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Abstract

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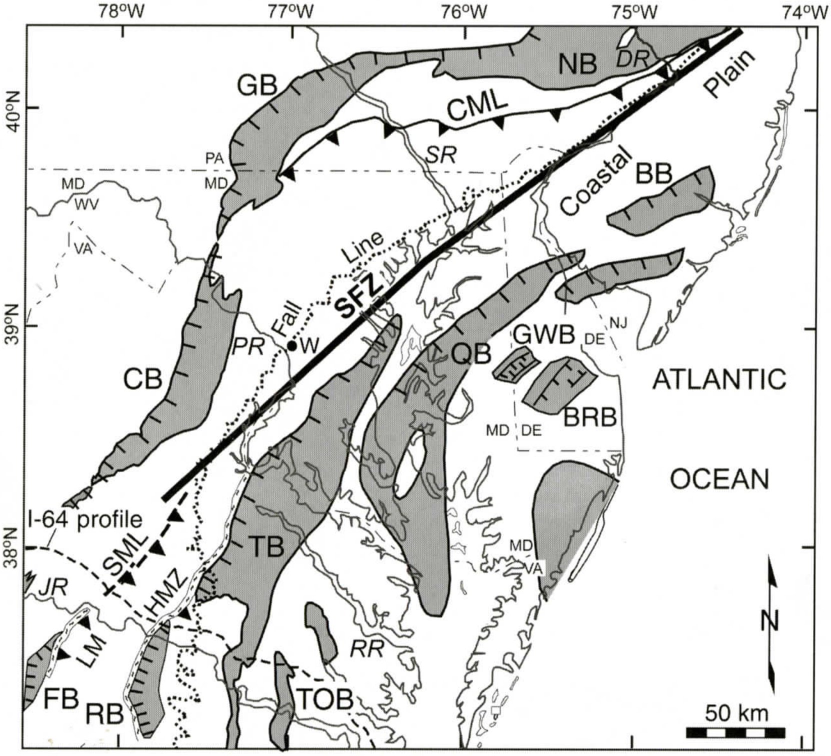

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Relationship of the Stafford Fault Zone to the Right-Stepping Bends of the Potomac, Susquehanna, and Delaware Rivers and Related Upstream Incision Along the U.S. Mid-Atlantic Fall Line

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ABSTRACT

River geomorphology near the U.S. mid-Atlantic Fall Line, as well as geologic and geophysical data, support previous hypotheses that the right-stepping bends of the Potomac, Susquehanna, and Delaware rivers, and upstream incision are associated with a northeastward extension of the mapped Stafford fault zone in north-central Virginia, yielding a total length of 450 km for the fault zone. Late Miocene to Pleistocene up-to-the-west displacement on the steep, west-dipping Stafford fault system, possibly combined with other processes such as flexural isostatic deformation and sea level lowering, produced the stream incisions southwest of the Delaware River. The source of the up-to-the-west displacement on the Stafford fault is likely from two superimposed processes: 1) oblique, up-to-the-west dextral fault slip driven by the regional northeast–southwest-oriented maximum horizontal compressive stress field, and 2) up-to-the-west displacement caused by isostatic uplift of the Appalachians to the west in combination with sediment loading in the Salisbury embayment to the east. The river bends, in contrast, were likely produced by termination of down-to-the-southwest tilting of the Salisbury embayment to the west against the Stafford fault zone. The source of the tilting likely results from uplift at the South New Jersey arch located northeast of the river bends. The abrupt termination of regional tilting and the linear nature of the Stafford fault zone imply that it is a steeply dipping strike-slip fault that has decoupled the crust east of the fault from that to the west. Low level seismicity coincides with much of the Stafford fault system between Maryland and New York City, suggesting a causal relationship. Because of the large population along the fault system between north-central Virginia and New York City, further studies are needed to determine the level of seismic hazard that it presents.

INTRODUCTION

Many investigators have proposed that late Cenozoic activity on a NNE-trending fault zone produced the right-stepping bends of the Potomac, Susquehanna, and Delaware rivers and adjacent upstream incision along the U.S. mid-Atlantic Fall Line (e.g., McGee, 1888; Hobbs, 1904; Higgins and others, 1974; Mixon and Newell, 1977) (Figure 1). Mixon and Newell (1977), using drill-hole and trench data, identified an ~55-km-long buried fault zone along the river bend trend (RBT of Figure 1) southwest of Washington, DC, that they named the Stafford fault system (SF on Figure 1). The Stafford fault system predominantly consists of en echelon fault segments dipping steeply to the northwest (Mixon and Newell, 1977). The segments predominantly display west-side-up offsets of about 30 to 75 m that decrease upward through the Coastal Plain sediments, consistent with intermittent differential uplift of the Piedmont. Slickenside orientations and a stereo net analysis of subsidiary normal and reverse faults exposed in trenches across the Dumfries segment of the Stafford fault system in Virginia indicate a dextral strike-slip component (page 438 of Mixon and Newell, 1977; Newell and others, 1978). These results, combined with small (< 1 m) reverse offsets of Pliocene- to Pleistocene-

