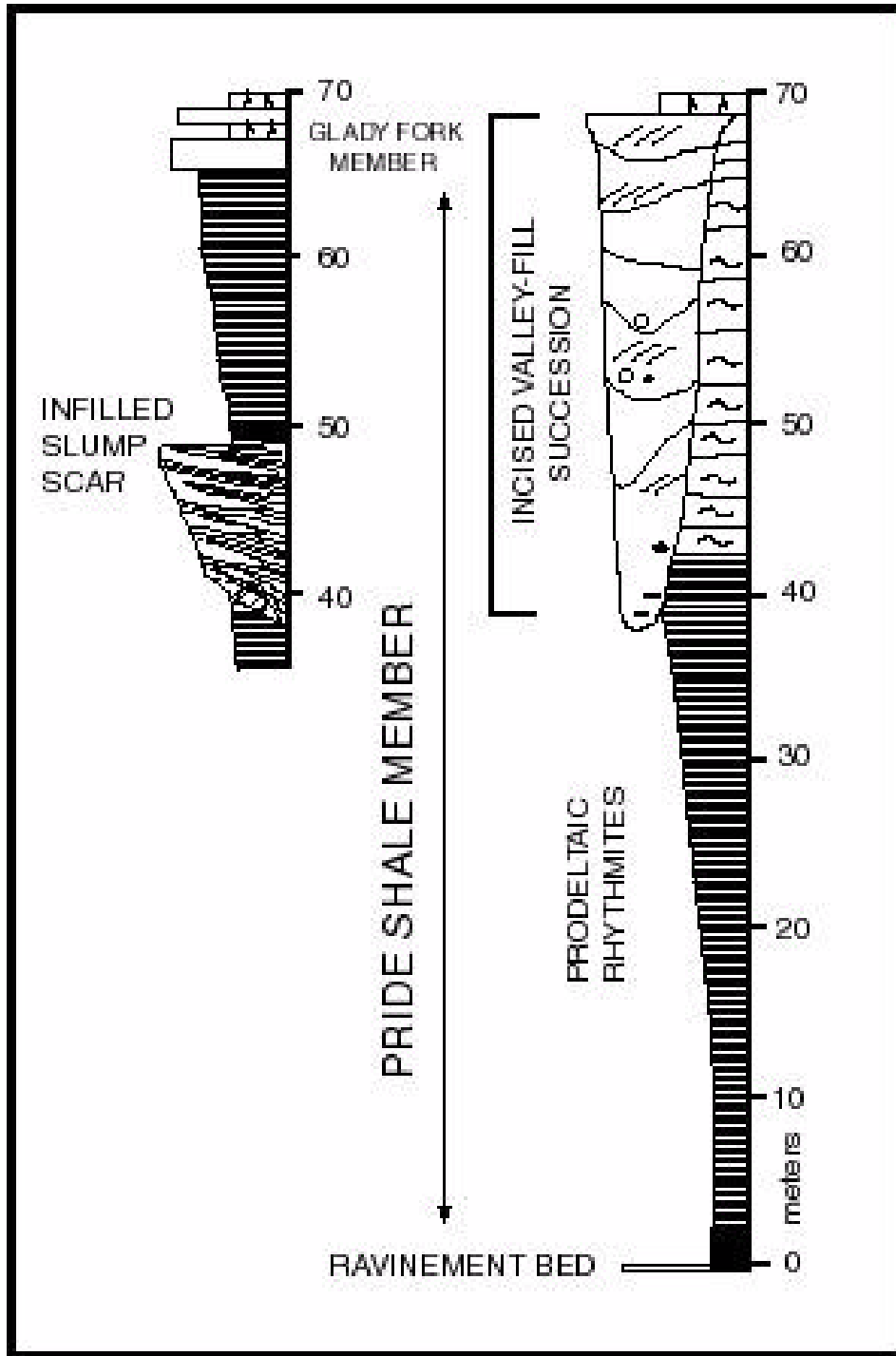
Figure 1.18- Measured section of Pride Shale and overlying Glady Fork Member from central part of Mercer County. These units are interpreted as a coarsening-upward deltaic succession that is truncated locally by a conglomeratic, minor incised valley-fill succession.

Locally, large (500 m across) discontinuities within the rhythmic beds define a series of channels or scoop-shaped infills. The Pride Shale typically grades upwards into a wavy and ripple-bedded

sandstone (which is mapped as the Glady Fork Member) that contains root traces and passes upward into a 1-3 m thick paleosol of regional extent. Locally, the Pride Shale is overlain by incised conglomeratic channel facies of the Glady Fork Member.

Two thinner, but similar successions are present in the Hinton Formation. These include a unit between the Stony Gap and Hackett Sandstone Members (lower part of sequence 3) that is up to 25 m-thick, and the 0-30 m thick Upper Member that defines the top of the Hinton Formation (middle part of sequence 12)(Fig. 1.5). Neither of these units diplays the distinctive laminae bundling seen in the Pride Shale, yet both coarsen upward from a basal shale or limestone into thin-bedded, sparsely-fossiliferous, gray-green siltstones. Gamma-ray and bulk density signatures of these successions are variable, yet nevertheless are comparable to those of the Pride Shale. In outcrop, the lowermost unnamed unit is overlain by red, sandstones and mudstones whereas the Upper Member of the Hinton Formation is sharply truncated by incised, conglomeratic channel facies of the Princeton Formation.

Interpretation - The Pride Shale is an extensive prodeltaic deposit of a prograding tide-dominated delta (Miller and Eriksson, 1997). The thickness of this unit indicates that infill was preceded by significant regional inundation. The hierarchy of cycles reflect fluctuating tidal energy related to semi-diurnal, diurnal, neap-spring, and anomalistic tidal periodicities; and seasonal pulses of sediment modulated by monsoonal fluvial discharge (Miller and Eriksson, 1997). The fossil plant material and thin-valved or delicate invertebrates are consistent with a nearshore environment characterized by salinity fluctuations and high turbidity that served to exclude most benthic fauna. The clinoform geometry is suggestive of a gently southwestwardly-sloping prodeltaic slope. Deltaic progradation is indicated by the coarsening-upward of shales into rippled sandstones of probable distal bar or distributary mouthbar origin (cf. Wright, 1985) and ultimately into rooted, terrestrial mudstones.

The two thinner, shale-dominated units in the Hinton Formation are similarly interpreted as deltaic deposits based on their coarsening upward from shales or fossiliferous shales into thin-bedded marine sandstones. Both are both truncated by terrestrial facies (incised valley-fill successions) that mark a basinward shift of fluvial systems onto underlying progradational marine facies in response to a fall in relative sea level.

Coastal Plain / Floodplain Facies

Terrestrial Redbeds

Description - Approximately 50% of the study interval is comprised of alternating channelized, and non-channelized, tabular sandstone bodies; heterolithic channel-fill facies; red mudstone (paleosols); and nonmarine carbonates. These facies comprise the thick (tens to hundreds of meters) redbed successions that dominate the lower part of the Hinton Formation and the upper part of the Bluestone Formation (Fig. 1.2).

Channelized sandstones are erosively based and range from a few to 10 m thick and from several to 50 m wide. Basal contacts of the larger channelized sandstones truncate underlying paleosols and tabular sandstone bodies (Fig. 1.19), whereas the smaller channelized sandstones are erosive into paleosols (Fig. 1.20). Internal fill consists of very-fine to fine grained lithic wacke. Rarely, basal beds are conglomeratic with locally derived mudstone and siltstone clasts up to 3 cm in diameter. Channel-fills are typically compound consisting of several upward-fining or laterally-accreted units. Sedimentary structures include plane lamination, large-scale low-angle trough crossbeds, and inclined beds of alternating sandstone and mudstone. Mudcracks are locally present within the mudstone-dominated parts of the compound fills. Tabular, non-channelized sandstone bodies are up to 3 meters thick and persist laterally for hundreds of meters. Most can be traced into incised channel facies (Figs. 1.19, 1.20). Sedimentary structures include planar laminations and small-scale current ripples draped by red mudstone.

Figure 1.19- Channelized, and non-channelized, tabular sandstone bodies in the lower Hinton Formation, Hinton WV. Tabular sandstone bodies are interbedded with paleosols. The sandstones are interpreted as fluvial channel and overbank (crevasse splay) deposits of a low-sinuosity fluvial system. Scale bar is 1 meter.

Figure 1.20- Small channelized sandstone bodies in the upper Hinton Formation, Princeton ,WVinterpreted as tributary channels that incised locally into floodplain paleosols during periods of overall aggradation. Channel sandstone is overlain by a thick vertisol. Scale bar is 1 meter.

At least two examples of inclined heterolithic units overlie vertic paleosols in the redbeds of the lower Hinton Formation (Fig. 1.21). These successions consist of a basal conglomeratic deposit up to 40 cm thick that is overlain by a series of laterally-accreted, interbedded very fine-grained red sandstones and mudstones. The basal conglomerate is comprised of locally derived mudstone chips, carbonate (caliche nodules), and rare vertebrate (amphibian) teeth up to 2 cm in length. The inclined surfaces of the overlying beds dip at angles of a few degrees to 10 degrees.
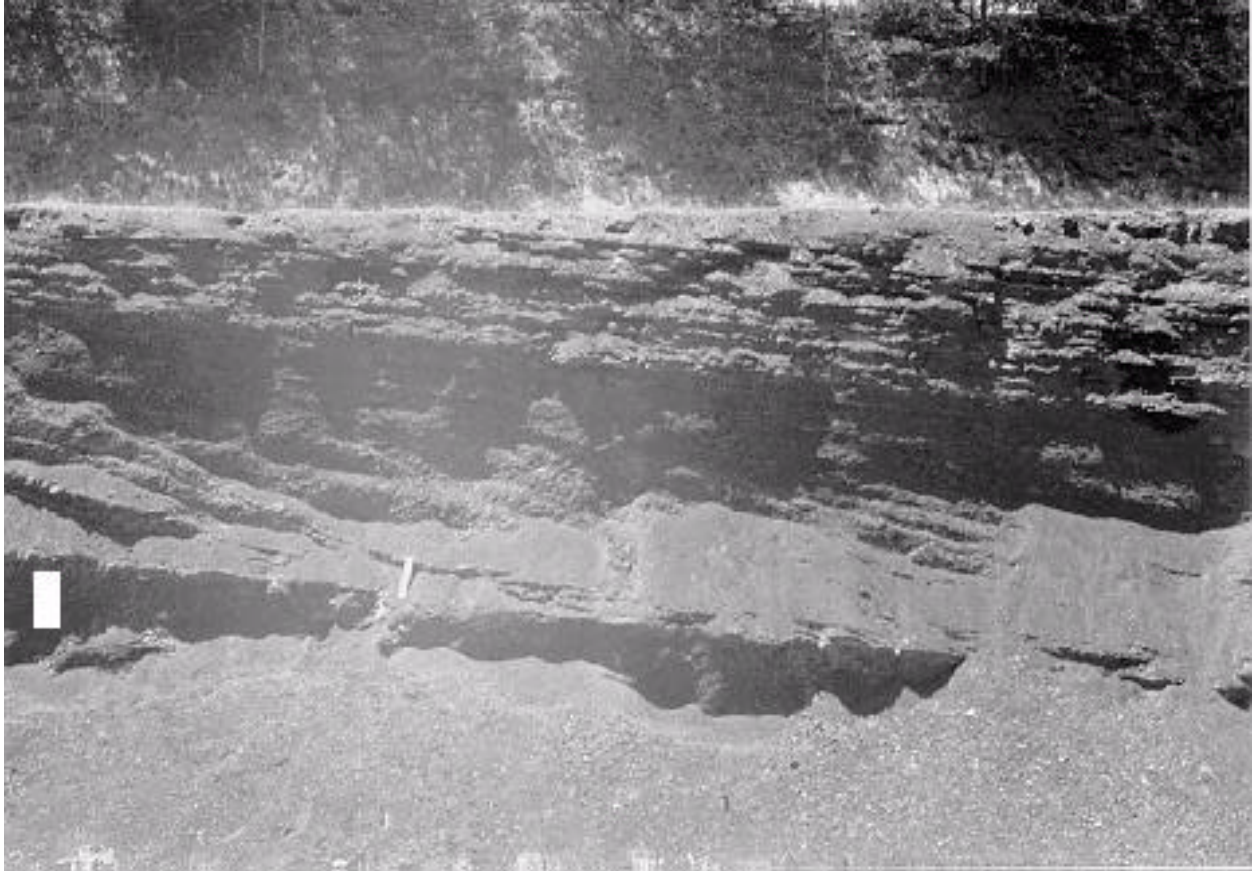
Figure 1.21- Inclined heterolithic strata overlying red vertisol in the lower Hinton Formation , Possum Hollow, West Virginia. Inclined strata represent muddy point bar deposits. Scale bar is 1 meter.

The most common floodplain facies is red, silty mudstone that occurs as tabular beds up to 4 m-thick. Typically, these units have been modified into 2-3 m-thick, mature vertic paleosols (Fig. 1.22). These paleosols are characterized by well-developed pedogenic slickensides and blocky pedogenic fabrics. Irregularly shaped carbonate nodules up to several centimeters in length locally define calcic horizons. Sparse root traces occur as reduction haloes that branch downward over a few to tens of centimeters. The uppermost few to tens of centimeters of the paleosol horizons characteristically are leached to a light green color. The red-green color banding commonly defines several generations of paleosols within an otherwise continuous mudstone unit. Paleosols are thinner and less mature where interbedded with tabular sandstone bodies.

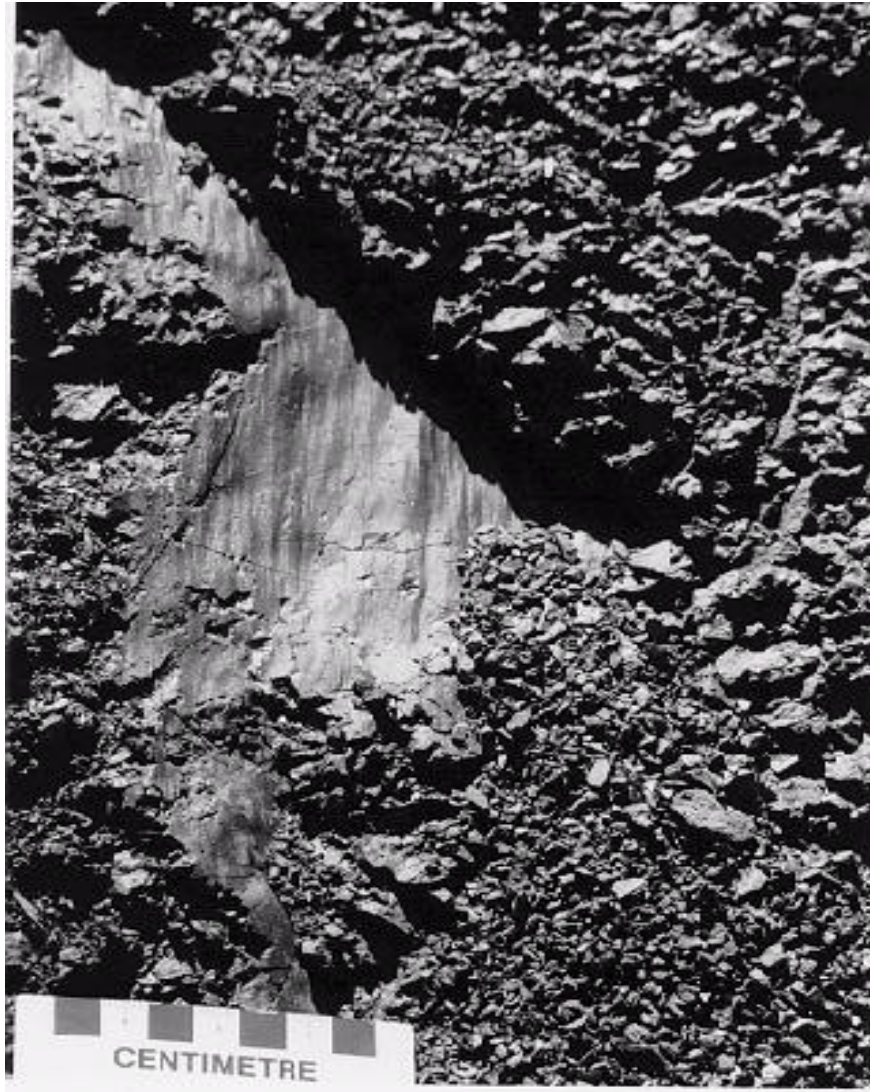Figure 1.22- Vertisol in Hinton Formation in central Mercer County. Note pedogenic slickensides in thick (2 m) B horizon. Thick paleosol reflects times of extended soil formation upon an aggradational floodplain.

Two varieties of nonmarine carbonates are enclosed within the red paleosols. These include gray units of coarse, nodular, argillaceous lime mudstone that occurs as discontinuous units up to 1 meter thick and commonly grade laterally over tens of meters into red paleosols with scattered carbonate nodules. Thinner beds (a few to 40 cm) of yellow to gray lime mudstone form more continuous beds that pinch out over several hundred meters. These units are sparsely fossiliferous with ostracodes, and many are locally brecciated by pedogenesis.

Interpretation - The large and small channellized sandstone bodies are interpreted, respectively, as the deposits of low-sinuosity trunk and tributary fluvial channels that incised into underlying floodplain facies in response to relative sea level fall (cf. Leeder and Stewart, 1996). In the outcrop belt, these channel deposits typically occur as isolated units rather than as components of nested, channel fill complexes. This geometry is corroborated by the fact that few examples of the fluvial channel facies are recognized in well-log data. Parallel laminations, low-angle trough crossbeds, and mudcracks within the channel fill facies are suggestive of episodic deposition

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within ephemeral streams in a semi-arid setting (cf. Tunbridge, 1981). The non-channelized, sandstone bodies are interpreted as crevasse splay deposits related to overbank flooding (cf. Miall, 1996; Bown and Kraus, 1981). Parallel laminated sandstones capped by current ripples and clay drapes are suggestive of high flow stage deposition followed by rapid waning of flow (cf. Tunbridge, 1981).

Inclined heterolithic units represent lateral accretion, point bar deposits of a low-gradient fluvial system characterized by high suspended sediment loads (Turner and Eriksson, 1997; in press). These units lack the rhythmic interlamination of sandstone/ mudstone couplets characteristic of the estuarine successions described previously.