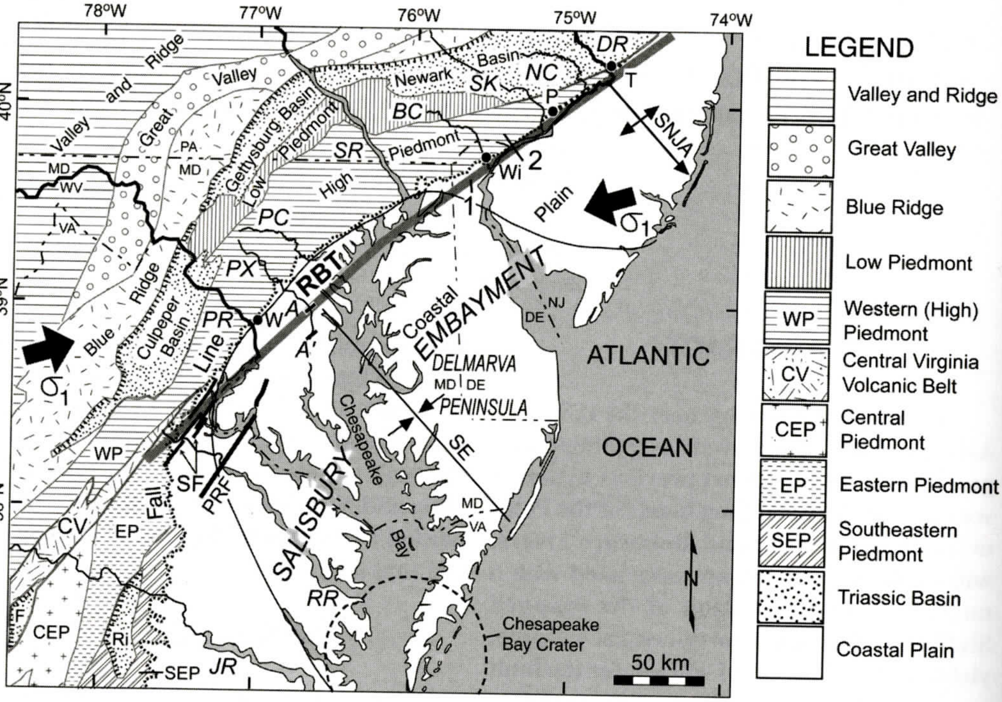


Figure 1. Location of river bend trend (RBT, thick gray line) that defines the location of the Stafford fault zone. Stafford fault system (SF) of Mixon and others (1992) denoted by thick black lines. AA' denotes portion of Patuxent River valley with down-to-the-southwest slip-off terraces in Figure 5. Locations 1 and 2 are other areas where Cretaceous to Cenozoic faulting have been documented along the RBT (see text). Other features shown are Fall Line (thin dotted line), South New Jersey arch (SNJA, Volkert and others, 1996), and Chesapeake Bay impact crater (thin dashed circular outline, Poag, 1997). Regional maximum horizontal compressive stress, σ_1, denoted by arrows (Zoback and Zoback, 1991). Approximate location of depocenter of Salisbury embayment (SE) also shown. Black contour in the Coastal Plain delineates the Salisbury embayment and South New Jersey arch. Physiographic provinces taken from Mixon and others (1989), Geologic Map of Pennsylvania (1990), Geologic Map of Virginia (1993), and Plank and others (2000). PRF is Port Royal fault. Triassic basins: F, Farmville; Ri, Richmond; and other dotted patterns. Rivers from south to north: JR, James; RR, Rappahannock; PR, Potomac; PX, Patuxent; PC, Patapsco; SR, Susquehanna; BC, Brandywine Creek; SK, Schuylkill; NC, Neshaminy Creek; DR, Delaware. Cities from southwest to northeast near the RBT: W, Washington, DC; Wi, Wilmington, DE; P, Philadelphia, PA; T, Trenton, NJ.

age stream-terrace deposits (Mixon and others, 1992), indicate that the fault zone periodically has undergone compressional deformation since at least the Early Cretaceous (110 Ma) and was active as recently as the latest Pliocene to early Pleistocene. Higgins and others (1974) and Mixon and others (1992) postulated that the fault system continues north-northeast along the RBT to Long Island, yielding a total length of about 450 km. This hypothesis is supported by evidence for Cenozoic faulting to the northeast along the Susquehanna and Delaware river bends (e.g., Thompson, 1978, 1979; McKenna and others, 1999a, 1999b) (Figure 1).

Based on gradients of larger rivers traversing the mid-Atlantic Piedmont, Reed (1981) and Hack (1982) attributed the Fall Line incision to a combination of Plio-Pleistocene sea-level regressions and differential uplift of the Piedmont along a narrow flexure or fault zone. Pazzaglia and Gardner (1994, 2000), in contrast, attributed the faulting and Fall Line incision to a combination of Plio-Pleistocene sea level regressions and post-Oligocene uplift (last 15 million years) along a convex-up flexural hinge zone produced by denudation to the west and sediment loading offshore and in the Salisbury embayment. They based their conclusion on a two-dimensional flexural isostatic geodynamic model for the mid-Atlantic margin that they derived from petrography-based lithostratigraphic correlations of Susquehanna River terraces with marine facies in the embayment.

Based on stream gradients in the Coastal Plain and Appalachian provinces and ages of Piedmont-derived sediments in the Coastal Plain of Virginia, Weems (1998) concluded that the Fall Line in northern Virginia, which he referred to as the Tidewater Fall Line, resulted mainly from two major episodes of uplift to the west. These uplifts occurred between 5 and 10 million years ago and within the last 2 million years.

Despite these and other efforts, the cause of the river bends and incision has remained enigmatic for several reasons. First, most river valleys are drowned below the mid-Atlantic Fall Line, thus neotectonic studies using fluvial geomorphology are mainly limited to the Piedmont. Second, seismicity recorded during the past 20 years has been sparse, rather than along well-defined trends (e.g., Bollinger and others, 1991). Third, faulting along most of the RBT cannot be observed directly because the Coastal Plain sediments there are up to a few hundred meters thick; faulting beneath thick unconsolidated sediments commonly results in folding of overlying strata rather than surface ruptures (Stein and Yeats, 1989). Fourth, very low Cenozoic slip rates on faults in the eastern U.S. (Prowell, 1988; Gardner, 1989) result in small displacements, thus making it difficult to identify fault trends. Finally, if the incision and bends were produced by a basement fault rooted beneath the Blue Ridge-Piedmont allochthon, the relatively high magnetization and 7- to 15-km-thickness of the allochthon would make its detection difficult using magnetics and gravity techniques.

My objective herein is to evaluate the various hypotheses by further investigating smaller rivers traversing the outer (eastern) Piedmont and inner (western) Coastal Plain of the mid-Atlantic area and then to integrate these results with available geologic, geophysical, and seismicity data. My results suggest that late Cenozoic up-to-the-west displacement on the Stafford fault zone, possibly combined with flexural deformation, is the primary cause of the Fall Line incision in the mid-Atlantic states. The river bends were produced by the western boundary of down-to-the-southwest tilting of the Salisbury embayment to the west against the Stafford fault zone.

FLUVIAL GEOMORPHOLOGY OF THE MID-ATLANTIC PIEDMONT AND COASTAL PLAIN

Rivers traversing the mid-Atlantic Piedmont flow over various types of Proterozoic to Paleozoic igneous and metamorphic rocks, as well as over sedimentary rocks and igneous dikes of the exposed early Mesozoic Newark rift basins (Figure 1). In the easternmost Piedmont the valleys are deeply incised with narrow terraces, while farther upstream they are typically broader with well-developed terraces and sometimes flood-plains (Hack, 1982). The incision developed contemporaneously with development of the Fall Line starting in late Miocene time by warping, faulting, and differential erosion (Cleaves, 1989); the Fall Line is the linear trace of rapids and falls along various rivers near the western edge of the Atlantic Coastal Plain province (Weems, 1998).

Most Piedmont rivers are drowned downstream of the mid-Atlantic Fall Line, thus forming estuaries, such as the Chesapeake and Delaware Bays. The largest rivers (the Potomac, Susquehanna, and Delaware) bend abruptly to the southwest along a linear trend that several authors have suggested may be fault-related (e.g., Higgins and others, 1974; Mixon and others, 1992). These large right-stepping bends decrease in amplitude to the southwest away from the South New Jersey arch; the sizes of the bends are about 100 km, 50 km, and 35 km for the Delaware, Susquehanna, and Potomac rivers (Figure 1). Based on sediment distribution patterns, Poag and Sevon (1989) postulated that the river bends began to form during middle Miocene time. Owens and Minard (1979) and Stanford (1993), in contrast, postulated that the Delaware paleovalley in the Coastal Plain was temporarily occupied by other large rivers, such as an ancient Hudson, flowing from the north-northeast across central New Jersey during late Miocene and Pliocene time. In either case, the larger rivers, including the ancient Hudson, generally migrated to the southwest as they flowed southeast across the mid-Atlantic Coastal Plain during mid-Miocene to Pliocene sea-level low stands, leaving behind progressively older sediments to the northeast (Hack, 1955; Schlee, 1957; Owens and Minard, 1979; Pazzaglia, 1996).

Down-to-the-southwest tilting within the Salisbury embayment and on the southern flank of the South New Jersey arch has been postulated to be the primary cause of the southwest migration (Newell and Rader, 1982; Pazzaglia, 1993). Subsidence within the Eocene-age (~35 Ma) Chesapeake Bay impact crater (Figure 1) and Pleistocene longshore drift also caused the Susquehanna River to migrate southwestward during late Pleistocene sea level low stands, but mainly in the outer Coastal Plain (Colman and others, 1990; Powars and Bruce, 1999). Until now, no one has presented a convincing tectonic model to explain fully the interrelationships of the southwest river migration, the river bends, and the Fall Line incision.

METHODS

I have examined the geomorphology of river valleys traversing the mid-Atlantic Fall Line to determine the origin of both the river bends and incisions. To do this, I separately evaluated changes in bedrock type, sea level fluctuations, longshore drift, deformation along a convex-up flexural hinge zone, and deformation on the Stafford fault zone. The river investigation was mainly restricted to the Piedmont because most rivers traversing the Piedmont are drowned just downstream of the mid-Atlantic Fall Line.

Methods for the Piedmont River Study

For the Piedmont river study I investigated changes in bedrock type, gradients, and incision patterns along smaller river valleys traversing the eastern Piedmont between the Potomac and Delaware rivers to help determine the cause of the Fall Line incision. I used primarily 1:24 000- to 1:500 000-scale topographic and surficial geologic maps. Longitudinal profiles, combined with bedrock geology taken from detailed 1:24 000 maps, were used to locate and quantify knickpoints and to infer differences in erodibility in the valley floors of rivers along the Fall Line.

Methods for Patuxent River study in the Coastal Plain

The Patuxent River valley in Maryland (PX in Figure 1) provides a unique opportunity to search for evidence of neotectonism within the Coastal Plain where it traverses the RBT because it is the only river valley that is not drowned along the inner (western) edge of the Salisbury embayment (Figure 1).

To search for local uplift along the Coastal Plain section of the Patuxent valley, I evaluated local changes in river geomorphology, hereafter referred to as river anomalies. River anomalies are produced by changes in channel slope, which affect stream power (product of stream discharge and channel slope; Schumm, 1986). Uplift-related anomalies include channel incision, upwarped terraces, low channel sinuosity across the uplift, and increased channel sinuosity downstream and sometimes upstream (Ouchi, 1985; Schumm, 1986). Aggradation also occurs upstream and downstream from the uplift because of lowered stream power in those reaches.

To search for sinuosity patterns that might be from uplift, a graph of sinuosities (ratio of channel length to valley length) was calculated for 1-km-long valley segments, which is about twice the average meander wavelength of the Patuxent River. Longitudinal profiles and stream length-gradient indices were used to detect and

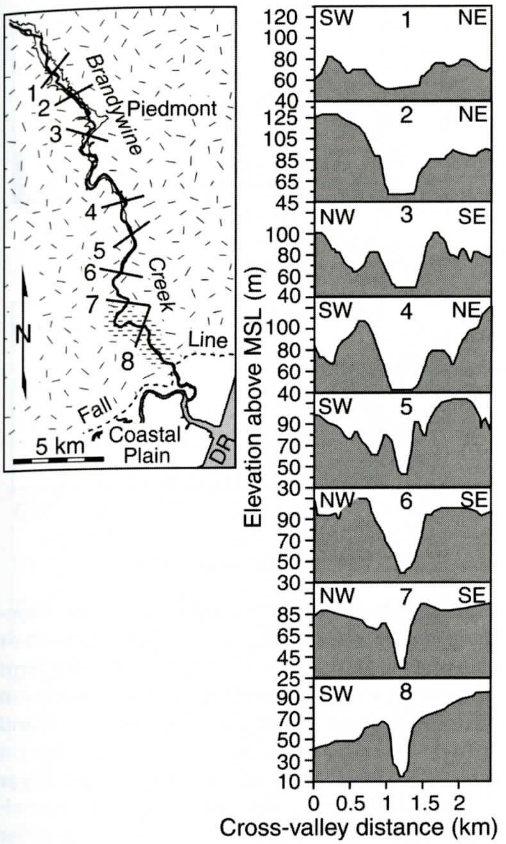


Figure 2. Profiles across valley of Brandywine Creek. Note the V-shaped, incised profiles (7-8) in the outer Piedmont in contrast to the wider valley upstream (profiles 1-4) characterized by a broader valley floor. Area of short dashes in index map denotes increased incision in the outer Piedmont. DR is Delaware River.

quantify gradient changes across the RBT. The stream length-gradient index for a particular reach is defined as the product of channel slope at that particular reach and the total channel length from that reach to the highest channel elevation (Hack, 1973).

Using Cross-valley Profiles to Distinguish Between Buried Faulting and Deformation Along a Convex-up Flexural Hinge Zone

I also sought cross-valley anomalies that deformation along a hinge zone could not produce. One type is a local cross-valley tilt along a buried fault zone that locally causes lateral river migration, development of unpaired slip-off terraces, and a local curve in the river valley (Figure 5A of Marple and Talwani, 2000). The other type of cross-valley anomaly is an abrupt change from a relatively symmetric cross-valley shape upstream to a cross-valley asymmetry downstream, suggesting cross-valley tilting, often with the river laterally offset (Marple, 1998; Figure 5B of Marple and Talwani, 2000). This anomaly suggests the presence of a laterally and vertically through-going fault along which tilting on the flank of a regional arch, such as the South New Jersey arch, stops to the west. I investigated the large river bends near the mid-Atlantic Fall Line to determine if they were produced by this mechanism.