The singular and stacked (multi-generation) paleosols are overbank mudstone deposits that reflect suspension sedimentation lateral to fluvial channels during seasonal (or sporadic) flooding of a low-gradient floodplain. Subsequent pedogenic alteration transformed the mudstones into vertisols. The red (hematite) color, blocky soil textures, clay coatings and pedogenic slickensides reflect alternate wetting and drying within the B horizon of well-drained, clay-rich soils (Retallack, 1988; Mora et al., 1991). Such structures are related to clay-shrinkage and expansion at depths above the level of pedogenic carbonate formation (Wright and Robinson, 1988). Root traces in the form of downward branching haloes also suggest well-drained rather than waterlogged conditions that would favor preservation of original organic material (Bown and Kraus, 1981; Retallack, 1986, 1988). The calcareous nodules also provide compelling evidence for seasonal dry conditions in which moisture deficiency permitted carbonate accumulation in the soil profile (Retallack, 1988; Wright and Robinson, 1988; Mora et al., 1991). Where carbonate precipitation occurred over long periods, the nodules coalesced into a horizon within the lower B-zone of the soil profile. The vertisol textures and calcic nodules that indicate seasonal climatic conditions are consistent with paleogeographic reconstructions that place the central Appalachian basin within the southerly monsoonal belt (5-10 degress south of the equator; Cecil and Englund, 1989; Golonka et al., 1994). Grey-green horizons that cap some of the paleosol profiles may reflect organic matter or impeded drainage through the A horizon (cf. Bown and Kraus, 1981). In most cases, the gray-green color represents a reduced zone in the B horizon beneath fossiliferous, dark shale units.

The stacked character of the paleosols with respect to fluvial sandstone bodies and channel fills is suggestive of repetitive episodes of soil development. The thick, mature paleosols within the redbed successions are interpreted to mark prolonged episodes of pedogenesis in distal floodplain settings (cf. Kraus and Bown, 1988; Wright and Robinson, 1988). Such units are candidates for defining times of relative lowstands in the mudstone-dominated parts of the study interval. Thinner (<1m) paleosols that lack distinct horizons reflect development near channels where frequent overbank sedimentation covered developing profiles (cf. Kraus and Bown, 1988; Retallack, 1988). The immaturity of these soils suggests that development was a function of intrabasinal controls on sedimentation such as avulsion frequency (Wright and Robinson, 1988).

The laterally continuous, bedded carbonates are interpreted as coastal lake or playa deposits. These sparsely fossiliferous (ostracodes, bivalves) units presumably define times during which local areas on the floodplain was shallowly inundated, and did not receive significant riverine clastic influx (cf. Platt and Wright, 1991; 1992). Some of these playa carbonates were subesequently altered by pedogenic processes during development of overlying vertic/ calcic paleosols.

Collectively, the association of floodplain/ coastal plain facies suggests that the paleoshoreline was basinward of the study area during much of the depositional history. The predominance of laterally persistent paleosols in parts of the study interval implicates prolonged episodes of soil development upon areally expansive floodplains. Although thin interbeds of marine shales formed during short-lived flooding events, the thickness of terrestrial facies suggests net aggradation when sediment influx was balanced by accommodation.

SEQUENCES

Stratigraphic stacking of the depositional systems defines at least 17 unconformity-bounded, transgressive-regressive sequences in the eastern part of the study area (Fig. 1.5). These sequences converge and amalgamate into the distal part of the basin where ~12 sequences are recognized in the westernmost well logs (Fig. 1.6). Five principal sequence types are distinguished on the basis of the dominant depositional system (Fig. 1.23). Each type is interpreted within a sequence stratigraphic context that relates depositional and erosional episodes to base level change.
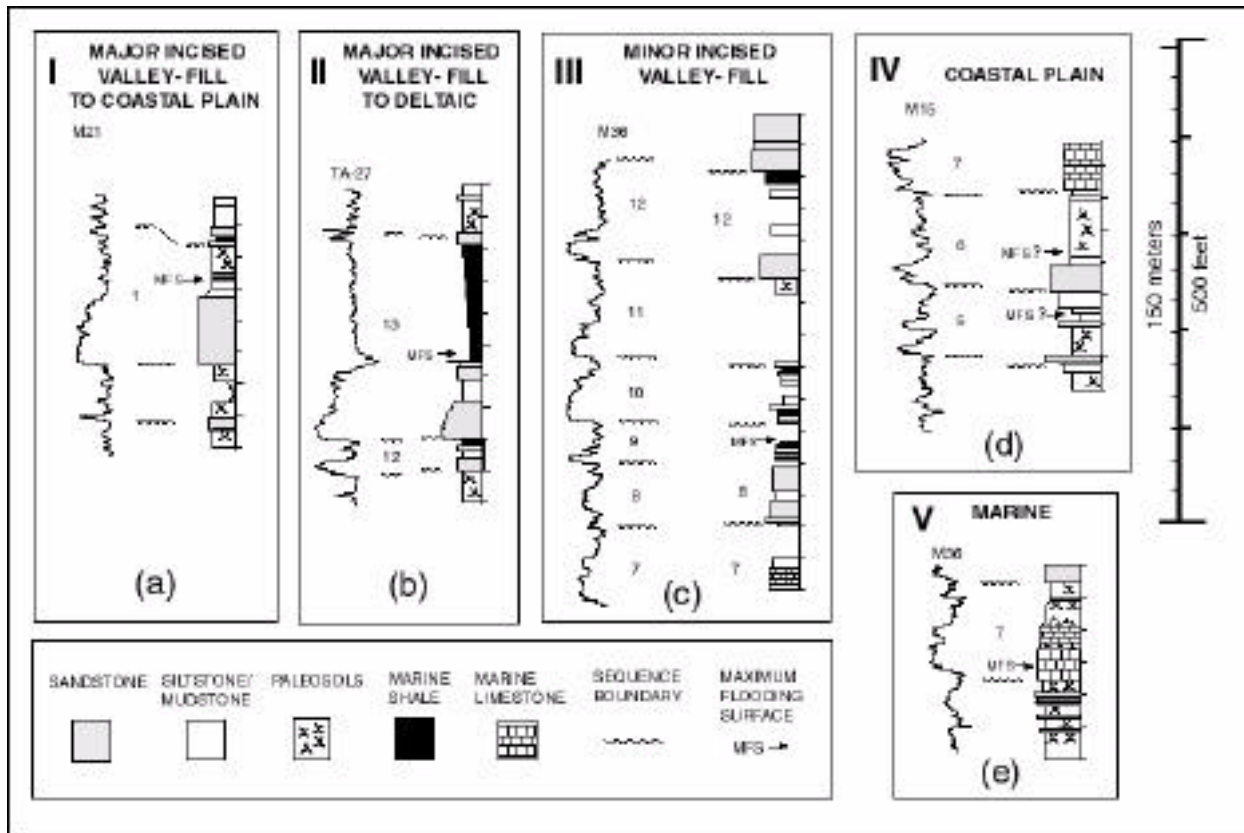
Figure 1.23- Types of fourth-order sequences recognized within the study interval. Lower sequence boundaries of Types I, II, and III define times of valley incision during relative lowstand. Bases of channelized fluvial sandstones (Type IV) are interpreted as the updip equivalent of incised sequence boundaries that developed downdip in the less rapidly subsiding part of the basin (cf. Fig. 1.3). Paleosols define lower sequence boundaries of marine-dominated sequences (Type V).

Major Valley Fill To Coastal Plain-Dominated Sequence

Sequence 1 is comprised of a thick incised valley-fill succession (the Stony Gap Member), a thin (few meters) dark, fossiliferous shale unit, and an uppermost coastal plain succession dominated by paleosols (Figs. 1.5, 1.6, 1.23a). The regional basal unconformity defines a time of significant fall in relative sea level when fluvial incision occurred throughout much of the basin. The thickness of the overlying incised valley-fill facies (up to 50 m) indicates that most of the accommodation space generated during the subsequent relative sea level rise was infilled by braided fluvial and tidal-estuarine facies that represent transgressive deposits within

the extensive paleovalley. In easternmost outcrops, upper estuarine inclined heterolithic facies are overlain by a thin (< 5 m) fossiliferous shale that is interpreted as an inner estuarine or nearshore marine deposit. This shale defines the maximum flooding surface near the landward (eastward) limit of transgression. Upward transition into rooted horizons and coastal plain facies (redbeds and paleosols) marks progradation and aggradation of coastal plain facies during highstand as relative sea level began to fall. An erosional basal contact with overlying channelized sandstone marks the upper sequence boundary.

Major Incised Valley-Fill To Delta-Dominated Sequence

Sequence 13 is comprised of a conglomeratic valley-fill succession (the Princeton Formation) and marine to deltaic units of the Pride Shale and Glady Fork Sandstone Member of the Bluestone Formation (Figs. 1.5, 1.6, 1.23b). This thick (up to 70 m) sequence overlies a regional unconformity that truncates progressively lower units in the Hinton Formation into distal parts of the basin. Basal braided fluvial and tidal estuarine incised valley-fill deposits represent transgressive deposits, and are sharply truncated by a ravinement bed that developed as the valley was transgressed. The dark marine shale horizon at the base of the Pride Shale represents a condensed section that defines maximum flooding. The anomalous thickness (up to 60 m) of overlying marine/ prodeltaic facies suggests that relative sea level rise provided as much as 100 m of accommodation in the eastern part of the basin, assuming 40 m water depths. Highstand deposition is represented by progradational tide-dominated deltaic facies. The upper sequence boundary is marked by rooted horizons at the top of ripple-bedded, distributary mouthbar facies. This rooted horizon is correlative with deeply-incised fluvial channels that were cut locally into the prodeltaic facies during lowstand (cf. Fig. 1.18).

Minor Incised Valley-Fill Dominated Sequences

Sequences 6, 8-12, and 14 have an incised lower sequence boundary that truncates underlying marine or terrestrial facies (Figs. 1.5, 1.6, 1.23c). Overlying intervals are represented rarely by fining-upward conglomeratic fluvial deposits (sequence 10 locally; and sequence 14) or, more commonly, by mixed wave- to tidal-estuarine deposits (Fig. 1.23c). These facies infilled shallow incised paleovalleys during a relative sea level rise. Maximum flooding surfaces are defined by dark, fossiliferous, estuarine to shallow-marine shales that overlie the incised, valley-fill units. Highstand deposits of green gray siltstones and rooted coastal plain facies record coastal progradation as sediment supply outpaced accommodation.

Coastal Plain-Dominated Sequences

Sequences 2-5 and 15 are dominated by coastal plain facies (Figs. 1.5, 1.6). Sequence boundaries are defined locally by fluvial channels that are incised locally into underlying paleosols (Fig. 1.23d). In some cases, these erosional horizons are interpreted as the updip equivalent of deeply-incised sequence boundaries that developed downdip in the less rapidly subsiding part of the basin (cf. sequence 6 in Figs. 1.3, 1.5, 1.6). Aggradational channel fills and overbank deposits of fine-grained, meandering fluvial systems record increased accommodation. Maximum flooding surfaces are defined by thin (typically < 1 m) dark shales that likely reflect short intervals of marine incursion over a low relief coastal plain. Highstand deposition is marked by paleosols and intercalated non-channelized sandstones that define times of floodplain aggradation.

Signatures in the westernmost well logs indicate that the coastal-plain facies were deposited throughout the study area as part of an arcuate apron of sediment in the western, northern, and eastern parts of the basin. Coeval marine facies to these sequences are present to

the southwest in Black Warrior basin and the Ouachita trough (Donaldson and Schumaker, 1981; Hatcher et al., 1989).

Marine-dominated sequences

Sequences 7 and 17 are comprised of fossiliferous marine units that are bounded by sequence boundaries defined by thick paleosols (Figs. 1.5, 1.6 and 1.23e). The lower sequence boundaries were not incised prior to regional inundation. Marine fossils immediately above the sharp base of the Little Stone Gap Member (Hinton Formation) (Sequence 7) suggest that coastal inundation was rapid. In contrast, the Bramwell Member (Bluestone) Formation (Sequence 17) overlies a few to 30 cm-thick carbonaceous shale that likely represents a transgressive, swamp deposit. The thickness (tens of meters) and areal extent of the marine-dominated sequences is suggestive of extended periods of inundation. Late highstand deposits include calcareous shales. These upper shale units in both sequences have been pedogenically altered into vertic or calcareous paleosols in association with development of the upper sequence boundary.

CONTROLS ON SEQUENCE DEVELOPMENT

Given that the thick eastern section in the study area is the most complete record of Late Mississippian sedimentation in the Appalachian Basin (Arkle, 1974), a minimum average sequence duration can be calculated by dividing the maximum number of sequences (17) into the ~7 million years total time estimated to be represented by the study interval. The resulting figure of ~412 ka is an approximation because: 1) overall time constraints are based upon limited biostratigraphical data and indirect geochronological controls, and 2) the calculation assumes preservation of all Late Mississippian sequences. Although the study interval has been regarded as part of a continuous Mississippian-Pennsylvanian succession (Englund and Randall, 1981), recent work on stratigraphic relationships in the subsurface (Figs. 1.3, 1.5, 1.6; and Beuthin, 1994) and lithologic evidence from the outcrop belt (Wilson, 1984; Beuthin, 1997) are suggestive of a depositional hiatus at the top of the study interval. Although the existence of a lengthy (1-4 million year) Mississippian-Pennsylvanian hiatus does not directly affect the above calculation, the presence of such a long-term depositional hiatus highlights the uncertainties regarding the preservational completeness of Upper Mississippian strata in the Appalachians (cf. Sutherland and Manger, 1984). Nondeposition or erosional loss of section during Late Mississippian regressional episodes would contribute to an overestimated average duration for the preserved sequences. Nevertheless, sequence development appears to reflect recurring changes in base-level changes over 300-500 k.y. Such cyclicity corresponds with fourth-order cycle within the hierarchy of Mitchum and VanWagoner (1991) and Vail et al. (1991).

Earlier studies of similar high-frequency stratigraphic cycles in the Appalachian basin and neighboring basins have invoked delta-switching (Ferm, 1970), eustasy (Chesnut, 1994; Van Tassell, 1994, Filer, 1994, Aitken and Flint, 1995; Smith, 1996; Al-Tawil, 1998), regional climatic change (Cecil, 1990), and tectonically induced subsidence (Pashin, 1994) as primary forcing mechanisms on sequence development. In the present study, the stacked character of the sequences is suggestive of recurring changes in relative sea level. The lateral (along strike) and downdip continuity of sequence boundaries and marine units (Figs. 1.5, 1.6) is suggestive of basinwide accommodation change forced by a regional or global mechanism rather than by autogenic effects such as delta-switching. Comparison of the interpreted relative sea level curve (Fig. 1.24) with that proposed by Swann (1964) for the midcontinental Chesterian type section shows that erosional and depositional episodes may have been synchronous in the Appalachian and Illinois basins. Although accumulation of these strata coincided with a time of increasing tectonic activity, it seems unlikely that short-term (< 0.5 m.y.) tectonically-induced changes in base level forced the development of cyclothem-scale stratigraphic packaging in such widely separated basins (cf. Klein and Kupperman, 1992).

In the present study, sequence development is interpreted to reflect a glacioeustatic control because glacial-interglacial cycles can account for both the frequency and magnitude of the base level changes. The calculated duration of the Upper Mississippian sequences in this study compares well with the 413 k.y. Milankovitch long-eccentricity period that is regarded as a primary forcing mechanism for late Paleozoic climate change and attendant glacial episodes (Crowell, 1978; Algeo and Wilkinson, 1988). This long-eccentricity period is thought to have remained stable through the past 600 m.y. (Walker and Zahnle, 1986). Algeo and Wilkinson (1988) note a positive clustering about the 413 k.y. cycle duration in the late Mississippian through upper Pennsylvanian stratigraphic record based upon an analysis of cycle durations reported for 200 stratigraphic successions. Similar sequence durations have been recognized in the Pennsylvanian stratigraphic record in the midcontinent (Heckel, 1986; Boardman and Heckel, 1989) and western United States (Soreghan, 1994a;1994b). In many midcontinental cyclothemic successions, a eustatic control on sequence development is strongly suggested by the fact that some cyclothems can be correlated over hundreds of kilometers and between widely separated basins (Connolly and Stanton, 1992). In the present study, direct correlation with equivalent sequences in the midcontinent is limited because several of the Appalachian sequences are confined to the eastern part of the basin where the subsidence rate was greatest. Nevertheless, the fourth-order (~400 k.y.) cyclicity identified herein is consistent with cyclicity identified by Al-Tawil (1998) within Mid-Mississippian carbonates in the Appalachian basin, and by Wilson (1984) and Aitken and Flint (1995) within the Pennsylvanian rocks of eastern Kentucky.

The incised lower boundaries and the overall thicknesses of the Upper Misisippian sequences suggests that the magnitude of relative sea-level change was typically tens of meters. In regards to the thickest sequences (1 and 13), decompacted thicknesses are suggestive of sea level rises of up to 100 m. Crowley and Baum (1991) estimate that the magnitude of eustatic change with respect to Late Paleozoic glacial-interglacial cycles was 60 m $\pm$ 15 m (cf. Maynard and Leeder, 1992). Significant glacioeustatic changes likely occurred during the Namurian when the Gondwanan ice cover may have been at its greatest areal extent (21 x 10^6 km^2 ; roughly equivalent to the Pleistocene; Gonzalez-Bonorino and Eyles, 1995).
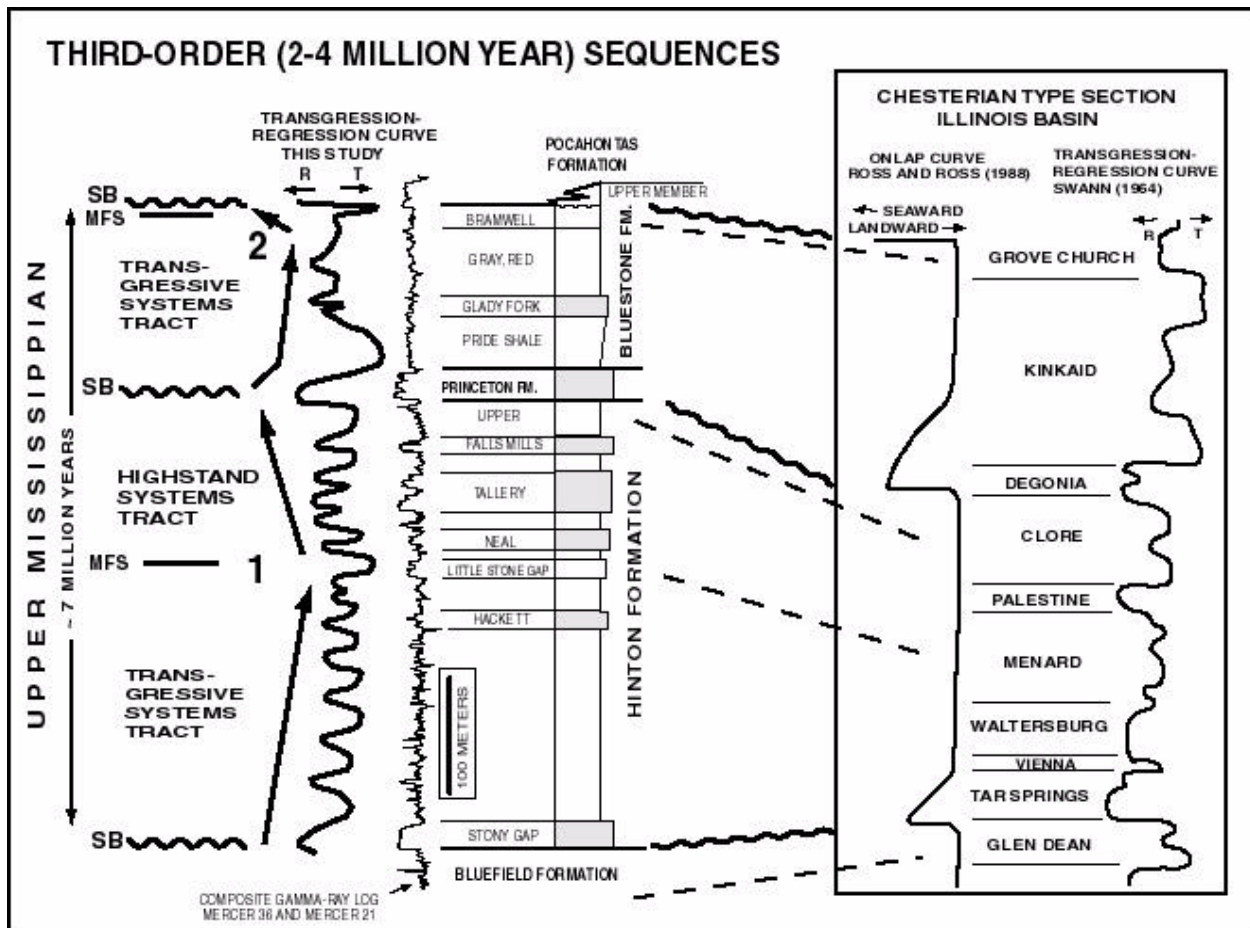
Figure 1.24- Interpreted long-term (third-order) cyclicity defined by major unconformities and stacking patterns of high-frequency (fourth-order) sequences into sequence sets. Third-order transgressive systems tracts are represented by aggradational sequence sets whereas the third-order highstand systems tract is comprised of a progradational sequence set. Interpreted sea-level curve, is compared with the onlap curve of Ross and Ross (1988) and the sea level curve of Swann (1964) for equivalent-age stratigraphic units in the Illinois basin.

COMPOSITE SEQUENCES

The study interval can be subdivided into two thick (100's of meters) stratal packages that represent third-order (0.5-5 m.y.) sequences of Vail et al. (1991). These third-order sequences represent composite sequences (Mitchum and Van Wagoner, 1991; cf. Weber et al., 1995) that are bounded by regional unconformities, and display internal stacking of component fourth-order sequences into retrogradational-aggradational and progradational sequence sets. These sequence sets are regarded, respectively, as third-order transgressive and highstand systems tracts (Fig. 1.24). The composite sequences are notably asymmetric in that the retrogradational-aggradational sequence sets comprise ~60% and nearly 100% of the total respective thicknesses (Fig. 1.24).

The lower composite sequence (representing the Hinton Formation) is comprised of a retrogradational, major incised valley-fill to coastal plain sequence (fourth-order sequence 1); a thick (~250 m) aggradational sequence set of coastal plain-dominated sequences (fourth-order sequences 2-6); a regional marine-dominated sequence (fourth-order sequence 7); and a

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progradational sequence set that is dominated by minor incised valley-fill deposits (fourth-order sequences 8-12).

The transgressive systems tract of the lower composite sequence is dominated by coastal plain sequences that define episodes of fourth-order base-level change during a time when the shoreline was several to tens of of kilometers west of the present day outcrop belt. These strata accumulated in the proximal part of the foreland basin where subaerial accommodation was greatest as a result of tectonically-induced subsidence (cf. Schwans, 1995). In such settings, dip-parallel fluvial profiles undergo long-term backward rotation that favors aggradation in the upstream part of the fluvial system as streams strive to maintain their equilibrium profile (Posamentier and Allen, 1993). Fourth-order base level changes during such times were manifested by attenuated incision during fourth-order eustatic falls, increased fluvial aggradation during initial transgression, and short-lived episodes of coastal inundation during rapid sea-level rises. The predominance of terrestrial facies suggests that fluvial sediment influx was sufficient to maintain aggradation of the coastal plain during overall relative sea level rise.

The marine-dominated sequence (7) that contains the fossiliferous Little Stone Gap Member marks regional inundation of the coastal plain, and represents the third-order maximum flooding surface. This unit contains fossils that indicate open-water marine conditions (Gordon and Henry, 1981; Hoare, 1993) and seemingly marks a time when basin configuration, anomalous relative sea-level rise, and/ or climatic factors limited clastic input to the central Appalachian basin.

The overlying progradational sequence set is comprised of a stacked series of fourth-order sequences (8 to 12) that record alternating terrestrial, estuarine, and marine deposition during third-order highstand. Fourth-order sequences are marked by fluvial/ estuarine systems in minor paleovalleys that record incision and backfill, and fossiliferous marine shales and thin (1-3 m) marine limestones that mark flooding events. The limited thickness of coastal plain deposits between the sequences suggests that punctuated erosional episodes took place during a time of reduced overall accommodation. This relationship is best displayed by the section in the western part of the basin where subsurface data indicate that fourth-order, incised-valley sandstones are stacked with little to no intervening coastal plain facies.

The base of the upper composite sequence is defined by the incised boundary at the base of the Princeton Formation, that marks a lowstand surface of erosion throughout the basin. The anomalous 60-m thickness of the overlying marine/ prodeltaic Pride Shale Member indicates that significant inundation occurred during initial development of the retrogradational-aggradational sequence set. Overlying units that comprise the remainder of this transgressive systems tract are dominated by coastal plain deposits. The aggradational sequence set of the upper package is overlain by a marine-dominated sequence (Bramwell Member; sequence 17) that defines third-order maximum flooding and likely represents the younger counterpart of the Little Stone Gap Limestone. No progradational (highstand) sequence set is recognized in the upper package. Instead, this stratigraphic position is occupied by a well-developed paleosol atop the Bramwell Member that likely marks an extended, pre-Pennsylvanian depositional hiatus (Beuthin, 1997) at the culmination of second-order eustatic fall (Mid-Carboniferous eustatic lowstand). This possibility is corroborated by paleontological evidence (Pfefferkorn and Gillespie, 1981; Saunders and Ramsbottom, 1986) of a significant time break between the Mississippian and Pennsylvanian systems in the Appalachian basin.

ORIGIN OF THE COMPOSITE SEQUENCES

The wedge-like geometry of the Upper Mississippian section (Fig. 1.3) indicates that relatively uniform asymmetric accommodation occurred throughout the ~7 Ma depositional history. Because deposition of these strata was coincident with second-order eustatic fall (Vail et al., 1977; Saunders and Ramsbottom, 1986), long-term accommodation was generated by thrust loading and sediment loading in the active foreland region.

Based on existing time constraints (Fig. 1.4), the composite sequences identified in this study each record 2 to 4 million years. The retrogradational-aggradational to progradational stratigraphic stacking patterns of fourth-order sequences within the composite sequences defines long-term changes in accommodation. Similar cyclicity is noted for the Devonian (Van Tassel, 1994; Filer, 1994), middle Mississippian (Al-Tawil, 1998), and Pennsylvanian (Busch and Rollins, 1984; Aitken and Flint, 1995) stratigraphic record in the Appalachian Basin. The two composite sequences also correspond with the 1.2 to 4.0 m.y., transgressive-regressive sequences identified in the Illinois Basin by Ross and Ross (1988)(Fig. 1.24). The origin of third-order composite sequences is controversial (Plint et al., 1992; Kendall et al., 1995).

Late Mississippian tectonism in Appalachians was characterized by thrust-loading episodes in the deformational belt (Tankard, 1986; Quinlan and Beaumont, 1984; Ettensohn, 1994). It is possible that the two third-order sequences recognized in the study interval may represent tectono-stratigraphic units linked to a discrete thrust event (cf. Kauffmann, 1984; Leckie, 1986). This interpretation is consistent with Flemings and Jordan's (1990) model that shows that alluvial deposits aggrade in the proximal part of foreland basins during times of thrust-induced subsidence, whereas progradation dominates during times of tectonic quiescence. However, the fact that third-order sequences are expressed in both the Appalachian and Illinois basins (Fig. 1.24) requires far-field tectonic effects to have influenced accommodation and thereby long-term sedimentary stacking in both foreland and intracratonic settings. Quinlan and Beaumont (1984) suggest that synchronous subsidence in these basins relates to flexural interaction across the Cincinnati arch in response to the episodic emplacement of thrust loads.

Alternatively, a global-scale control on sea level could account for the formation of the composite sequences. A global mechanism is supported by the observation of Ross and Ross (1988) that third-order cycles identified in midcontinental North America can be recognized worldwide. Therefore, long-term glacioeustatic (cf. Olsen and Kent, 1996) or tectono-eustatic control on sea level may provide a better explanation than can regional tectonism related to thrust episodes. The lack of conclusive evidence for multi-million year glacioeustatic cycles has led to the suggestion that global tectonic mechanisms such as variation in the rate of sea-floor spreading (cf. Kauffman, 1984; Harrison, 1990) or changes in intraplate stressfields (Cloetingh, 1988) may control third-order eustatic change.

CONCLUSIONS

Upper Mississippian strata in the central Appalachian basin preserve a cyclic record of deposition in fluvial, estuarine, marine, and deltaic environments. Interpretation of these units within a sequence stratigraphic context suggests that transgressive-regressive packages record cyclothem scale (~400 ka) base-level fluctuations that permitted incision and backfill of paleovalleys along the eastern margin of the basin. Sequence development likely reflects glacioeustatic cycles related to coeval Gondwanan glacial-interglacial episodes. Glacioeustasy explains the record of high frequency, high amplitude (20-60+ m) base level change better than punctuated tectonic episodes or cyclic changes in local sediment supply related to climatic variation.

The strata in the study interval were deposited over ~7 million year period that coincided with second-order sea level fall. Long-term accommodation was provided by tectonic loading along the eastern margin of the Appalachian basin. Two third-order (2-4 Ma) composite sequences manifested by bundling of 4th-order sequences reflect variation between times of relatively high and low accommodation. The third-order sequences are notably asymmetric with thick (100's of m) packages of retrogradational, major incised-valley fills and aggradational coastal plain mudstones comprising the transgressive systems tracts, and alternating estuarine and subordinate terrestrial facies comprising comparatively thin highstand systems tracts. The older of the two third-order sequences is more complete, whereas the upper sequence was truncated by a prolonged erosional event. These composite sequences are comparable to similar long-term sequences known from the Appalachian and Illinois basins, and are thought to reflect global glacio- or tectono-eustasy.

Chapter 2:
Late Mississippian Prodeltaic Rhythmites in the Appalachian Basin: A Hierarchical Record of Tidal and Climatic Periodicities

ABSTRACT

The Pride Shale is a 60 meter-thick succession of thinly interlaminated dark shales, siltstones, and fine-grained sandstones within the Bluestone Formation (Upper Mississippian) of southern West Virginia. A hierarchy of submillimeter-to meter-scale cycles in this succession preserves a spectrum of tidal and climatic periodicities. Submillimeter-thick, fine-grained sandstone/ shale or siltstone/ shale couplets are interpreted as the product of suspension settling associated with individual ebb-tidal flows. Up to 17 couplets systematically thicken and thin within millimeter-to centimeter-scale bundles interpreted as neap-spring tidal cycles. Successive neap-spring cycles display a thick-thin relationship interpreted to reflect unequal perigean and apogean tides. The fine scale and abbreviated character of these microlaminated rhythmites is suggestive of a distal, subtidal setting wherein subordinate daily and neap and apogean ebb flows were generally of insufficient strength to transport sand and silt. Decimeter-scale cycles in the Pride Shale are manifested by the progressive upward thickening and thinning of up to 18 millimeter-to centimeter-thick (neap-spring) cycles. This bundling is interpreted to reflect a climatic (monsoonal) signal which defines that part of the year when adequate fluvial-tidal discharge forced suspended sediment into the basin. Lastly, a multi-year cyclicity is recognized in thick exposures where 17-22 annual beds display a crude upward thickening and thinning within meter-scale bundles. This bundling is interpreted to reflect the 18.6-year lunar nodal cycle, with the thick annual beds representing years during which the inclination of the lunar orbital plane favored increased tidal amplitudes.

Decompacted thicknesses of cycles indicate that accumulation rates for the Pride Shale typically ranged between 3 and 20 cm per year, but reached over 60 cm per year where sandy rhythmites developed as the marginal infill of large slump scars. The stratigraphic relationships and sedimentation rates for the Pride Shale are consistent with paleontological evidence of rapid sedimentation in turbid waters seaward of multiple points of fluvial discharge. The Pride Shale is interpreted as an extensive prodeltaic deposit which, in places, records hundreds of years of nearly continuous tidal sedimentation along the eastern margin of the Appalachian Basin.

INTRODUCTION

Tidal cycles preserved within flat-laminated, fine-grained sediments are well documented from Upper Proterozoic and Carboniferous successions in Australia (Williams, 1989a) and North America (Kuecher et al., 1990; Kvale and Archer, 1991; Archer et al., 1994; Chan et al., 1994). These tidal rhythmites are high-resolution depositional records from which tidal-forcing mechanisms and absolute rates of sedimentation can be deduced. Ancient tidal rhythmites also have been used to demonstrate changes in the Earth-moon system through time (Sonett et al., 1988; Williams, 1989b; Sonett et al., 1996).

Numerous tidal rhythmite sequences within the Carboniferous strata of the mid-continental U.S.A. highlight the significance of tidal sedimentation in intracratonic settings. Recent studies (Kvale and Archer, 1989; 1990; Kvale et al., 1989; Kvale et al., 1994; Martino and Sanderson, 1993; Greb and Archer, 1995) focusing primarily on localized sequences of estuarine rhythmites show that these deposits reflect rapid sedimentation (up to meters per year) and may record up to tens of years of continuous deposition (Kvale and Archer, 1990; Feldman et al., 1993).

Recent analysis of hierarchical laminae bundling in the Pride Shale of southern West Virginia reveals a unique tidal rhythmite deposit within the Upper Mississippian strata of the central Appalachian Basin. Extending into southwestern Virginia and eastern Kentucky, the Pride Shale is distinguished by its considerable thickness, unusual weathering pattern, and distinctive gamma-ray signature. The fine laminae within the Pride Shale have been interpreted as the result of low-energy tidal deposition (Cecil and Englund, 1989) or seasonal oxygen-level fluctuations (Neal, 1994), but conclusive evidence regarding the origin of the strikingly regular and cyclic bedding has been lacking. Detailed analyses of the Pride Shale at the thin-section and outcrop scale indicate that the cyclicity is commensurable with a hierarchy of tidal and climatic periodicities. Herein we discuss the nature of these depositional cycles and present further evidence which suggests that the Pride Shale records hundreds of years of tidal rhythmite deposition in a prodeltaic environment. As such, the Pride Shale differs from most known tidal rhythmites in its depositional setting and in its unusually long record of sedimentation.
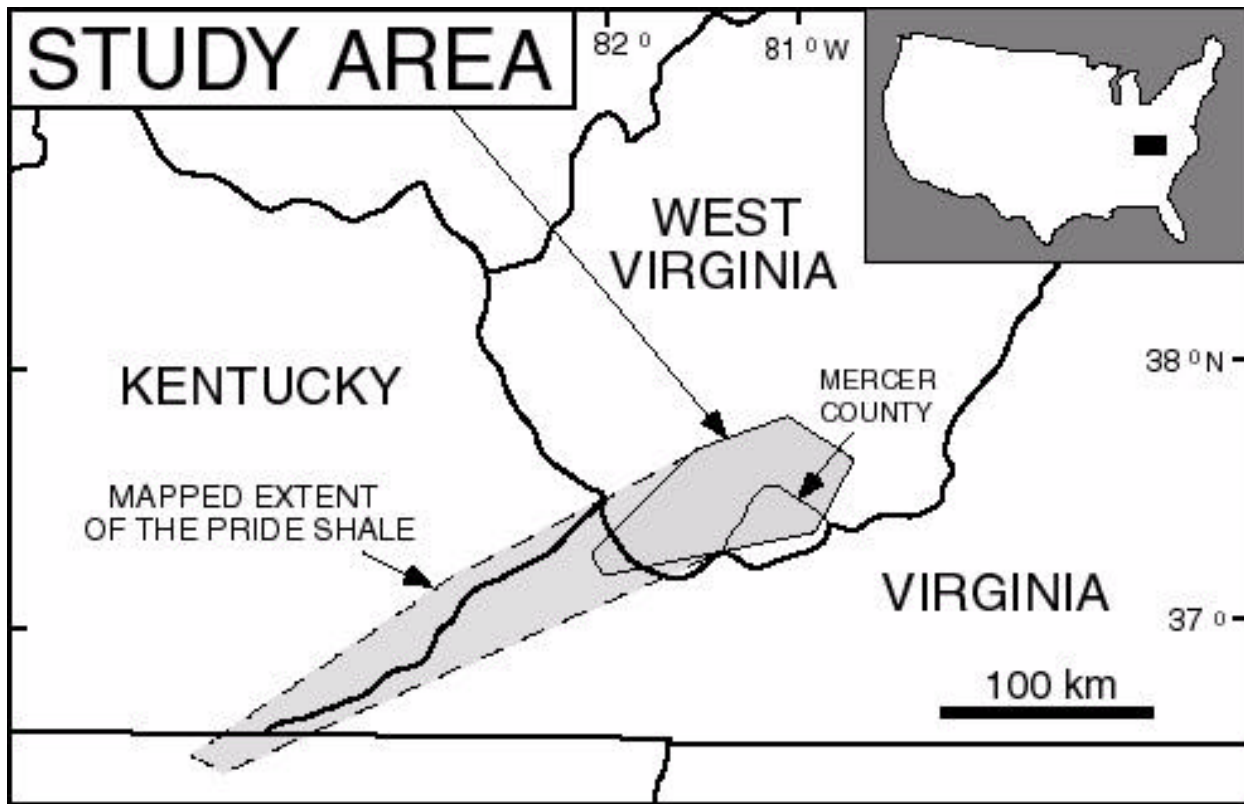

Figure 2.1- Map showing the regional mapped extent of the Pride Shale and the study area in southern West Virginia and southwestern Virginia. The Pride Shale crops out in the eastern part of study area, and is particularly well exposed in the central part of Mercer County.