INVESTIGATION OF INCISION AND LARGE RIVER BENDS NEAR THE MID-ATLANTIC FALL LINE

Investigation of Rivers Traversing the Piedmont

The Piedmont streams that I investigated are (from southwest to northeast) the Patuxent River, Patapsco River, Brandywine Creek, Schuylkill River, and Neshaminy Creek (Figure 1). The Patuxent and Patapsco rivers and Brandywine Creek are deeply incised in the outer (eastern) Piedmont, but farther upstream in the inner Piedmont their valleys widen, sometimes with flood-plains, indicating aggradation. The valley floor of Brandywine Creek, for example, upstream from the incision is as much as 0.5 km wide, displays meander cutoffs, and is swampy in some areas (Figure 2). The Schuylkill River and Neshaminy Creek, in contrast, are not deeply incised in the outer Piedmont. Longitudinal profiles of the more deeply incised rivers have large knickpoints (~15 to 44 m) near the Fall Line while the Schuylkill River has a small knickpoint of only about 6 m and the Neshaminy Creek profile doesn't have a knickpoint (Figure 3). The change in incision depth is also reflected in the Fall Line, which forms a topographic scarp southwest of Wilmington, but not to the northeast (Hack, 1982).

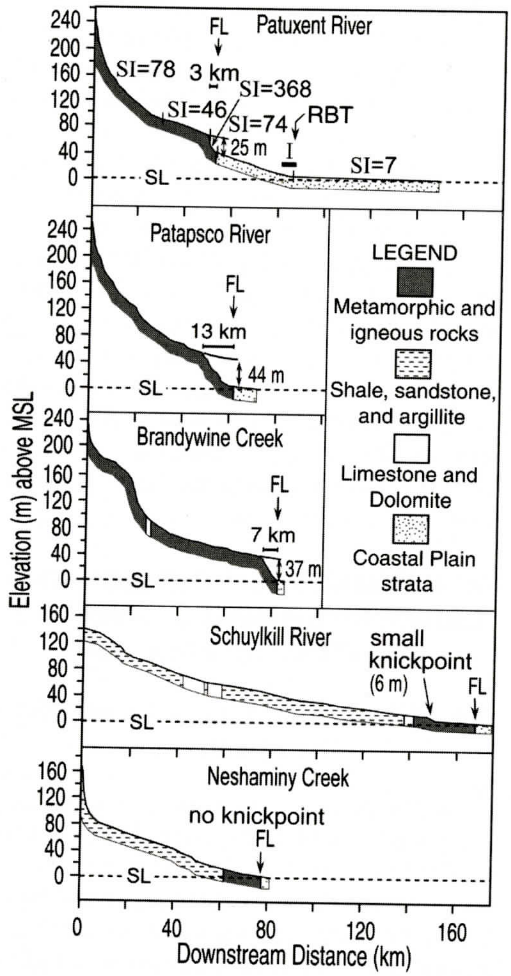


Figure 3. Longitudinal profiles of Patuxent, Patapsco, and Schuylkill rivers, and Brandywine and Neshaminy creeks compared with bedrock geology (profiles from south to north). SI denotes stream length-gradient indices above different segments of Patuxent profile. Length (km) of incised reach in the outer Piedmont denoted by horizontal bar above profiles. Incision depth in meters. FL, Fall Line; I, incision (Patuxent profile); RBT, river bend trend (Patuxent profile). Schuylkill profile is downstream from Reading, PA. See Figure 1 for river locations.

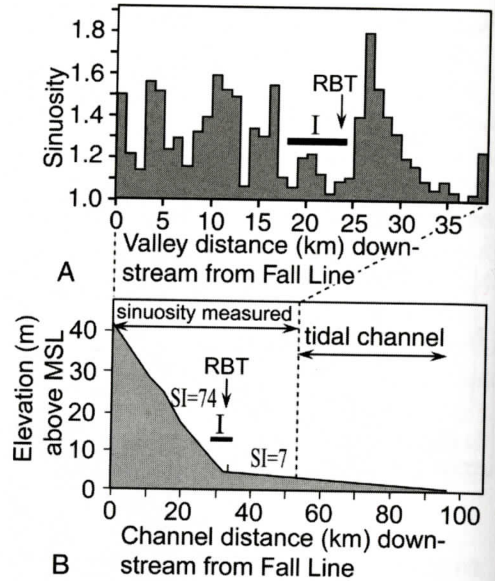
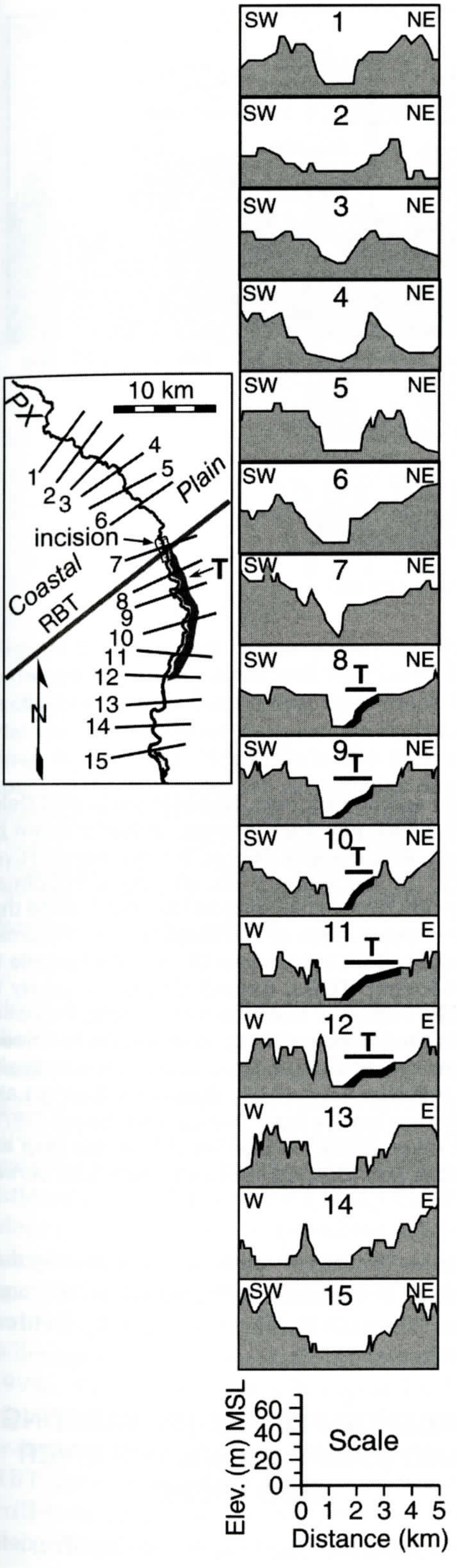


Figure 4. A. Sinuosities (averaged over 1-km-long valley segments) of the Patuxent River channel downstream of the Fall Line and upstream of tidal channel. Area of incision denoted by horizontal line. RBT is river bend trend. B. Longitudinal profile and stream length-gradient indices (SI) of Patuxent River corresponding to same reach in A. Horizontal-axis scales (distance) in graphs A and B differ because graph A distance was measured along the center of the valley whereas graph B distance was measured along the channel.