METHOD OF STUDY

This study is based on field investigations of the Pride Shale and its bounding units in southern West Virginia and southwestern Virginia (Fig. 2.1). The Pride Shale is thickest and most rhythmic in the eastern outcrop belt, where differential weathering of sand-dominated vs. clay-dominated layers imparts a washboard-like surface to vertical outcrop faces. Because of variation in grain size and degree of weathering, the rhythmic character of the Pride Shale is rarely uniform throughout single exposures. Nevertheless, a similar hierarchy of laminae

bundling is recognized in both thinly laminated and thickly laminated Pride Shale throughout most of the study area. Rhythmicity was noted in over 14 roadcuts in the central part of Mercer County (West Virginia), yet only the outcrops that best displayed the millimeter- to meter-scale bundling were selected for detailed analysis. These exposures include a long roadcut along Interstate 77 immediately northwest of Exit 20 at Camp Creek, West Virginia (see Englund 1989), and a borrow pit/ roadcut along Route 19 at Spanishburg, West Virginia. Data collected from the lower part of the Pride Shale (approximately 5 m above the base) included measurement of over 130 sub-millimeter-thick layers in thin section. The thicknesses of the fining-upward microlaminae were measured with a calibrated occular micrometer from the base of each recognizable sub-millimeter-thick sandstone/ siltstone layer. The thicknesses of over 100 centimeter-scale cycles and of 200 decimeter-scale cycles from the middle of the Pride Shale interval were measured with a digital micrometer on polished slabs or on the outcrop from the midpoints of the clay-dominated intervals. Fast Fourier Transform and Maximum Entropy spectral analyses of the cycle sequences were used to quantify the number of components within the decimeter and meter-scale bundles. Spectral analyses of components in the microlaminated cycles ($<$ 1 cm thick) were not performed because: (1) the poor resolution of individual clay-dominated couplets precluded an accurate assessment of the maximum number of components within most cycles, and (2) the vertical succession of centimeter-scale cycles that contain a relatively high number of internal components is limited by decimeter-scale thickening and thinning. In any given section, the few cycles that appear most complete are invariably bounded by abbreviated, clay-dominated cycles whose components are thin, incomplete, or otherwise too diffuse to be measured.

STRATIGRAPHIC AND GEOLOGIC SETTING

The Pride Shale is exposed in a northeast-southwest-trending belt along the eastern margin of the Allegheny Plateau, and extends westward into the subsurface over approximately 12,000 km^2 in West Virginia, Virginia, Kentucky, and Tennessee (Fig. 2.1). In West Virginia, the Pride Shale is the basal member of the Bluestone Formation, the youngest of four formations assigned to the Upper Mississippian Mauch Chunk Group (Fig. 2.2). These Mauch Chunk strata, including the basal Bluefield, Hinton, and Princeton Formations, represent a thick (up to 1000 m) succession of predominantly siliciclastic sediments that were deposited in various terrestrial and shallow-marine settings along the eastern margin of the central Appalachian Basin (Englund and Thomas 1990). Vertic and calcic paleosols within the red mudstones of the Hinton and Bluestone Formations are suggestive of moderately to highly seasonal (semiarid) climatic conditions (Cecil and Englund, 1989; Cecil, personal communication 1996). Recent studies (Caudill et al., 1996) of equivalent-aged paleo-vertisols in Tennessee have shown the the Late Mississippian climate in the southern part of the Appalachian basin was semiarid with a mean annual precipitation comparable to that in extreme southern Texas.
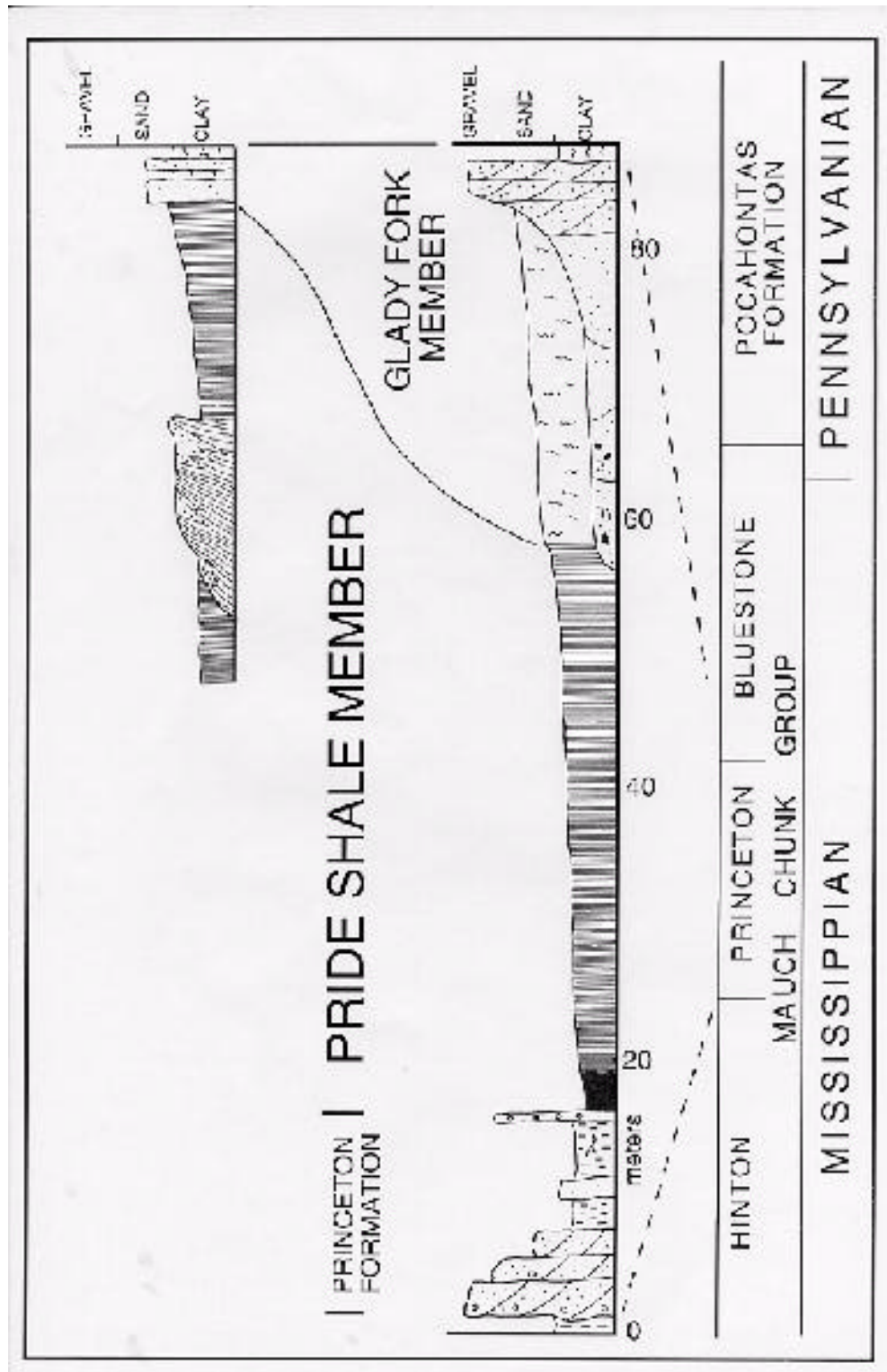
Figure 2.2- Stratigraphic column of the study interval in Mercer County, West Virginia. The Pride Shale grades upward from a basal condensed section into the wavy-bedded, fine-grained sandstone facies of the Glady Fork Sandstone member throughout most of the study area.

In southern West Virginia, the Pride Shale is sandwiched between two conglomeratic sandstones: the underlying Princeton Formation and the overlying Glady Fork Sandstone Member (Fig. 2.2). The Princeton Formation is a regional conglomeratic unit that unconformably overlies gray to red mudstones of the upper Hinton Formation. Polymictic conglomerates and cross-bedded sandstones of the lower Princeton Formation are overlain by a laterally variable succession of rooted sideritic mudstones, tabular sandstone bodies, thin coals, and leached paleosols. Also within the upper Princeton interval are heterolithic rhythmites that fill small (< 1 m depth) channels, and display laminae bundling similar to that as described below in the Pride Shale. The Princeton-Pride contact is marked locally by a thin (to 0.5 m), quartz pebble conglomerate that truncates the underlying rooted facies and contains a variety of gastropods, bivalves, and brachiopods. This conglomerate has been interpreted as formed during a widespread marine-transgressive event (Weems and Windolph, 1986), and likely represents a lag deposit developed upon a tidal ravinement surface (cf. Dalrymple, 1992). Where this conglomerate is absent, the distinction between the uppermost Princeton Formation and the basal Pride Shale Member is difficult in weathered outcrops (cf. Englund, 1989). We consider that the base of the Pride Shale Member is marked by the few-meter-thick, black, fissile shale that overlies the conglomerate/ fossil lag (or rests upon older Princeton or Hinton facies) and commonly encloses carbonate concretions bearing marine invertebrate (foraminifera and bivalves) and vertebrate (fish) fossils (Weems and Windolph, 1986). High gamma ray readings obtained from a scintillometer on the outcrop suggest that this unit is the source of the distinct gamma ray spike seen in subsurface well logs. This "hot shale" marker unit is recognized throughout the study area (in both outcrop and subsurface sections), and is interpreted as a condensed section developed during regional inundation. The rest of the Pride Shale is composed of rhythmically interlaminated dark shales and fine-grained sandstones in which plant fossils (*Stigmaria Stellata*, and unidentifiable disseminated material) and small, thin-valved bivalves (*Sanguinolites, Modiolus* sp.) are locally abundant. Scattered specimens of a shrimp-like arthropod are preserved as delicate carbonized impressions high up in the Pride Shale. In general, the thin laminae are laterally continuous and have not been disrupted by bioturbation, though small vertical escape structures are common in the sandy rhythmite facies. The largest and most freshly exposed outcrops are characterized by thick (20+ m) intervals of rhythmic beds whose gentle dip (a few degrees greater than the regional dip) and grain-size trends are suggestive of westward-prograding clinoforms.

Prominent discontinuities within the Pride Shale occur as smooth to irregular, concave-upward surfaces that extend in roadcuts for hundreds of meters. The axes of these trough-like features strike SW-NE, and are characterized locally by meter-scale rotated/ deformed blocks. The rhythmic infill lacks a sandy or conglomeratic basal lag, yet drapes the blocks and is coarsest along trough shoulders where sand-dominated rhythmites pinch out into finer-grained beds of the trough axes. A slump-block origin (Cooper, 1961) and a channel-scour origin (Englund, 1989) proposed for the discontinuities do not explain all aspects of their geometry and sedimentary fill. An alternative interpretation of the discontinuity-bounded sedimentary packages is that they may represent infilled slump scars (cf. Davies, 1977). Such a scar-infill interpretation invokes subaqueous gravity sliding of thick (to 15 m) packages of semicoherent sediment into deeper parts of the basin. Meter-scale blocks derived from the trailing edge of the slump block were left in the scar interior prior to progressive infill by rhythmites. The irregular distribution of sand-dominated rhythmites may reflect cannibalization of sands from disrupted sediments along the updip scar surface and redistribution by bottom currents moving oblique to the slump direction.

In most of the study area, the Pride Shale grades upward into a wavy-bedded, fine-grained sandstone unit (mapped as the Glady Fork Member), 3-20 m thick. The distinction between the Pride Shale and this wavy-bedded facies is recognized in outcrop by weathering profile, and in well logs by the shift toward lower gamma ray/ bulk density values. This facies is crudely rhythmic, with mud flasers and drapes between ripple-bedded or flat-laminated fine-grained sandstones. Scattered sand-filled vertical burrows (some containing fecal pellets) and disseminated plant material characterize this facies in some parts of the study area. Englund and

Thomas (1990) suggested a deltaic environment of deposition for this facies. Uppermost rooted horizons of this unit appear correlative with a brecciated paleosol that overlies the Pride Shale in the northernmost part of the study area. The Glady Fork Member in its type area is a coarse conglomerate and cross-bedded sandstone unit that lies unconformably upon the Pride Shale within a 1 km wide, southwestward-trending paleovalley.

The Pride Shale Member had previously been interpreted as a shallow-marine, lagoonal, or estuarine deposit (Miller, 1974; Englund, 1989). On the basis of the stratigraphic relationships and sedimentological features discussed above, the Pride Shale is interpreted as a marine to prodeltaic unit whose basal condensed section rests upon truncated fluvial to tidal-estuarine facies of the underlying Princeton Formation. A prodeltaic interpretation of the rhythmic-bedded Pride Shale is consistent with the lateral continuity and the general upward coarsening into wavy-bedded facies of the Glady Fork Member. Furthermore, the inferred clinoform geometry, infilled slump-scar discontinuities, abundant plant fragments, and limited bioturbation are consistent with a gently sloping prodeltaic setting where significant turbidity and/ or salinity fluctuations excluded most benthic fauna.

TIDAL AND CLIMATIC CYCLES IN THE PRIDE SHALE

At least five orders of cyclicity are recognized in the Pride Shale (Fig. 2.3). The cycles are most apparent locally within the eastern outcrop belt, and are less well developed in the western and southern parts of the relatively narrow (~ 15 km across) outcrop belt.

Semidiurnal and Diurnal Couplets

Submillimeter- to millimeter-scale fining-upward couplets are composed of angular, quartz grains of fine silt to fine sand size capped by a dark layer of clays and micas. Couplets generally occur singly, but rarely occur as thick/ thin pairs separated by a minor clay drape.
Each microlaminated couplet is interpreted as a product of suspension fallout whereby silt /sand was deposited during peak tidal flows, and clays during the subsequent slackwater period. Though it is unclear whether these laminae were deposited from flood tides or from ebb tides, their fine-grained lithology and stratigraphic position above the basal condensed section and below the deltaic/ fluvial Glady Fork Member (Englund and Thomas, 1990) is consistent with a basinward setting in which ebb tides distributed suspended sediment across a prodeltaic slope. The paired thick/ thin couplets are interpreted to reflect deposition of sand/ silt from both the dominant and subordinate ebb flows in a mixed, semidiurnal tidal system (cf. Kvale et al., 1989). The rarity of such couplet pairs in the Pride Shale suggests that sand deposition took place only from the strongest ebb flows. Nevertheless, the presence of paired thick-thin couplets reveals the effect of the tropical cycle (changes in lunar declination) on the relative strength of the semidiurnal tides.
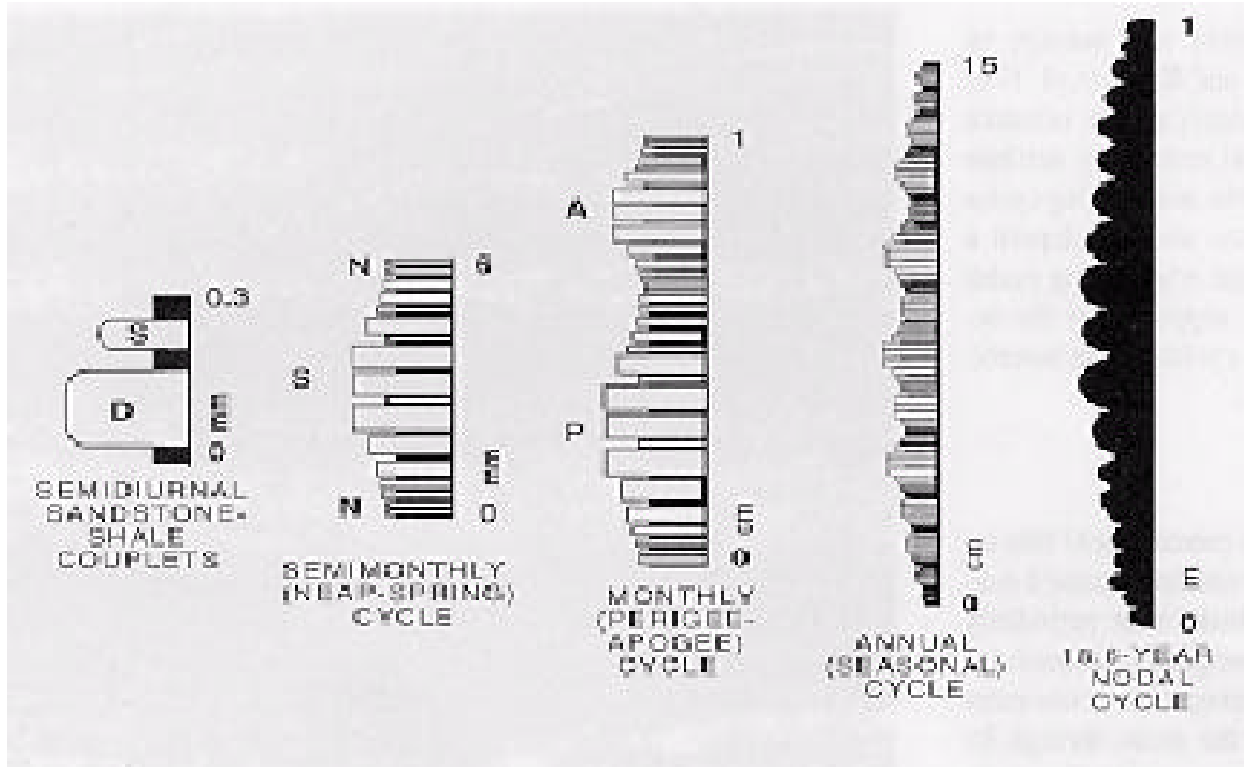
Figure 2.3- Idealized representation of the hierarchy of cycles and cycle thicknesses in the Pride Shale. The five orders of laminae bundling are interpreted to reflect tidal and climatic controls on prodeltaic sedimentation.

Neap-Spring (Semimonthly) Cycles

Up to 17 of the microlaminated couplets are bundled into vertically-accreted cycles 0.1-3 cm thick (Fig. 2.4). Though these cycles commonly contain fewer than 15 individual couplets, nearly all are characterized by microlaminae that display a progressive upward thickening and thinning. The general relationships between couplets and cycle thicknesses in some cycles are suggestive of greater than 17 individual couplets per cycle. Unfortunately, an accurate count of the maximum number of couplets per cycle is limited by the poor resolution of the the thinnest, clay-dominated couplets.

These cycles are interpreted to reflect neap-spring (semimonthly) tidal cyclicity, whereby the thickest couplets were deposited by spring tides during times when the earth, moon, and sun were aligned (syzygy). Despite the low number of couplets per cycle, a semimonthly synodic origin for the neap-spring cyclicity is favored over a tropical mechanism, which is known to force neap-spring cyclicity in some diurnal tidal settings by changes in lunar declination (see Archer et al., 1991 and Kvale et al., 1995 for a discussion of various lunar periodicities). By applying the rationale that laminae preservation is a function of depositional setting, we attribute the abbreviated character (i.e. less than 28 layers) of the neap-spring cycles to a distal setting in which most neap tides were too weak to deposit a recognizable sand-shale couplet. The argument that the neap-spring cycles in the Pride Shale reflect abbreviated tidal cycles is supported by the occurrence of nearly complete (28-layer) neap-spring cycles in the heterolithic deposits of the underlying Princeton Formation.
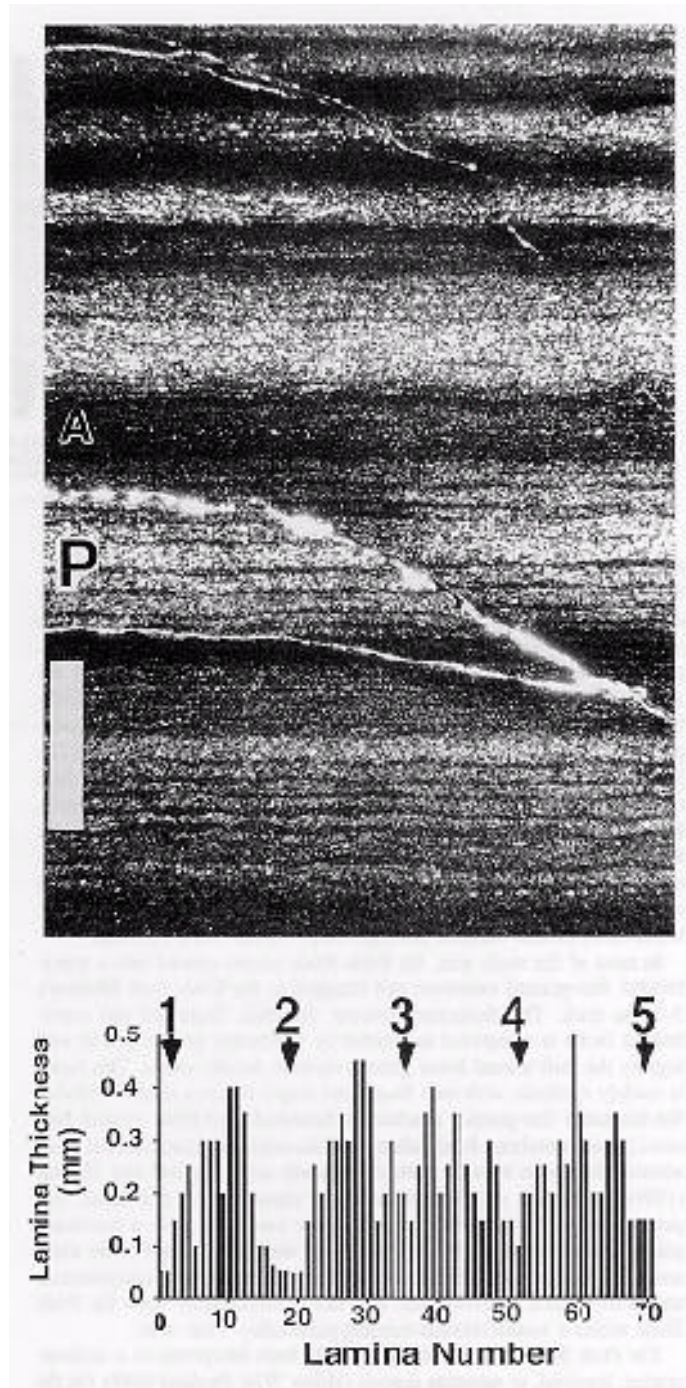
Figure 2.4- a.- Photomicrograph of microlaminated neap-spring cycles in the Pride Shale. Cycles typically contain 15 or fewer distinct sandstone/ shale couplets, which represent semidiurnal/ diurnal laminae deposited primarily during the spring parts of each tidal cycle. Successive thick/thin cycles are interpreted as perigean (P) and apogean (A) neap-spring cycles. Scale bar is 2 mm long. b.- Histogram showing thickness variation in thin section of individual couplets through approximately two months of deposition. Neap periods (arrows) 2 and 5 are characterized by clay-dominated laminae too indistinct to be accurately measured.

Monthly Cycles

Successive fortnightly cycles commonly display a paired, thick/ thin relationship in which the thinner cycles contain fewer and finer-grained couplets (Fig 2.4a). This association reflects the anomalistic lunar periodicity whereby higher-than-average tides accompany the perigean (minimum lunar distance) passage of the moon, and lower-than-average tides accompany the apogean (maximum lunar distance) passage of the moon through its elliptical orbit. Because the anomalistic signal is characteristic of synodically driven semidiurnal tidal systems rather than tropically driven diurnal tidal systems (Wood, 1986), the thick-thin pairing of neap-spring cycles in the Pride Shale provides further evidence that the Late Mississippian paleotidal system in the central Appalachian Basin was one of mixed, dominantly semidiurnal tides.

Seasonal Cycles

The Pride Shale in weathered roadcuts is characterized by a systematic thickening and thinning of 11-18 neap-spring cycles within bundles 2-50 cm thick . Each bundle typically consists of 6-13 relatively well-developed neap-spring cycles bounded by or interspersed with several thin, fine-grained neap-spring cycles. The thin, clay-dominated cycles lack distinctive sandstone/ shale couplets and are bounded by diffuse dark (organic-rich?) layers with a systematic thickness variation consistent with the thickening/ thinning trends seen in the sandier neap-spring cycles. Differential weathering of the well-developed vs. poorly developed neap-spring cycles imparts a washboard-like surface to Pride Shale exposures (Fig. 2.5). The thickest bundles (to 50 cm thick) are composed of sand-dominated neap-spring cycles (Fig. 2.6).

On the basis of the time represented by their component neap-spring cycles, each decimeter-scale bundle represents 6-8 months of deposition (Fig. 2.5b). This cyclicity is interpreted as an annual (seasonal) climatic cycle which defines that part of the paleoyear when fluvial-tidal discharge was sufficient to supply sediment to the subaqueous delta. A similar nontidal annual cyclicity has been described by Chan et al. (1994) from Proterozoic rhythmites in central Utah and by Kvale et al. (1994) from the Pennsylvanian Mansfield Formation in Indiana. The annual cyclicity in the Pride Shale provides strong evidence for a monsoonal climate, and is consistent with seasonality as indicated by vertic to calcic paleosols within the terrestrial redbeds of the Mauch Chunk Group (Cecil and Englund, 1989). Paleogeographic reconstructions (Scotese, 1986; Witzke, 1990) place the central Appalachian Basin within the monsoonal belt (10-15^0 south of the Equator) during Late Mississippian time.
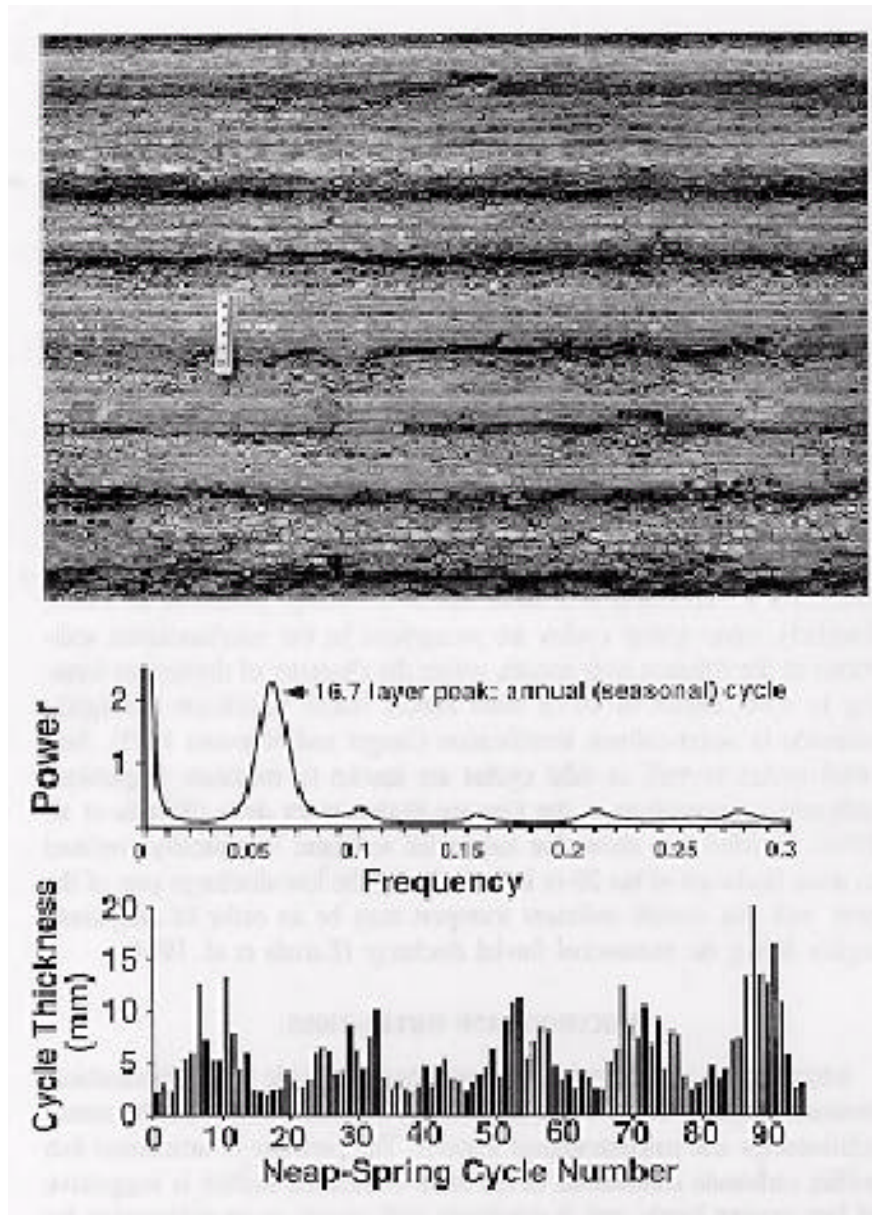
Figure 2.5- a.- Decimeter-scale cycles in outcrop. Each furrow-rib-furrow contains up to 18 recognizable neap-spring cycles, and is interpreted to reflect a nontidal, climatic control on sedimentation. b.-Histogram showing thickness variation of neap-spring cycles through approximately five years of deposition. Maximum entropy spectral plot shows a strong peak at 16.7 neap-spring cycles. Bundling is interpreted to reflect a seasonal (monsoonal) signal wherein the thick neap-spring cycles formed during times of maximum fluvial discharge.

Figure 2.6- Succession of sandy rhythmites within large slump scar. The half-meter-scale thickening/ thinning represents annual bundling of individual sandy neap-spring cycles.

A weakly developed, multiyear cyclicity is recognized in outcrop by light/ dark banding, differential weathering, and progressive thickening and thinning of 17-21 annual beds within meter-scale bundles (Fig. 2.7). Such bundling closely matches the 0.5-2 m-thick cycles which characterize the Pride Shale signature in subsurface gamma-ray and bulk density logs.
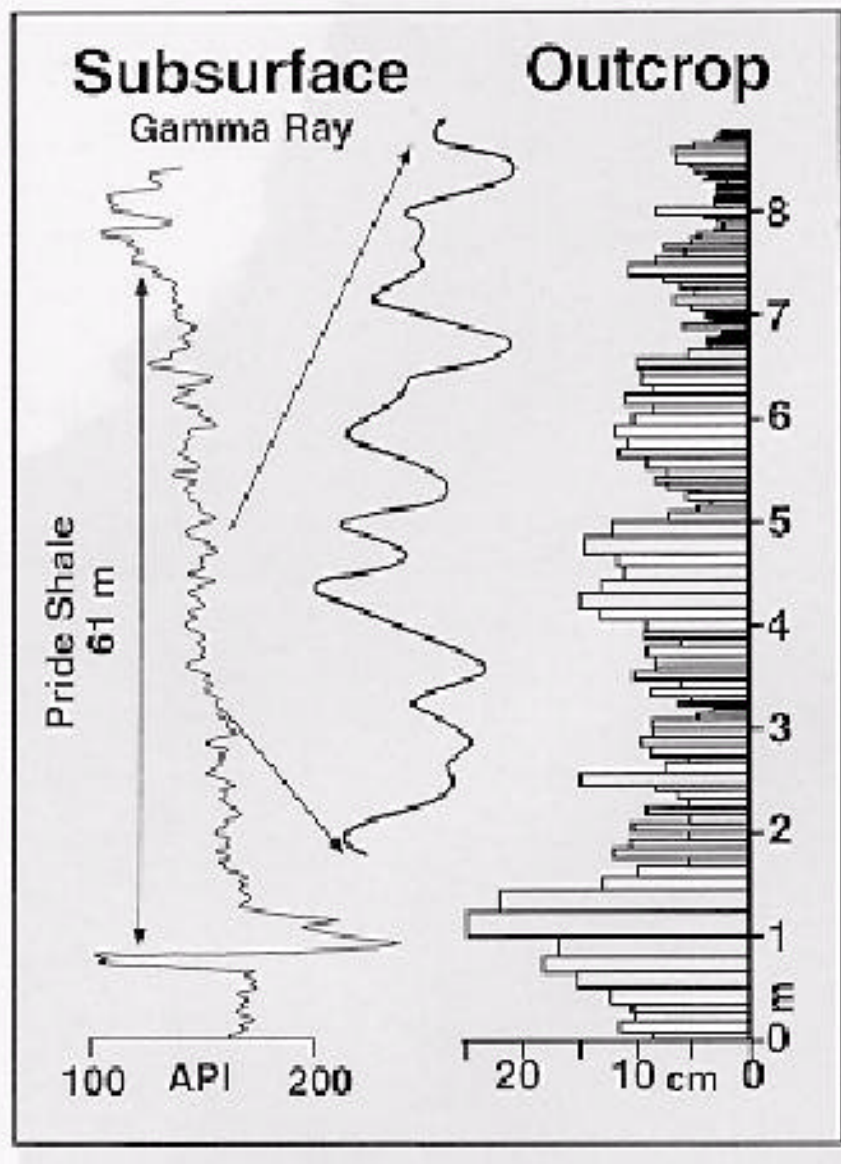

Figure 2.7- Scaled representation of meter-scale cycles in outcrop showing systematic thickening and thinning of annual beds. Bundling of 18-22 annual beds is interpreted to reflect tidal modulation by the lunar nodal cycle. Bundle thickness in outcrop is consistent with meter-scale variation in subsurface gamma ray signatures throughout the study area.

From the number of annual beds, each meter-scale cycle represents (on average) about 19 years of deposition. These cycles are interpreted to reflect tidal amplitude modulation at the lunar nodal period, which at present has an 18.6-yr duration. The nodal cycle relates to the slow rotation of the moon's nodes (points at which the inclined lunar orbital plane crosses the plane of the ecliptic) such that the inclination of the lunar orbital plane varies between 18^0 and 28^0 from the Earth's equator. This $\pm 5^0$ shift in the attitude of the lunar orbital plane is known to weakly modulate all tidal systems and can have a significant effect on sediment transport capacity in tidal settings (Wood, 1986; Oost et al., 1993). The nodal cycle has been invoked as a control on modern sedimentation including mudflat progradation in Guiana (Wells and Coleman, 1981) and tidal-channel fills in the Dutch barrier islands (Oost et al., 1993). Nodal cycles are rarely recorded by ancient sediments, because nearly two decades of uninterrupted sedimentation must be preserved (Oost et al., 1993). One exception is the Elatina Formation, a Precambrian tidal rhythmite succession in Australia, in which Williams (1989a) recognized both the ~ 9 year lunar apsides cycle (rotation of the Moon's perigee) and a 19.5-year nodal cycle from spectral analyses of a 60-year-long rhythmite record.

The meter-scale cycles in the Pride Shale may be the first nodal cycles to be recognized in outcrops of ancient sediments. If a purely tidal explanation is used, the thickest seasonal beds within each nodal bundle likely reflect years during which the lunar orbital plane was best aligned to force tidal oscillation in the Appalachian Basin. Higher-energy ebb flows during such times would have flushed more suspended sediment into the prodeltaic environment. It has been suggested (Marmer, 1951; Kaye and Stuckey, 1973) that the highest-amplitude tides occur during times of extreme high (e.g. 28^0) lunar declination along modern coasts with semidiurnal tides, and during times of extreme low (e.g. 18^0) lunar declination along coasts with diurnal tides. Because the tidal cycles in the Pride Shale are consistent with a synodically driven, mixed semidiurnal setting, it is possible that the thickest annual beds mark those years of maximum lunar declination. Alternatively, the meter-scale cycles in the Pride Shale could reflect lunar nodal control on seasonal precipitation, whereby the thick annual beds reflect wetter-than-average years with elevated fluvial discharge. Such an effect has been noted by Hameed (1984) in the historical record of Nile River floods and by Currie (1988) in the historical record of rainfall/ crop production in North America. Given the degree to which tidal cyclicity is preserved in the Pride Shale, the nodal modulation of tidal amplitudes, rather than climatic shifts, is favored as a primary forcing mechanism behind the meter-scale cycles. It is possible that both forcing mechanisms may have operated in conjunction.

ANALOGS

Known ancient tidal rhythmites that appear similar to the Pride Shale include the Elatina Formation of South Australia (Williams, 1989a). These finely laminated siltstones preserve a long-term (60-year) sequence of neap-spring cycles, and have been interpreted to reflect ebb-tidal deposition in a distal deltaic setting (Williams, 1989a). The thickness, lateral extent, and internal variation of the Pride Shale, together with the lateral variability of the overlying Glady Fork Member, are suggestive of a broad subaqueous slope that received suspended sediment from several points of fluvial discharge along the eastern margin of the central Appalachian Basin (Fig. 2.8). Prodeltaic settings are a possible environment where tidal rhythmites could form, provided that storms,wave action, or bioturbation does not disrupt the stratificaton (Williams, 1989b; Dalrymple, 1992). Smith et al. (1990) documented the tidal control on sedimentation in modern fjord-head deltas in Glacier Bay, Alaska. The prodeltaic sediments that
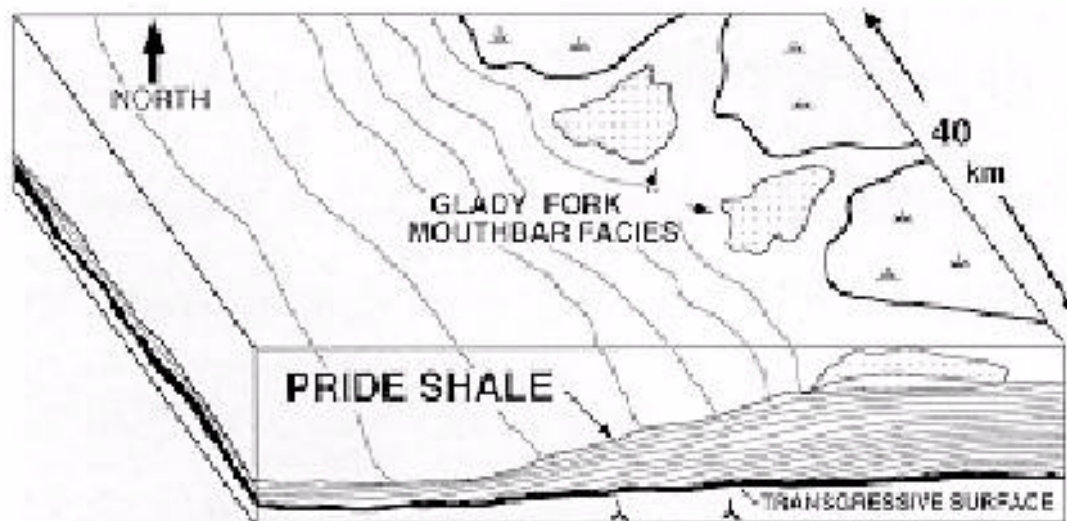
Figure 2.8- Block diagram showing prodeltaic Pride Shale deposition along northeastern margin of the central Appalachian Basin.

accumulated during neap periods were faintly laminated or structureless, whereas distinctly laminated coarse-fine couplets were deposited during spring tides. These authors attributed the alternation of poorly laminated vs. well-laminated sediments to a mechanism of tidal drawdown, whereby high-energy spring-tidal ebb flows remobilize sediment that accumulated on the subaqueous delta plain during the neap part of the tidal cycle. Similar fortnightly flushes of sediment may have taken place in the central Appalachian Basin during deposition of the Pride Shale.