Coastal Plain Study of Patuxent River valley

The Patuxent River valley in the Coastal Plain displays a variety of anomalies along the RBT in the inner (western) Coastal Plain. Instead of a smooth concave-up longitudinal profile typical of Coastal Plain rivers (Marple and Talwani, 2000), the Patuxent River profile displays a sudden gradient change where it traverses the RBT within the Coastal Plain province, rather than at the Fall Line. The stream length-gradient index changes abruptly from about 74 (relatively steep) upstream of the RBT to 7 downstream (Figures 3 and 4B). The channel is more deeply incised (~3 m) for ~7 km near this gradient change (Figure 4). Sinuosity is low (1.03-1.2) for about 8 km upstream of the RBT while downstream the sinuosity increases to as much as 1.8 (Figure 4A).

The Patuxent valley also displays down-to-the-southwest, late to middle Pleistocene-age slip-off terraces (Glaser and Hansen, 1973; Glaser, 1984) for about 13 km downstream of the RBT (McCartan, 1996) (A-A' on Figure 1, Fig-



ure 5). The fact that the channel does not flow preferentially toward either side of the valley floor here indicates that the cross-valley tilt that produced these terraces ended before development of the modern (Holocene) valley.

Other Cross-valley Anomalies in the Inner (Western) Coastal Plain

River cross-valley profiles show an abrupt change across the RBT. A cross-valley profile across all three major rivers just downstream of the RBT generally displays down-to-the-southwest cross-valley tilting (Figure 6). Sedimentologic studies indicate that the major rivers migrated southwestward downstream of the RBT mainly during late Miocene time, leaving behind sediments that are progressively older to the northeast (e.g., Schlee, 1957; Owens and Minard, 1979; Pazzaglia, 1993) (Figure 6). The Susquehanna River, for example, migrated about 45 km to the southwest during late Miocene time, leaving behind Bryn Mawr fluvial deposits that are progressively older to the northeast (Pazzaglia, 1993). Upstream of the RBT, in contrast, river valleys display no evidence for cross-valley tilting or river migration to the southwest. The Potomac and Susquehanna Rivers, for example, in the outer Piedmont and inner Coastal Plain west of the RBT have relatively symmetric cross-valley shapes with Miocene- to Pleistocene-age terraces on both sides of their valleys (Figure 7, Pazzaglia and Gardner, 1993) while the Delaware valley displays terraces on its southwest side (Figure 8). Smaller valleys in the outer Piedmont, such as that of Brandywine Creek (Figure 2), also do not display a cross-valley tilt.

Another anomalous feature associated with

Figure 5. Profiles across Patuxent River valley. RBT (gray line in index map) denotes river bend trend and postulated location of Stafford fault zone to the northeast. Incision denoted by gray area along Patuxent River (PX) in index map. Location of down-to-the-southwest slip-off terrace (middle to late Pleistocene, Glaser and Hansen, 1973; Glaser, 1984) denoted by shaded area (T) in index map and horizontal line labeled T in cross sections.