Modern tide-dominated deltas possibly analogous to the Pride Shale include the Fly (Harris et al., 1993), the Amazon (Jaeger and Nittrouer, 1995), and the Ganges-Brahmaputra (Baruda et al. 1994). Laminated sediments in distributary channels of the Fly River delta record neap-spring cycles, and deltaic sedimentation extends to 45 m water depths where prodeltaic sediments are accreting at a mean rate of 4.5 cm/ yr (Harris et al. 1993). Similarly, neap-spring cycles are recognized in the interlaminated sediments of the Amazon river mouth, where the character of rhythmites forming in water depths of 10 or more meters reflect significant fortnightly variation in water-column stratification (Jaeger and Nittrouer, 1995). Seasonal cycles as well as tidal cycles are known to modulate suspended-sediment concentration in the Ganges-Brahmaputra delta (Baruda et al., 1994). Satellite data show that suspended sediment is generally confined to areas landward of the 20-m isobath during the low-discharge part of the year, and that overall sediment transport may be an order of magnitude higher during the monsoonal fluvial discharge (Baruda et al., 1994).

DISCUSSION AND IMPLICATIONS

Interpretation of the forcing mechanisms responsible for the hierarchical laminae bundling in the Pride Shale must be compatible with the stratal relationships and paleontological features. The presence of articulated fish within carbonate concretions in the basal condensed section is suggestive of low oxygen levels, and is consistent with anoxia as an explanation for the general lack of bioturbation. Neal (1994) suggested that the fine, varve-like lamination in the Pride Shale may represent yearly variation in basinwide anoxia. Given that the condensed section represents maximal regional inundation, yearly cycles of deep-water oxygenation is a tenable explanation for the thinnest and presumably most distal laminae in this part of the Pride Shale. However, a yearly or seasonal interpretation for the individual millimeter-thick laminae

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in the bulk of the Pride Shale inadequately explains the hierarchy of cycles. An interpretation of the couplets as annual deposits would require additional forcing mechanisms to explain the centimeter- to meter-scale bundling. Sunspot cycles, for example, was once considered a possible forcing mechanism for the laminae bundling in the Upper Precambrian Elatina rhythmites of South Australia (Williams, 1989a). Like the Elatina rhythmites, the centimeter-scale bundling of couplets within the Pride Shale is an approximate match with known 11-year sunspot or 22-year Hale magnetic cycles. Yet a sunspot or Hale cycle origin for the Pride Shale bundles would require regular annual and decadal modulation of sand or silt influx by solar-forced climatic cycles for tens of thousands of years. This unlikely situation would also require additional forcing mechanisms to explain the decimeter-scale bundles (at a 200-400 yr periodicity) and the meter-scale bundles (at ~ 5000 yr periodicity).

Cyclicity in the Pride Shale is comparable with both modern tidal periods and the interpreted record of Carboniferous tidal rhythmite sedimentation elewhere in the eastern and midcontinental U.S.A. Tidal cycles in the Pride Shale that suggest mixed (semi-diurnal/ diurnal) tidal flows are consistent with the tidal rhythms interpreted from younger (Pennsylvanian) rhythmite successions in the western part of the Appalachian basin (Adkins and Eriksson, 1996), the Black Warrior Basin (Demko et al., 1991), and the Illinois basin (Kvale et al., 1989; Kvale and Archer, 1991). As a thick and finely laminated tidal rhythmite deposit, the Pride Shale preserves a lengthy, quasi-continuous sedimentary record. The abbreviated character of the preserved tidal record reflects: (1) reduced influx of sediment during subordinate semidiurnal, neap, and apogean ebb-tidal flows, and (2) pulselike annual sedimentation as modulated by seasonal fluvial discharge. The decimeter-scale cycles that reflect monsoonal climatic conditions are consistent with the evidence for seasonality recorded in paleosols in the terrestrial redbeds below and above the Pride Shale. Such climatic signals in both marine and terrestrial deposits supports independently constrained paleogeographic reconstructions (Scotese et al., 1986; Witzke, 1990) that place the Late Mississippian Appalachian basin within the southern, subtropical monsoonal belt.

The interpretation of the bundling in the Pride Shale as tidal and climatic cycles is further supported by the preservation of delicate-bodied faunas that suggest rapid burial. The limited bioturbation in the Pride Shale likely reflects turbid conditions, and the possibility that significant salinity fluctuations accompanied sediment dispersal as mud-laden fluvial systems emptied into the central Appalachian basin. Rapid accumulation rates (3-60 cm/yr) suggested by decompacted annual cycles in the Pride Shale compare favorably with other known tidal rhythmites, and highlight the punctuated nature of sedimentation in ancient tide-dominated systems. Thick intervals of rhythmic beds indicate that in any one vertical section the Pride Shale represents several hundred years of accumulation. Although single outcrops are consistent with rapid infill of available accommodation space, it seems likely that infill of the Pride basin occurred via progradation of broad, overlapping prodeltaic clinoforms. Any consideration of the time represented by the rhythmic Pride Shale intervals therefore must be weighed against the time represented by the basal condensed section and the outcrop-scale discontinuities, as well as the clinoform geometry and probable diachronous development of the the overlying Glady Fork facies.

CONCLUSIONS

1. Rhythmic, flat-laminated facies of the Pride Shale overlie a basal condensed section and grade upward into wavy-bedded sandstone facies. The Pride Shale is interpreted as a prodeltaic deposit, and together with the the younger Glady Fork facies, it may represent one of the few ancient examples of a prograding, tide-dominated delta.

2. A hierarchy of cycles within the Pride Shale is manifested as sub-millimeter- to half-meter-thick bundles of vertically accreted rhythmites. The cycles reflect fluctuating tidal energy related to semidiurnal, diurnal, neap-spring, and anomalistic tidal periodicities, and seasonal pulses of sediment modulated by monsoonal fluvial discharge.

3. Meter-scale variability in grain size and bed thickness in the Pride Shale is interpreted to reflect variation in tidal amplitudes forced by a ~ 19-yr lunar nodal cycle. These bundles are amongst the first long-term tidal cycles to be recognized in outcrop.

4. Sedimentation rates for the Pride Shale were on the order of centimeters to decimeters per year. Such rates may reflect accumulation in relatively proximal parts of the basin, yet highlight the character of prodeltaic sedimentation seaward of a tide-dominated, high-suspended-load fluvial system.

Chapter 3:
Linked Sequence Development and Global Climate Change: The Upper Mississippian Record in the Appalachian Basin

ABSTRACT

The character and relative stratigraphic position of paleoclimatic indicators within Upper Mississippian strata of southern West Virginia are suggestive of a link between eustasy and patterns of continental - to global-scale atmospheric circulation. At the cyclothem (~400 k.y.) scale, annual rhythms in marine facies, and vertisols and lacustrine carbonates in terrestrial units are suggestive of seasonal, semiarid climatic conditions during highstand progradation. In contrast, leached paleosols and coals that underlie sequence boundaries and occur within transgressive heterolithic facies are suggestive of humid climatic conditions during late highstand through early transgression.

Milankovitch-band climatic change that forced Gondwanan glacial episodes may explain both the cyclothemic sequence development and the lithologic evidence for fluctuating climatic conditions in the Appalachian basin during the late Mississippian. The shifts from humid climatic conditions during lowstands to seasonal conditions during highstands is attributed to systematic variation in monsoonal circulation during each glacial/ interglacial cycle. Such climatic change is consistent with global circulation models if : 1) the climate in the Late Paleozoic Pangean intertropical belt was driven by monsoonal circulation rather than by zonal circulation, and 2) the latitudinal shift of seasonal moisture belts was greatest (e.g. to $\pm$ 23$^{\text{O}}$) during non-glacial periods, and was restricted to the equatorial belt during glaciations. Given this model, everwet conditions in the early Pennsylvanian may reflect an orographic effect whereby, despite a tendency toward intertropical seasonality, a low pressure cell anchored by the central Pangean mountains drew moisture from the southwestern part of the basin.

INTRODUCTION

Glacioeustasy is a favored mechanism to explain the cyclothemic (fourth-order; 0.2-0.5 m.y.) Carboniferous stratigraphic record in North America and in Europe (Veevers and Powell, 1987; Heckel, 1986; Boardman and Heckel, 1989; Read, 1991; Maynard and Leeder, 1992; Crowley et al, 1993; Hampson et al. 1997). Glacial-interglacial oscillations and attendant eustatic change is thought to reflect Milankovitch insolation cycles during the Gondwanan polar glaciation (330-250 Ma)(Crowell, 1978; Frakes et al., 1992). Calculated sequence durations for cyclothemic successions suggests that periodic global warming and cooling in the Carboniferous was linked to the ~400 k.y., long-eccentricity cycle (Algeo and Wilkinson, 1988; Soreghan, 1994a, Smith, 1996; Al-Tawil, 1998).

The relationship between Milankovitch-band climatic change and glacial-interglacial cycles can be explored because lithologic indicators of paleoclimate (eg. paleosols, coals, eolianites) serve as proxy indicators of global change in atmospheric circulation. Tandon and Gibling (1994; 1997) noted that lowstand facies within Late Pennsylvanian cyclothems of the Sydney Basin (Nova Scotia) are associated with calcretes, whereas the highstand facies contain coals. This pattern implies that dry conditions prevailed during eustatic lowstands. Similarly, Soreghan (1992, 1994a) suggested that siltstones of eolian origin within Pennsylvanian and Permian cyclothems of the Ancestral Rockies define periods of aridity during glacial episodes. These examples are paralleled by Pleistocene and Holocene successions which suggest that tropical regions become wetter during interglacials and more arid during glacial maxima (Williams, 1985; Nanson et al., 1992).

This study focuses on Upper Mississippian strata in the Appalachian basin of southern West Virginia (Fig. 3.1). Paleosols within the predominantly terrestrial section suggest that

seasonal, semiarid climatic conditions were dominant over comparatively minor episodes of humid (coal-forming) conditions (Cecil, 1990; Caudill et al., 1996). Cecil (1990) suggests that this record reflects pulse-like climatic change toward everwet conditions as the basin moved northward into the equatorial rainbelt by the early Pennsylvanian. Continued study and interpretation of the Upper Mississippian section within a sequence stratigraphic framework reveals a record of base-level change that is consistent with glacioeustatically-forced cycles of
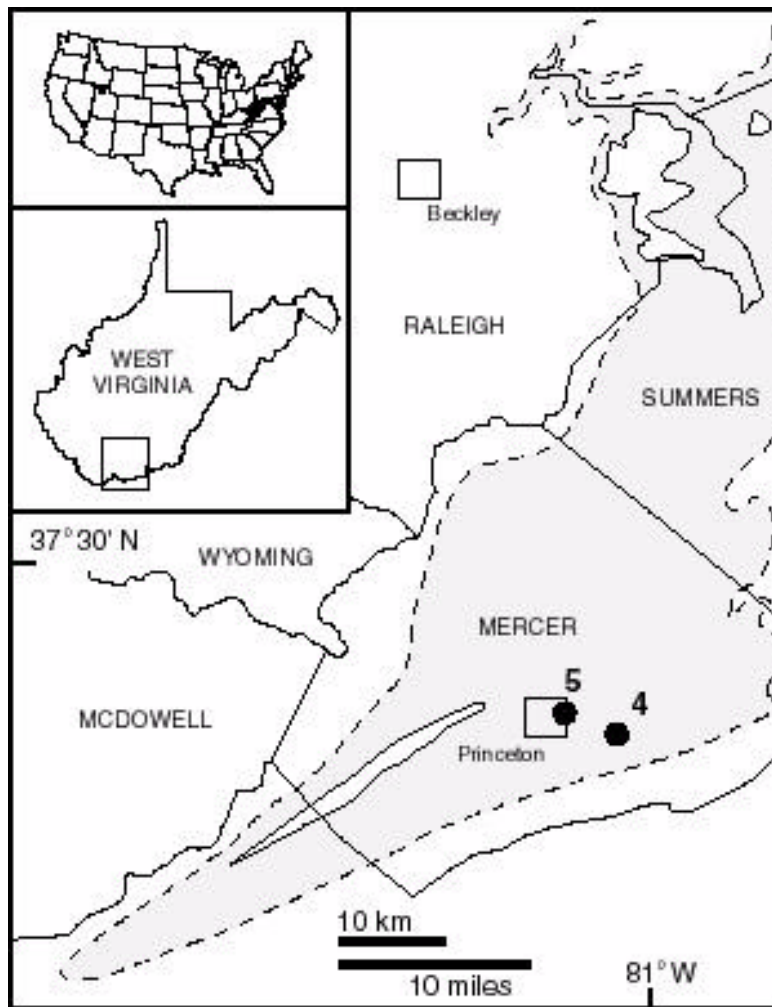
Figure 3.1- Map of the study area in southern West Virginia showing the outcrop belt of Upper Misssissippian strata (shaded) and the locations of cross-sections shown in figures 3.4 and 3.5 (bold dots).

~400 k.y. duration (Miller and Eriksson, in review). Furthermore, the relationship between indicators of wet climatic conditions and the bounding surfaces of depositional sequences suggests a link between eustatic and climatic fluctuations. In contrast with the Pennsylvanian examples mentioned above, the Upper Mississippian record in the Appalachian basin suggests that seasonal, semiarid conditions characterized the times of eustatic rise and highstand, whereas humid conditions prevailed during times of eustatic fall, and early eustatic rise.

STRATIGRAPHIC FRAMEWORK, DEPOSITIONAL AND PALEOGEOGRAPHIC SETTING

The Upper Mississippian Hinton, Princeton, and Bluestone formations of southern West Virginia comprise a stratal wedge that thins westward from a maximum thickness of ~600 m in Mercer County, West Virginia (Figs. 3.1, 3.2; Arkle, 1974). These strata are dominated by pedogenically-altered, terrestrial mudstone redbeds that filled the central Appalachian foreland basin as the region moved northward through near-equatorial latitudes (15 to 5 degrees south) in association with the collisional assembly of Pangea (Fig. 3.3; Scotese and Barrett, 1990; Golonka et al., 1994). Below the study interval, oolitic grainstones, aridisols, eolianites, and evaporites of the Greenbrier Formation developed within an semiarid tropical belt during the initial onset of Alleghenian orogenesis (Cecil et al., 1997; Al-Tawil, 1998). Overlying quartzoze sandstones, paleosols with aluminum-rich clays, and thick, low ash, low sulfur coals of Pennsylvanian-age indicate that swampy environments and everwet climatic conditions prevailed by the early Pennsylvanian (Phillips and Peppers, 1984; Cecil et al., 1985; Cecil, 1990).

Integrated outcrop and well-log correlation within the Upper Mississippian section permits characterization of 17 unconformity-bounded, transgressive-regressive sequences (Fig. 3.2; Miller and Eriksson, in review). Because existing biostratigraphic and geochronological data (Pfefferkorn and Gillespie, 1982; Gordon et al., 1983; Repetski and Henry, 1983; Roberts et al., 1995a; Jones, 1996) indicate that the study interval records ~7 million years, the average duration of individual sequences (~400 k.y.) is interpreted to reflect glacioeustatic cyclicity during the early stages of Gondwanan glaciation (330-250 Ma; Gastaldo et al., 1996; Miller and Eriksson, in review).

The unconformity-bounded sequences are comprised of transgressive and highstand deposits. Transgressive deposits occupy incised paleovalleys and consist of: 1) basal fluvial facies that overlie sequence boundaries in erosional contact with underlying paleosols and coals; and 2) fining-upward heterolithic deposits that contain thin coals. Maximum flooding is recorded by marine shale or limestone units, and overlying highstand systems tract deposits consist of progradational deltaic facies or pedogenically altered fluvial mudstones and sandstones.

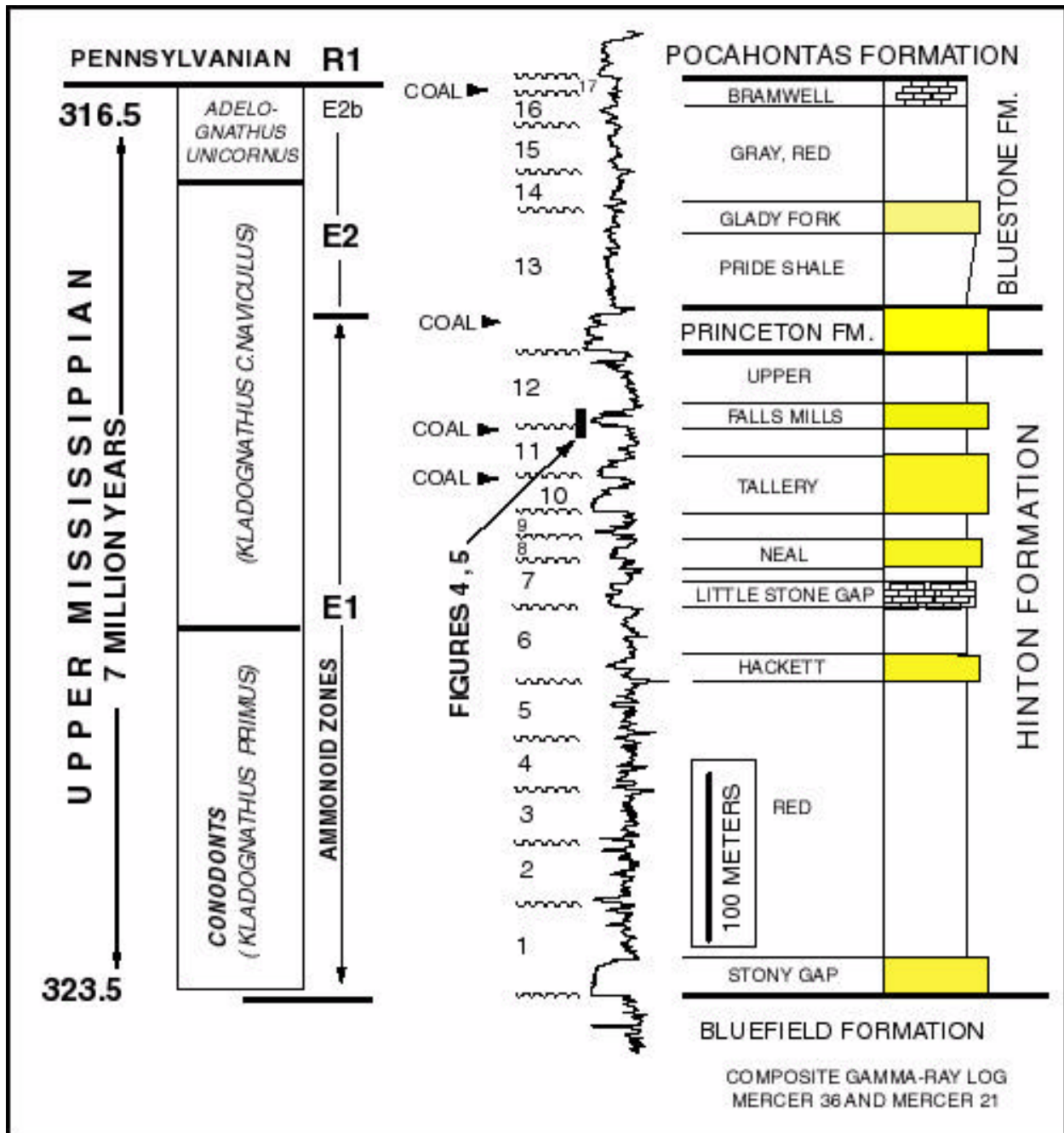Figure 3.2- Stratigraphic column of Upper Mississippian strata in southern West Virginia. Biostratigraphic and geochronologic age constraints from Manger and Sutherland (1984) and Jones (1996). Vertical pedogenetic trends and other indicators of paleoclimate such as coals are shown with respect to boundaries of 17 sequences (numbered). Sequences are interpreted to reflect eustatic sea-level changes on a ~400 k.y. time scale.

An unconformity that may represent several million years of non-deposition/ erosion separates the Mississippian and Pennnsylvanian strata in the Appalachian Basin (Beuthin, 1994; 1997). This unconformity apparently developed during eustatic fall toward a mid-Carboniferous lowstand (Saunders and Ramsbottom, 1986). Plant megafossils below the unconformity link the present study interval with Chesterian successions in the midcontinental U.S.A. and with the Namurian A strata of Europe (Gillespie and Pfefferkorn; 1979; Pfefferkorn and Gillespie, 1982). A significant paleobotanical shift is defined along the Mississippian-Pennsylvanian boundary by the dissappearance of the Namurian A index flora, and the first occurrence of more diverse plant species adapted to a swamp habitat (Pfefferkorn and Gillespie, 1981; Wagner, 1982; Jennings, 1986). Plant fossils in the lowermost Pennsylvanian units in the Appalachians (Fig. 3.2) suggest a correlation with the Namurian B strata of Europe (Kosanke, 1984; Pfefferkorn and Gillespie, 1982). The major Mississippian-Pennsylvanian (mid-Carboniferous) vegetational change is seemingly tied to a global climatic change (from semiarid to wet conditions) within the paleoequatorial belt (Gastaldo et al., 1996).
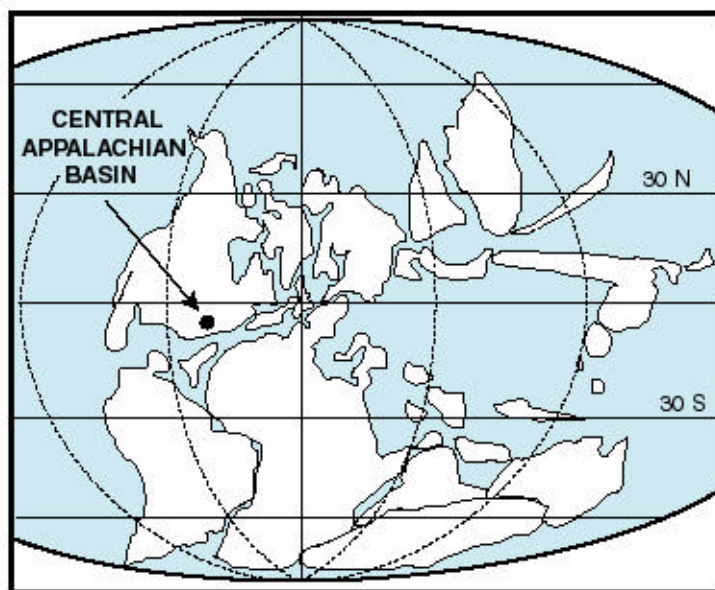
Figure 3.3- Paleogeographic reconstruction for Late Mississippian (Mid-Carboniferous) time redrawn from Cleal and Thomas (1996) after Scotese (1986). The central Appalachian Basin occupied a near-equatorial (5-10 degrees south) paleolatitude during the early stages of Alleghenian orogenesis.

LITHOLOGIC INDICATORS OF PALEOCLIMATE

Vertic and Calcic Paleosols

Vertically stacked paleosols constitute a significant proportion (50% or more) of the terrestrial redbeds in the Hinton and Bluestone formations (Fig. 3.2). These red beds are interpreted to represent fluvial channel and floodplain deposits upon a low-relief coastal plain (Miller and Eriksson, in review). The red vertisols are characterized by thick (up to 2 m), clay-rich B horizons with well-developed angular-blocky pedogenic fabrics and slickensides. Root traces are locally present as downward branching, gray-green reduction haloes. Isolated carbonate nodules up to 3 cm in diameter are locally developed. In rare cases, coalesced carbonate nodules occur as discontinuous horizons within the soil profile.

The vertisols and calcic vertisols are suggestive of seasonally dry climatic conditions in the Appalachian Basin during much of the Late Mississippian (Cecil, 1989; Cecil, 1990). Modern vertisols develop in strongly seasonal climates (4 to 8 dry months per year) in areas where the water table is low (Ahmad, 1983; Gibling and Bird, 1994). Geochemical analysis of equivalent paleosols from the Pennington Formation in Tennessesse has shown that the Late Mississippian climate in the Appalachian region was hot, and markedly seasonal with yearly precipitation similar to that in present-day south Texas (Caudill et al., 1996).

Lacustrine Limestones

Laterally continuous carbonate units up to 1 meter in thickness are interbedded with the vertisols. Ostracodes and bivalves are sparsely distributed within these units, which, in thin section, are comprised of dense micrite with a weakly laminar or peloidal fabric. These carbonates commonly show evidence of pedogenic disruption and grade upward into vertisols (cf. Tandon and Gibling, 1997).

The bedded carbonates are interpreted as lacustrine limestones that developed as a result of episodic inundation of the coastal floodplain. The lack of root traces and close association with vertisols suggests that the lacustrine carbonates developed under seasonal climatic conditions that may have limited colonization by plants (cf. Platt and Wright, 1992).

Tidal Rhythmites

Thinly laminated sandstone-shale couplets of the prodeltaic Pride Shale Member (Fig. 3.2) preserve a hierarchy of tidal cyclicities (Miller and Eriksson, 1997). Up to 17 individual neap-spring cycles are characteristically bundled into decimeter-scale cycles that record at least 8 months of near-continuous deposition. The decimeter-scale cycles are interpreted as yearly cycles of sedimentation that reflect seasonal fluvial discharge during highstand deltaic progradation. The thickness of annual sedimentary cyclicity in the Pride Shale is suggestive of an average of 10 cm of local accumulation per year. Such high depositional rates may be localized in the active part of the subaqueous delta, yet are nevertheless supportive of the general model of Cecil (1990) that suggests that highly seasonal climates are characterized by high sediment yields.

Leached Paleosols and Highstand Coals

In contrast with the indicators described above as characteristic of semiarid or monsoonal climates, leached, grey-white paleosols and thin (<40 cm) coals and are present in the upper parts of at least 7 of the 17 sequences recognized in the study interval (Fig. 3.2). Intracylothem, vertical pedogenetic trends are best preserved by sequences in the upper Hinton Formation (Fig. 3.2) where the upper part of Sequence 11 is comprised of terrestrial redbeds (channelized sandstones within vertically stacked vertisols) overlain by green-gray to gray-white paleosols and thin coals (Figs. 3.4, 3.5). The transitional zone between the red vertisols and the thick (3 m) green-gray interval is characterized by numerous downward branching reduction haloes. The lowermost paleosol profile in the green-gray interval is characterized by pedogenic fabrics and slickensides that are typical of the red vertisols lower in the sequence. The overlying gray-white interval is a clay-rich paleosol up to 2 meters thick that lacks blocky pedogenic fabric and which is distinctly lighter in color in its upper 10 centimeters. A series of thin (few cm to 20 cm) coals and dark, carbonaceous paleosols are present locally above the gray-white paleosol and below the erosional lower contact with overlying sandstones (Figs. 3.4, 3.5). The green to white paleosols and carbonaceous intervals/ coals are not present 10 km to the north where a thick section of cross-bedded to wavy-bedded sandstones rest unconformably upon red vertisols.
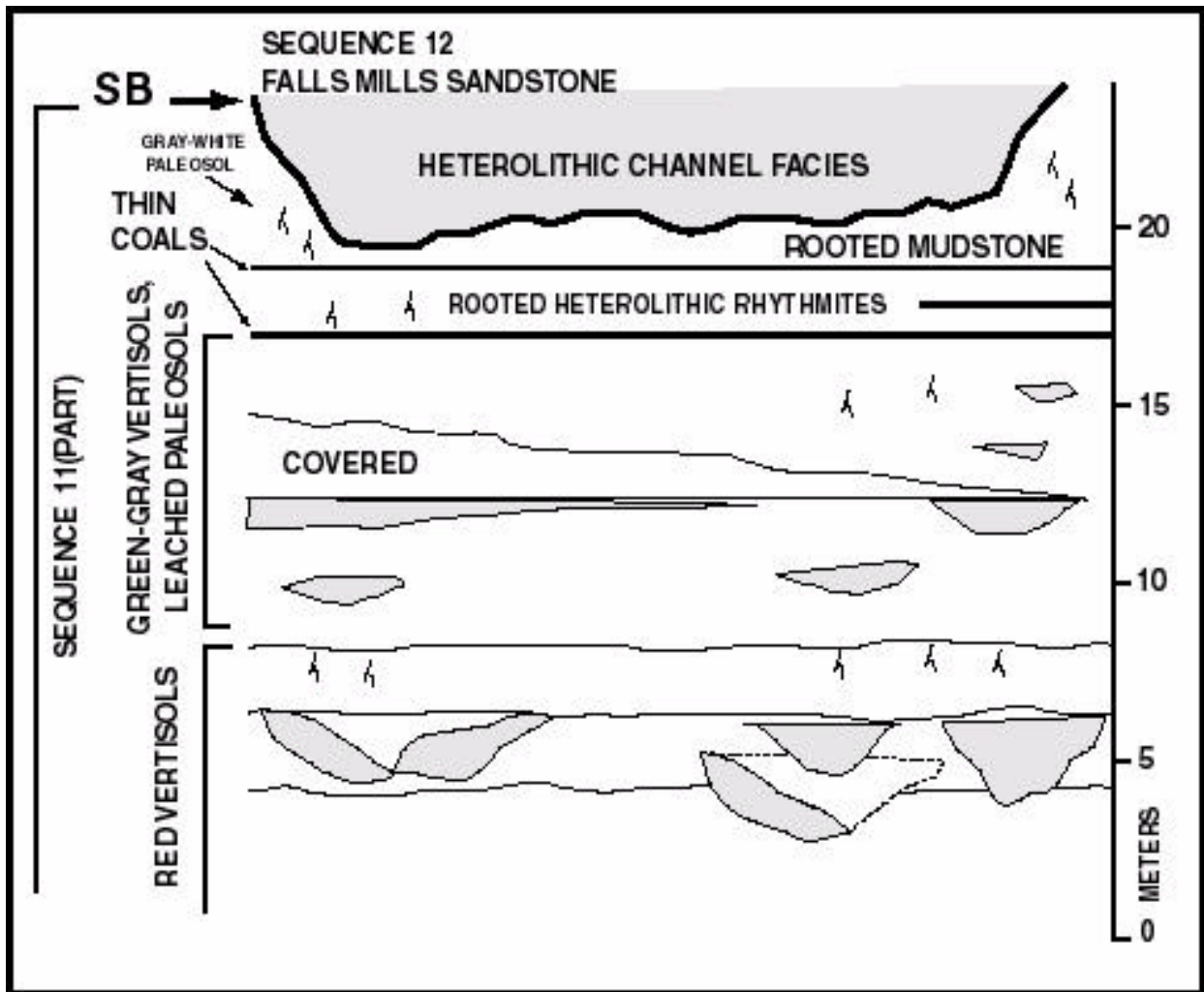
Figure 3.4- Cross-section showing upward transition from fluvial, coastal plain redbeds to green-gray to gray-white paleosols and coals at the top of Sequence 11. Exposure is located along Rt. 460, 2.5 km east of intersection with Interstate 77 near Princeton, West Virginia.
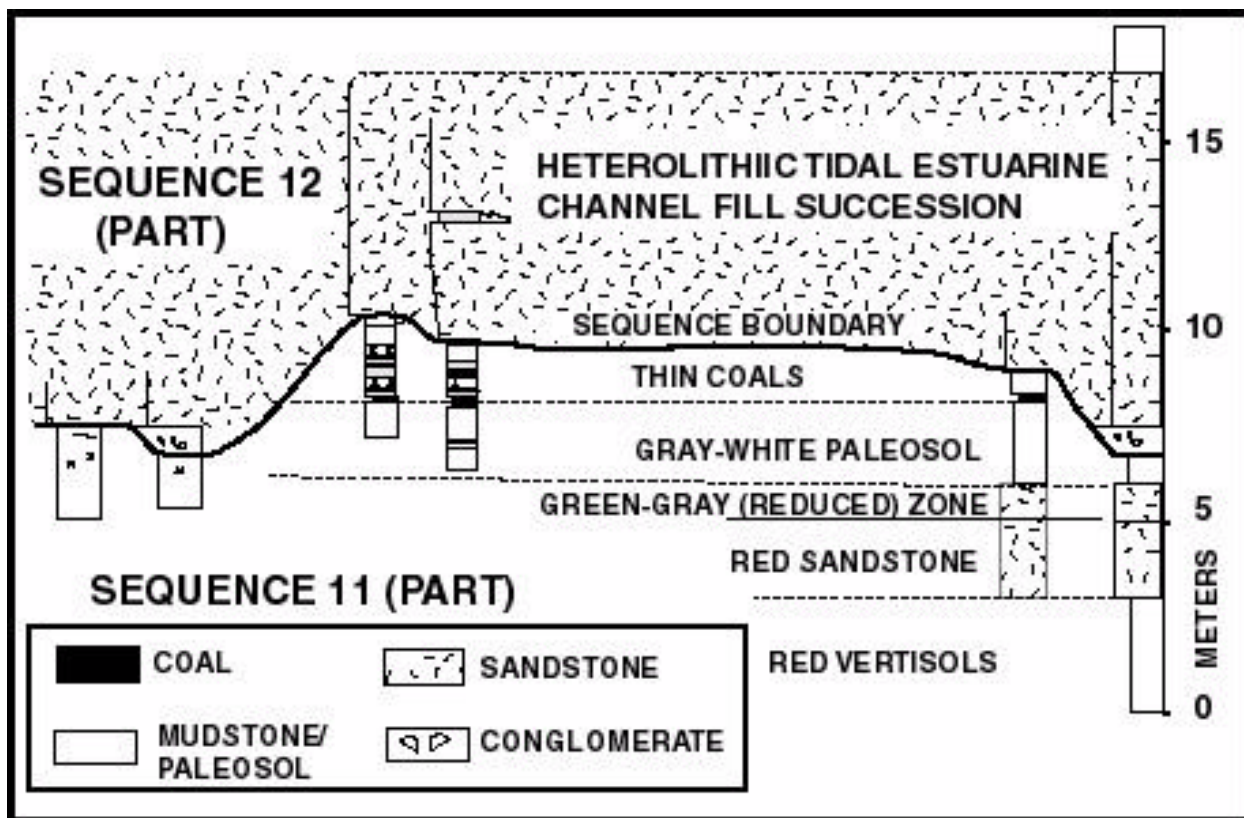
Figure 3.5- Cross-section of the upper part of Sequence 11 showing erosion of coal-bearing horizon in paleotopographic lows. Heterolithic deposits of overlying sequence are tidal estuarine facies deposited during relative sea level rise. Exposure is located 5 km east of section shown in Figure 3.4. Section extends north from the intersection of Rt. 460 (business) and Rt. 20 in Princeton, West Virginia.

     The green-gray interval is interpreted as a series of vertisols that were modified (gleyed) during a time of wet climatic conditions when a relatively high (seasonally?) water table permitted colonization by plants and reduction of iron in response to organic decay (cf. Retallack, 1990). The gray-white paleosol is interpreted as a highly leached soil that developed under waterlogged conditions; its uppermost horizon may represent the underclay of the overlying coal (cf. Retallack, 1990). Carbonaceous horizons and coals that mark the top of the sequence indicate that peat swamps were present prior to valley incision and backfill. The poor development and local extent of these coals suggests that humid conditions were short lived, yet this assumption ignores the potential for loss of section during eustatic falls. Where the valley-fill succession is thickest in the more deeply incised parts of the paleovalleys, basal fluvio-estuarine facies rest upon red mudstones of the underlying highstand deposits. The lesser thickness of the green to white soils and coals is likely a reflection of the fact that the most complete record of humid conditions is preserved only on paleovalley interfluves.

     A similar vertical progression from red vertisols to green to white paleosols to coal is present in the upper part of Sequence 10. The coal horizon (to 30 cm thick) that lies immediately below the sequence boundary is abruptly overlain by rhythmic facies that are interpreted as tidal estuarine deposits (Miller and Eriksson, in review; cf. Archer et al, 1994).

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Transgressive Coals

Thin (< 5 cm) coals are present within the estuarine deposits (Princeton Formation) of sequence 13. The coals are interpreted as marking times of organic productivity within coastal or estuarine-margin swamps during the step-wise infilling of a deeply incised paleovalley (Miller and Eriksson, in review). Such coals are evidence that climatic conditions were sufficiently wet for peat formation. However, such coal deposits within the transgressive parts of sequences are equivocal evidence for global climate change because the organic productivity may be linked to locally waterlogged conditions resulting from a rising water table during times of eustatic rise.