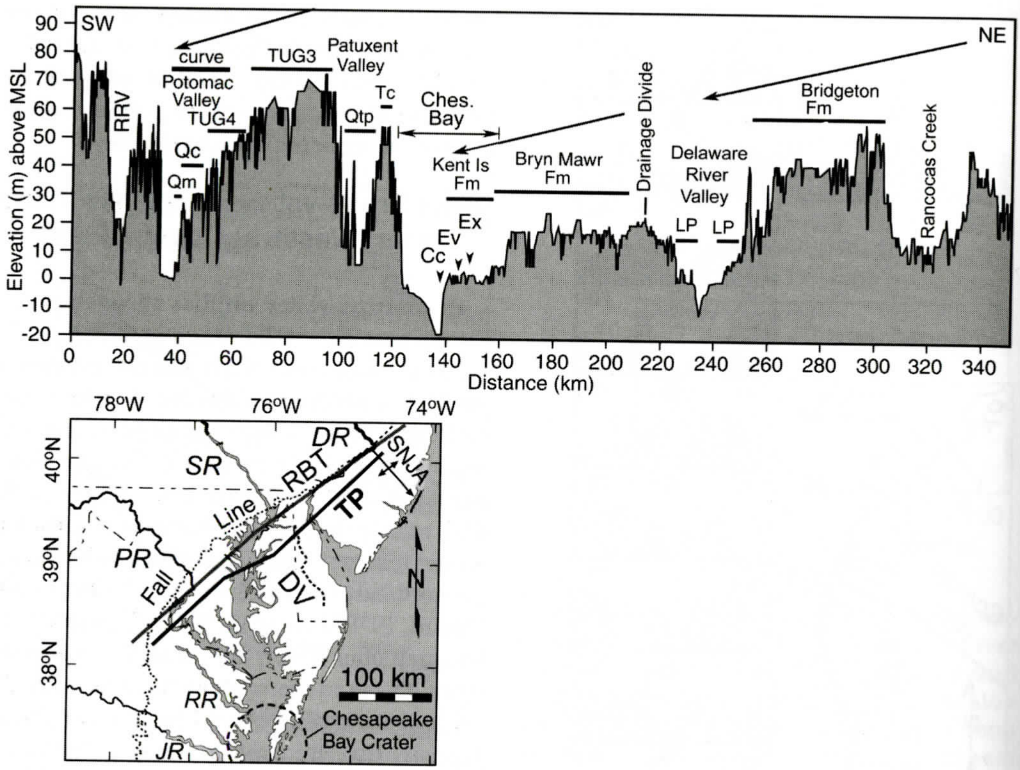


Figure 6. Topographic profile (labeled TP in index map) across Potomac, Susquehanna, and Delaware rivers just downstream of river bend trend (RBT, solid gray line). Arrows denote amount by which the major rivers are deflected to the southwest. Arrowheads denote Susquehanna River paleochannels: Cape Charles (Cc, 18 ka), Eastville (Ev, 150 ka), Exmore (Ex, 200-400 ka) (Colman and others, 1990). Fluvial terrace locations denoted by horizontal lines in cross sections. Note that the terrace sediments become progressively younger toward each of the major rivers. Potomac valley terraces: Qm, Maryland Point Fm (late Pleistocene); Qc, Chicamuxen Church Fm (middle to early Pleistocene); TUG4, Upland Gravel Fm (late Pliocene); TUG3, Upland Gravel Fm (early to upper Pliocene) (Glaser, 1978, 1984; McCartan, 1989). Patuxent valley terraces: Qtp, Patuxent Pleistocene terraces; Tc, Calvert Fm (middle Miocene) (from Glaser, 1976). Kent Isl. Fm in Chesapeake Bay is middle to late Pleistocene (Owens and Denny, 1979). Bryn Mawr subsurface deposits are from Pazzaglia (1993). Late Pleistocene terraces (LP, Van Sciver Lake Beds and Spring Lake Beds Formations) of Delaware valley and Bridgeton Formation are from Owens and Minard (1979) and Newell and others (1995). RRV is Rappahannock River valley. DV and SNJA in index map are drainage divide between Delaware River and Chesapeake Bay (dashed line) and South New Jersey arch. Index map modified from Figure 1.

cross-valley tilting along the inner Coastal Plain is the large southwest-convex curve in the Potomac River valley just downstream of the RBT (Figure 1). The terrain inside the curve is tilted down-to-the-southwest with unpaired terraces on the northeast side of the valley (Figure 6). Sedimentologic studies (e.g., Schlee, 1957) suggest that the curve and cross-valley tilt formed by river migration to the southwest since late Miocene time. This curve is superposed on the area of southwest migration by the ancestral Potomac River between the RBT and outer Coastal Plain postulated by Schlee (1957).

EVIDENCE FOR BURIED FAULTING AND UPLIFT ALONG THE RIVER BEND TREND

Evidence for buried faulting and uplift exists

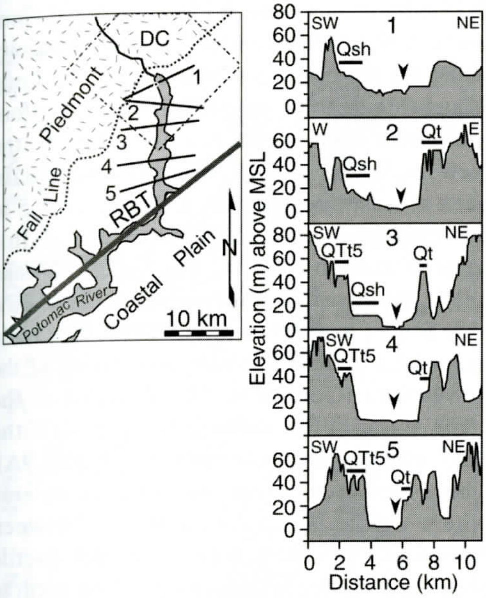


Figure 7. Potomac cross-valley profiles between the river curve and Fall Line. RBT (solid line) in index map is river bend trend. Fluvial terrace locations denoted by horizontal lines in cross sections. Note the lack of a southwest cross-valley tilt. Arrowheads denote channel location. Terrace units: QTt5, early Pleistocene; Qt, early Pleistocene; Qsh, middle Pleistocene (Froelich and Hack, 1975; Hack, 1977; Froelich, 1985). District of Columbia (DC) denoted by dashed outline.

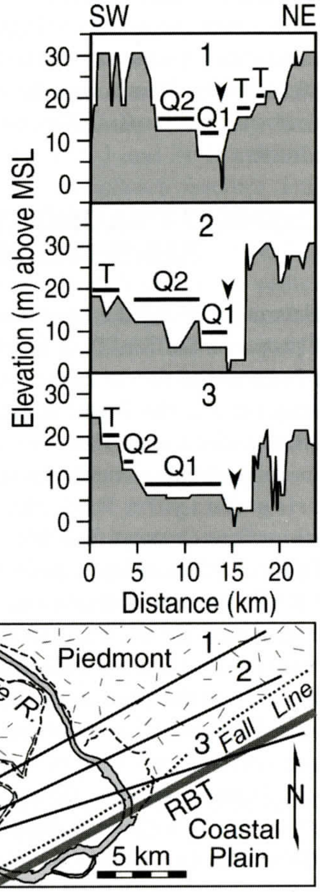


Figure 8. Delaware cross-valley profiles just upstream from river bend trend (RBT, solid line in index map). Horizontal lines in cross sections denote fluvial terrace locations. Arrowheads denote channel location. Terrace units: Q2, early Pleistocene; Q1, early Pleistocene (younger than Q2); T, older terraces, ages unknown (terrace ages and locations from Owens and Minard, 1975).

along various parts of the river bend trend (RBT), especially along its southwest end. The Stafford fault zone (SF on Figure 1) and north-northeast trending folds within Coastal Plain strata southwest of Washington, DC, (Mixon and others, 1992) are the strongest evidence for differential uplift and faulting along the RBT. More recently, a northwest-southeast-oriented 3-km-long seismic-reflection profile was imaged across the RBT just southwest of the Delaware River to investigate seismicity near Wilmington, Delaware (McKenna and others, 1999a, 1999b) (location 1 in Figure 1). This high-resolution profile suggests that Cretaceous sediments are extensively faulted along the RBT. Trenches up to 2.5 m deep and shallow drill-hole data revealed no surface faulting along this seismic line (McLaughlin and others, 2002). However, the Coastal Plain sediments in this area are hundreds of feet thick and the faulting, therefore, may not extend to the surface; either the faulting stopped before deposition of the shallower sediments or faulting at depth was muted by the Coastal Plain sediments as it propagated upward through them.

Based on aeromagnetic and drill-hole data, Thompson (1978, 1979, 1981) located two buried faults beneath the Delaware River bend between Wilmington and Philadelphia: the one to the west dips ~45° to 60°SE while the other is nearly vertical (location 2 in Figure 1). The pre-Cretaceous basement is offset down to the southeast some 45 m across the river bend, consistent with Cenozoic differential uplift of the Piedmont.