CLIMATIC FRAMEWORK

Climatic Zones

Latitudinal differences in solar insolation and the Coriolis effect tend to produce a zonal distribution of climatic belts around the modern Earth (Fig. 3.6). Hadley cell circulation of rising air (low pressure) in equatorial latitudes and descending air (high pressure) at the tropics results in an intertropical convergence zone where descending air turns eastward along the equator (Scotese and Barrett, 1990). The eastwardly flowing air tends to gain moisture over oceanic regions to produce a circumglobal belt of high rainfall in equatorial latitudes.
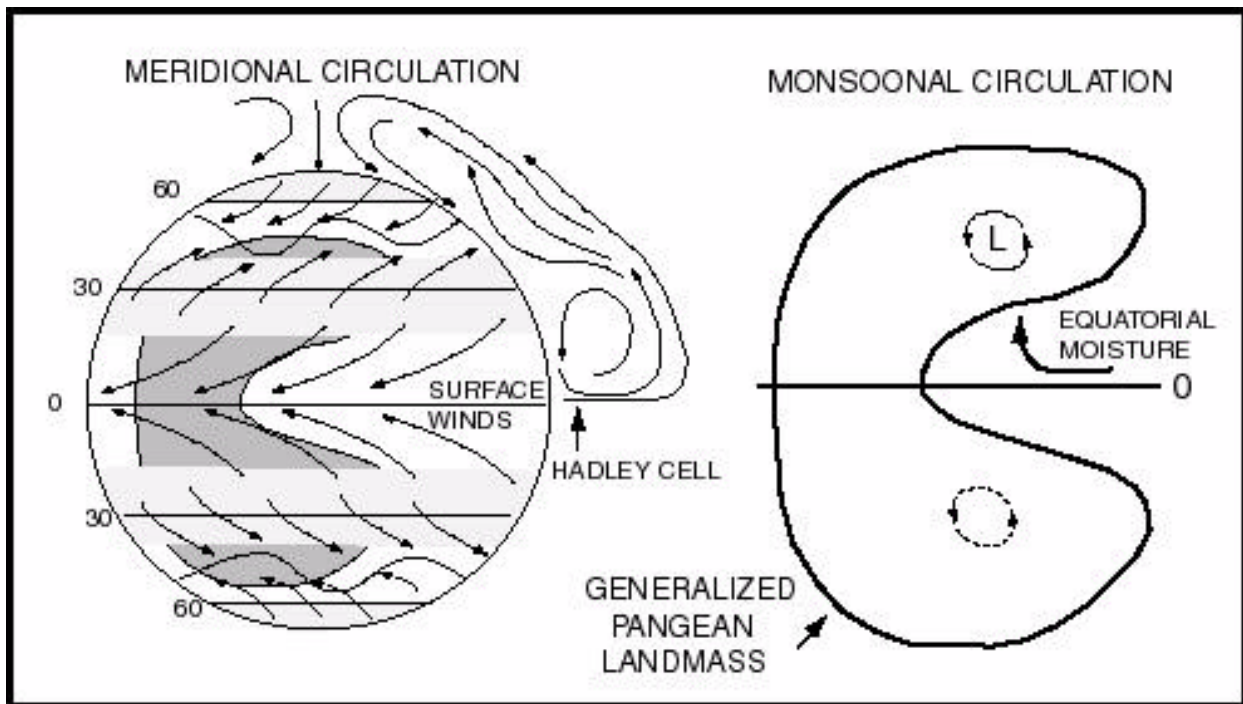
Figure 3.6- Model of atmospheric circulation showing a.) climatic zonation resulting from meridional circulation (after Scotese and Barrett, 1990), and b.) the diverting effect of monsoonal circulation into seasonally-developed low-pressure systems (after Parrish and Petersen, 1988).

<center>Monsoonal Effects</center>

Patterns of global atmospheric circulation (Fig. 3.6) are affected by the contrasting rates of heat exchange of continents and oceans. Because continents gain and lose heat rapidly, large landmasses experience significant seasonal temperature contrasts (Parrish, 1993). This seasonal warming favors the development of monsoonal circulation systems that draw the intertropical convergence zone away from the equator during the summer (Witzke, 1990; Scotese and Barrett, 1990; Parrish, 1993; Perlmutter and Matthews, 1994). The intertropical convergence zone can make especially large latitudinal shifts where continents and oceans are juxtaposed in low latitudes (Hay et al., 1990). Wright (1990) invokes strong monsoonal circulation to explain the seasonal to arid climatic conditions within paleoequatorial latitudes that are recorded by early Carboniferous (Tournaisian-Visean) carbonates of southern Britain. Additional lithologic indicators of seasonality that point to monsoonal conditions occur in Mississippian through Triassic sedimentary successions in other low-latitude Pangean basins (Witzke, 1990; Vanstone, 1991; Witzke, 1990; Dubiel and Smoot; 1994; Francis, 1994; Miller et al. 1996). Paleoclimatic models (Rowley et al., 1985; Parrish and Petersen, 1988; Crowley et al., 1989; Kutzbach and Gallimore, 1989; Crowley, 1994) confirm that the Pangean climate was likely dominated by monsoonal circulation as the continental landmasses assembled to lie symmetrically disposed about the equator (Fig. 3.3; Parrish, 1993).

<center>Effect of Glacial/ Interglacial Episodes</center>

Glacial episodes associated with Gondwanan continental glaciation (330-250 Ma) may have exerted a control on the latitudinal width of monsoonal systems in the Pangean intertropical belt. Seasonal effects are amplified during interglacial times when increased solar insolation (related to the lack of reflective ice-cover) forces strong temperature contrasts that result in a significant latitudinal increase in wet-dry seasonality (Witzke, 1990; Miller et al., 1996; Gastaldo et al., 1996). In contrast, during times of global cooling (continental glaciation), climatic belts and the intertropical convergence zone tend to contract about the equator (Ziegler et al., 1987; Scotese and Barrett, 1990; Gastaldo et al., 1996). During glacial episodes, monsoonal circulation systems may be weakened as the seasonally warm belt migrates over a narrower range of latitude. Consequently, low-latitudes that experienced significant north-south shifts in seasonal rainfall during interglacial times may experience more equable, warmer, and wetter climates during glaciations as the seasonal component of climate was lessened (Ziegler et al., 1987). Miller et al. (1996) proposed a climatic-eustatic model that invoked changes in the intensity of Permian monsoonal circulation to explain wet-dry cycles in paleosols of the midcontinental United States. These authors suggested that the wet intertropical climatic conditions during lowstands reflect weakened monsoons and greater influx of equatorial moisture into the continental interior.

<center>CARBONIFEROUS CLIMATE IN THE APPALACHIAN BASIN</center>

The Late Mississippian climate in the Appalachian basin has been interpreted as being seasonally dry with punctuated episodes of wet conditions (Cecil, 1990). Cecil (1994) suggests that the climatic changes during this time may be related to fluctuations in base level, but also suggests that the climatic/ base level cycles were not necessarily in phase (Cecil, pers. comm, 1998). The intracyclothem paleoclimatic trends described in the present study, however, suggest that seasonal (semiarid) climatic conditions prevailed during eustatic highstands (interglacials), and comparatively humid conditions prevailed during eustatic falls, lowstand (glacial maxima) and early eustatic rise. This pattern is reported for some Middle Pennsylvanian Euramerican strata in that extensive peat swamps developed under everwet conditions during glacial maxima, whereas marine units that mark interglacial periods correspond with a shift

toward more seasonal climatic conditions (Gastaldo, et al. 1996; Cecil et al., 1997). Such systematic climate change is suggestive of a link between paleoclimatic change in the subequatorial belt and the glacioeustatic cycles that seemingly forced fourth-order sequence development. If monsoonal circulation was dominant upon Pangea in the Late Mississippian, then variation in the intensity of monsoonal circulation during glacial versus interglacial episodes may account for the record of Late Mississippian wet-dry cycles in the Appalachian Basin (Fig 3.7; cf. Heckel, 1995). This glacioeustatic-climatic model is similar to the one proposed by Miller et al. (1996) whereby only during glacial periods are monsoonal circulation systems sufficiently weakened such that the equatorial region remains comparatively wet throughout the year.

CLIMATE CHANGE ASSOCIATED WITH PENNSYLVANIAN CYCLOTHEMS

The Upper Mississippian paleoclimatic record described above contrasts with intra-cyclothem climatic trends reported for some Pennsylvanian cyclothems. Tandon and Gibling (1994, 1997) suggest that arid climatic conditions were prevalent during lowstand phases of Late Pennsylvanian cyclothems in the Sidney Basin, Nova Scotia. In this region, sequence boundaries are marked by calcretes, whereas coals are associated with highstand deposits. Similarly, Soreghan (1992, 1994a) suggests that eolianites within the Upper Pennsylvanian cyclothems of the Ancestral Rockies are indicators of strong monsoonal winds in the northerly subtropical latitudes of western Pangea. These eolianites are not present within highstand (interglacial) deposits, and are interpreted as marking times of aridity during glacial lowstands. Both of these Late Pennsylvanian successions were deposited in near-equatorial latitudes, yet record climatic changes that parallel Pleistocene stratigraphic records (Williams, 1985; Nanson et al., 1992) which indicate that arid climatic conditions dominate tropical latitudes during glacial lowstands. It may be the case, however, that Pleistocene records may serve as a poor analog for Pangean successions because global circulation (eg. monsoonal vs. zonal) may differ upon contiguous vs. dispersed continental configurations (Witzke, 1990; Miller and West 1993; Parrish, 1993). Because of complicating geographic or topographic factors discussed below, the wet-dry polarity of the climate change within Pennsylvanian cyclothems of North America does not necessarily conflict with the monsoonal model that is applied to the present study.
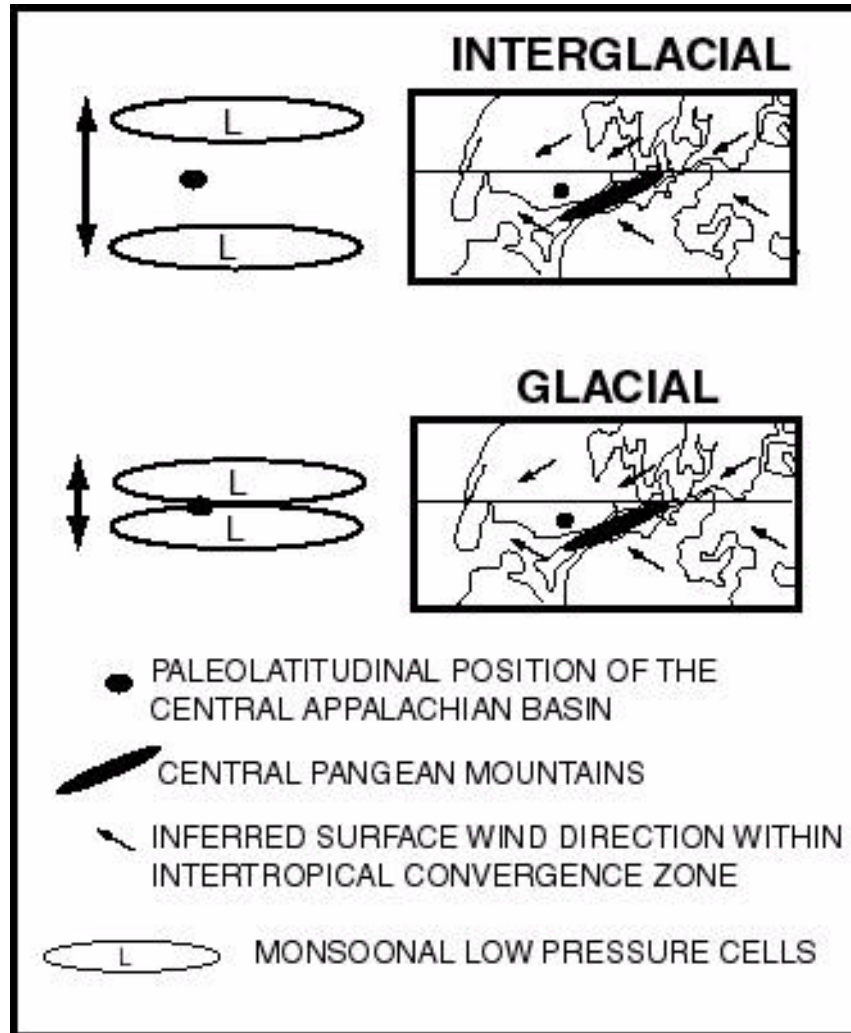
Figure 3.7- Model showing relative latitudinal shift of monsoonal low pressure systems (L). Interglacial times that favor strong seasonal heating may result in highly seasonal rainfall in subtropical and equatorial latitudes. A narrowing of climatic zones as a result of ice sheet formation (glacials) may diminish the monsoonal effect as seasonal heating is restricted to near equatorial latitudes.

LONG TERM PALEOCLIMATIC TREND

The study interval is located stratigraphically between middle Mississippian strata that were deposited under semiarid climatic conditions and Pennsylvanian coal-bearing strata that filled the basin during everwet climatic conditions (Cecil, 1990). This long-term climatic change has been ascribed to the northward drift of the Appalachian region into a wet climatic zone that is inferred to have been present in the Pangean equatorial belt (Cecil, 1990). Pennsylvanian paleosols that can be traced across North America indicate that climatic conditions became significantly dryer and more seasonal across the subequatorial belt from the Appalachian to Western Interior basins (Cecil et al., 1994). This trend may reflect the tendency for equatorial wet belts to be widest along the eastern margins of continents (Hay et al., 1990; Mack and James, 1994). However, long-term climatic change via paleolatitudinal drift is somewhat inconsistent

with sedimentologic evidence which indicates that seasonal climates were characteristic of Pangean equatorial latitudes (cf. Wright, 1990; Parrish, 1993).

As an alternative to paleolatitudinal drift through stable climatic belts, it is possible that the onset of everwet conditions (early Pennsylvanian) may reflect increased intensity of climatic change associated with glacial-interglacial episodes. Records of ice-rafted deposits (Frakes et al., 1992) and paleontological evidence (Raymond et al., 1989, Kelley et al., 1990) indicate that significant global cooling and attendant glaciation occurred in the late Namurian/ Westphalian (315-306 Ma). The Namurian A/B boundary (the Mississippian-Pennsylvanian boundary in the Appalachians) was a time of equatorial warming as climatic zones contracted in response to glaciation (Ziegler et al., 1987; Kelly et al., 1990). However, affirmation of intensified glaciation as the primary control on equatorial precipitation is difficult because other studies suggest that ice-sheets were of significant (maximum?) extent during the middle Namurian (325 Ma; Gonzalez-Bonorino and Eyles, 1995) and the Sakmarian (270 Ma; Otto-Bliesner, 1993; Gastaldo et al., 1996).

A third factor that would have influenced rainfall patterns in the early Pennsylvanian is the emergence of the Alleghenian mountains. High mountains (>2000 m) deflect winds and, depending on their latitude and orientation, can significantly modify patterns of rainfall (Witzke, 1990; Hay et al. 1990; Mack and James, 1994). Mountain belts are generally wetter on their windward sides. The eastern side of the central Pangean mountains is inferred to have experienced high rates of precipitation sourced from Tethyan equatorial winds, whereas the Appalachian basin would have developed in a rain shadow in the lee of the emerging mountains (Thomas, 1989; Hoffman and Grotzinger, 1993). Such a topographic/ geographic configuration exists in present-day equatorial Africa, where moderately high mountains in the east permit only seasonal rainfall across much of Kenya and Tanzania (Rowley et al. 1985; Parrish, 1993).

As the central Pangean mountains rose in response to collision, they would have provided a high altitude heat source that favored the development of a large, low pressure cell. This process takes place over India, where the Himalaya/ Tibetan Plateau (at ~30 degrees north latitude) dramatically amplifies subtropical low-pressures during northern hemisphere summer (Rowley et al., 1985). In the case of the central Pangean mountains, year-round low pressure associated with the emerging equatorial highlands may have countered the effects of seasonal heating to decrease the intensity of intertropical monsoonal circulation (Rowley et al., 1985). Numerical models of Beaumont et al. (1988) and Slingerland and Furlong (1989) suggest that Alleghenian thrust loads emplaced during the Mississippian were accommodated by downwarp of thinned crust, and that maximim topographic relief (3.5-4.5 km) did not develop until the Permian. Nevertheless, climatic modelling by Hay et al. (1990) indicates that 2-3 km of relief is sufficient to greatly affect rainfall and wind patterns upon continents with cross-equatorial mountains along their eastern margin. Otto-Bliesner (1993) confirms that the presence of 3 km high equatorial mountains in the Middle Pennsylvanian (306 Ma) likely set up a trough of low pressure on the lee (west) side of the mountains. In this way, an orogenically-forced reversal in equatorial winds offers an explanation for wet conditions in the Appalachian Basin because westerly winds may have introduced moisture derived from the seaway in the west (Otto-Bliesner, 1993; Crowley; 1996).

CONCLUSIONS

1. Paleoclimatic indicators associated with fourth-order (~400 k.y.) depositional sequences in the Appalachian basin vary systematically and in accordance with a glacio-eustatic control on sequence development. Indicators of seasonality are prevalent within the aggradational/ progradational coastal plain deposits. Leached paleosols and coals that cap terrestrial highstand facies and occur within incised valley-fill deposits are interpreted as marking times of reduced seasonality during periods of eustatic fall and lowstand. A monsoonal model for global atmospheric circulation may explain this paleoclimatic record which suggests that interglacial (highstand) periods were characterized by broad shifts in seasonal moisture across the tropical and equatorial latitudes. Seasonal latitudinal shifts in moisture belts were confined to near-equatorial latitudes during glacial episodes (lowstands).

2. The Late Mississippian depositional/ climatic cyclothems in the Appalachian basin are part of a long-term shift toward wetter climatic conditions. The seasonal character of Late Mississippian climates may have been modified by the early Pennsylvanian in response to northward paleolatitudinal drift, intensified glaciation, or orographic effects that decreased the intensity of monsoonal circulation in the lee of rising equatorial mountains.

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APPENDIX

PRINCIPAL EXPOSURES/ MEASURED SECTIONS

1. Route 102, Tazewell County, Virginia. Exposure of the lower Red member (Hinton Formation) through the Glady Fork Member of the Bluestone Formation between Bluefield and Hales Bottom.

2. Route 460, Mercer County, West Virginia. Discontinuous exposure of the Stony Gap Member (Hinton Formation) through the Pride Shale Member (Bluestone Formation) between Oakvale and Princeton.

3. Interstate 77: Upper Hinton (mid-"Ravencliff" interval) through Bramwell Member of the Bluestone Formation exposed from Brush Creek (west of Athens) northwest along I-77 to the U. S. Rt. 19 overpass.

4. Route 20: Summers County, West Virginia. Stony Gap Member (Hinton Formation) through the Princeton Formation exposed between Hinton and Pipestem, West Virginia.

4. Interstate 64: Stony Gap Member (Hinton Formation) through Red Member of the Bluestone Formation exposed from Sandstone (Summers County) westward along Interstate 64.

WELL LOG DATABASE

Logan County
1096 1139 1154

McDowell County
212 246 273 275 281 375 404 456 459 509 588 635 639 647 648 664 690 704 730
742 743 756 770 772 775 783 789 792 854

Mercer County
14 15 21 22 23 25 30 31 32 34 35 36 40 41 42 44 45 46 48 49 51 52 58 59 61 62
69 70 73 78 91 106 107 109 110 112 120 121 123 147 148 150 152 153 160 173

Mingo County
916

Raleigh County   **Tazewell County (Virginia)**

294 300 301 406 475 533 550 619 1184   TA-27

Wayne County
1697 1382

Wyoming County
762 764 789 792 796 825 841 1080 1146 1380 1384 Pocahontas # 12

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SELECTED PUBLICATIONS :

Miller, D. J., Adkins, R.M., and Eriksson, K. A., 1995, Sedimentation rates of prodeltaic and delta front tidal rhythmites in the central Appalachian basin: Abstracts with Programs, Geological Society of America Annual Meeting, v. 27 (6), p. 324.

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