SEISMICITY ALONG THE RIVER BEND TREND

Seismicity along the river bend trend (RBT) and proposed Stafford fault is difficult to evaluate because of the regional scatter of small earthquakes on the Paleozoic thrust faults and Triassic basin faults. However, some observations suggest that some of the seismicity is occurring along the RBT. Several notable earthquakes (≥ modified Mercalli intensity [MMI IV]), for example, have occurred along the RBT between Wilmington, Delaware, and Trenton, New Jersey, since 1800 (Coffman and others, 1982) (Figure 9). The Wilmington area is the most active area along the RBT between central Virginia and central New Jersey; several dozen small events have occurred there since 1871 (Stefanie Baxter, Delaware Geological Survey, 1999 written comm.). The largest known event (MMI VII) occurred near Wilmington on October 9, 1871, toppling chimneys and breaking windows (NOAA, 1971). A study of the 1973 Wilmington (M 3.8, MMI V-VI) earthquake (Sbar and others, 1975) suggests that the earthquakes between Wilmington and Trenton are associated with a steep, northeast-trending, northwest-dipping fault zone along the Delaware River bend. Its fault plane solution and aftershock distribution indicate up-to-the-west reverse motion on a nearly vertical plane (northwest dip direction) striking N28°E, subparallel to the Delaware bend (Sbar and others, 1975) (Figure 9A). Its highest intensity isoseismals are elongate along the Delaware bend (Figure 9A). The up-to-the-west motion of the Wilmington earthquake is consistent with late Cenozoic differential uplift of the Piedmont and late Cenozoic offsets along the Stafford fault zone and faults at locations 1 and 2 of Figure 1.

Among the regional scatter of small (< M 5.2) earthquakes between Delaware and Connecticut is a northeast alignment of seismicity along the RBT between Wilmington and Long Island (Figure 9A), suggesting that the Stafford fault zone may extend northeastward to the New York City area. Historically, there have been at least three estimated magnitude 4.5 to 5.0 earthquakes at the northeast end of this trend in western Long Island (1737, 1783, and 1884), the largest of which occurred in 1884 (M ~ 5, Seeber and Armbruster, 1988). The main axis of the highest intensity isoseismals of the 1884 event (Gordon, 1988), like that of the 1973 Wilmington earthquake, parallels the RBT and alignment of seismicity (Figure 9A). These New York events occurred in an area where triangulation data acquired between 1862 and 1973 show high horizontal ductile shear strain in the lower crust trending north to northeast (Zoback and others, 1985; Prescott and others, 1987). Other nearby northeast-trending structures include the New York Bight fault and the early Paleozoic suture called Cameron's–Martic Line near the eastern side of the Newark Triassic basin (Figure 9). These features, however, lie south and west, respectively, of the 1884 epicenter on Long Island. Also, rocks along Cameron's-Martic Line are fully ductile and ductilely folded, making that feature an unlikely candidate for brittle reactivation.

At its southwest end is another historical earthquake that might be associated with the Stafford fault zone. Here, the northeast-trending isoseismals of the larger of two earthquakes that occurred near Fredericksburg, Virginia in 1971 (M_L 3.5, MMI IV, Bollinger, 1971) closely coincide with the RBT (Figure 9C). Although this coincidence is compelling, no focal mechanisms were derived for these earthquakes to determine if they occurred on the Stafford fault zone.

DISCUSSION

One process that I evaluated to determine the cause of both the river bends and incision upstream along the mid-Atlantic Fall Line was bedrock erodibility. Bedrock erodibility, though, is an unlikely cause of the incision be-

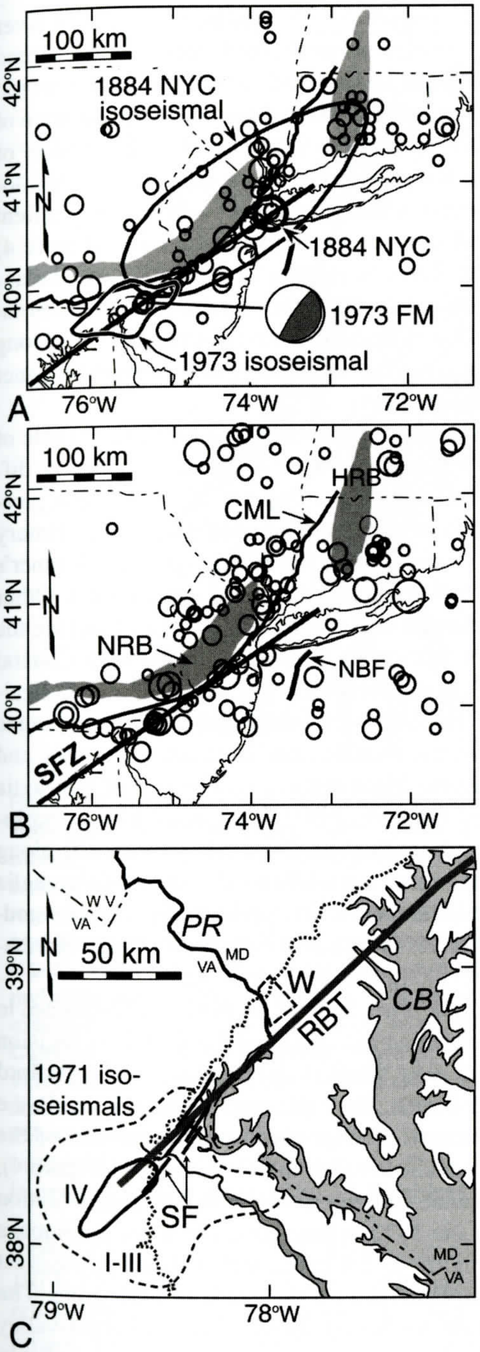


Figure 9. Historical and network seismicity (open circles) compared to the Stafford fault zone (SFZ, black line), Newark and Hartford rift basins (NRB, HRB, gray patterns), Cameron's-Martic line (CML, thick black line), and New York Bight fault (NBF, thick black line). A. Epicenters for period 1815-1974, 3.0≤ m ≤5.2, and focal mechanism (compressional quadrant shaded) for 1973 Wilmington, Delaware earthquake (from Sbar and others, 1975). Highest intensity isoseismals of 1973 Wilmington and 1884 NY earthquakes (closed contours) are from Sbar and others (1975) and Gordon (1988). B. Epicenters for period 1975-1994, 2.0≤ m ≤4.7. Note the alignment of seismicity along the Stafford fault zone southwest of Long Island in both maps. Epicenters near Wilmington in map B are from Stefanie Baxter (Delaware Geological Survey, 1999 written commun.). Modified from Kafka and Miller (1996). C. MM I-III and IV isoseismals (closed contours) of the larger of the two 1971 Fredericksburg, Virginia earthquakes (Bollinger, 1971). RBT (thick gray line) denotes river bend trend defining the location of the extended Stafford fault zone. CB, Chesapeake Bay; PR, Potomac River; W, Washington, DC.

cause, although the more erodible strata of the Newark basin are closer to the Fall Line north of Brandywine Creek, Piedmont rivers northeast of Brandywine Creek flow over crystalline rocks that are lithologically similar to those to the southwest (Figures 1 and 3). Nor is there a noticeable change in bedrock fracture density along the Fall Line that could have produced the different depths of incision between the Patuxent and Delaware rivers. Changes in sediment type also cannot explain the river bends and cross-valley tilt downstream. The Coastal Plain sediment along and downstream from the river bends within the embayment vary in lithology and do not show a consistent southwest dip of bedding that could have consistently caused the paleochannels to migrate to the southwest.

Pleistocene sea level regressions also cannot fully explain the incision. Most of the rivers that I investigated are similar in size (based on discharge and drainage basin area), flow over similar bedrock types, and respond to the same base level (i.e., Chesapeake Bay and Delaware River). For these reasons, they should display similar-sized knickpoints across the Fall Line, which is not true (Figure 3). Nor can sea level regressions explain the southwest river migration and cross-valley tilting downstream of the RBT along the Potomac, Patuxent, Susquehanna, and Delaware rivers (Figures 5 and 6).

The cross-valley migration of ancient rivers in the Salisbury embayment cannot be attributed to differential subsidence within the embayment. The right-stepping bend of the Potomac River, for example, lies southwest of the embayment depocenter (Figure 1). Subsidence-induced tilt likely would have deflected the Potomac River northeastward toward the depocenter located near the central Delmarva Peninsula (Olsson and others, 1988) (Figure 1).

Baymouth shifting along shorelines sometimes deflects rivers (e.g., Gayes and others, 1992), but is not a likely cause of the large river bends. Sedimentologic studies in the Salisbury embayment (e.g., Hack, 1955; Owens and Minard, 1979) indicate that the southwest river migration associated with these river bends occurred during sea level low stands when the shoreline was in the outer Coastal Plain or continental shelf. With the shoreline so far to the east, even a large shift in a river's course by baymouth shifting could not have produced the river bends. For example, although the southwest growth of the Delmarva Peninsula and subsidence in the Chesapeake Bay impact crater (Figure 1) caused the late Pleistocene Susquehanna paleochannels to migrate southwestward across the outer Coastal Plain, their divergence from the original course gradually decreases to the northwest until they converge with the old channel without producing an abrupt bend (Colman and others, 1990).

Having evaluated nontectonic processes as possible causes of the Fall Line incision and river bends, I then investigated the possibility that late Cenozoic flexural isostatic deformation produced these features. However, flexural deformation also cannot explain most of the geomorphic observations, except perhaps the Fall Line incision. First, flexural isostatic deformation could not have caused the abrupt change in cross-valley shape of river valleys along the RBT, such as that along the Patuxent River valley (Figure 5). Secondly, most of the river anomalies occur abruptly along a linear trend whereas flexural isostatic deformation would be associated with a broader, nonlinear zone of deformation like that of the curved boundary of the Salisbury embayment, perhaps several kilometers wide. The abrupt changes in gradient and sinuosity of the Patuxent River (Figure 4) and their coincidence with the RBT and the linearity of the RBT, for example, are more consistent with up-to-the-west displacement on a long buried strike-slip fault system beneath the inner Coastal Plain.

Regarding the Fall Line incision, the role of any flexural isostatic deformation would be difficult to evaluate because of the possible presence of the Stafford fault zone as a primary tectonic feature, which Pazzaglia and Gardner's (1994) flexural model did not consider. Other potential sources of error in their model are the correlation of Piedmont straths with Coastal Plain sedimentary units, which do not physically trace to each other, and the large distance (several tens of kms) between the straths and Coastal Plain units (e.g., Figure 3 of Pazzaglia and Gardner, 1994). The overall steeper basinward gradient of Coastal Plain deposits along the inner Coastal Plain (Figure 3 of Pazzaglia and Gardner) is also problematic for their model and is more easily accounted for by deformation along a major fault zone.

The other tectonic process that I evaluated to determine the origin of the Fall Line incision and river bends is deformation on the Stafford fault zone. The Fall Line incision (Figure 3), the gradient change and sinuosity pattern of the Patuxent River in the Coastal Plain (Figure 4), and the steep up-to-the-west fault plane solution of the 1973 Wilmington, Delaware earthquake (Figure 9) are consistent with up-to-the-west displacement on the Stafford fault zone. The river bends and abrupt change in cross-valley shape of several river valleys are also probably fault related, but in a different way (see later section on river bends). The strongest case for late Cenozoic faulting is along the southern end of the RBT where the ~55-km-long Stafford fault zone and folded strata were mapped in the subsurface (e.g., Mixon and others, 1992). Al-

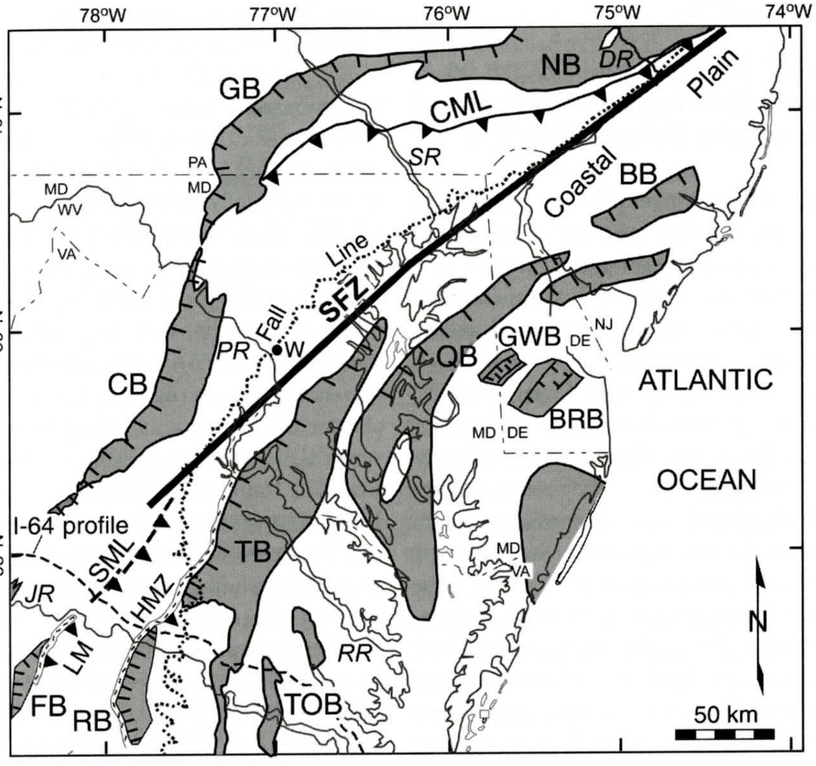


Figure 10. Map of north-central Virginia to southern New Jersey region showing the Stafford fault zone (SFZ) proposed herein and in previous studies (e.g., Higgins and others, 1974; Mixon and others, 1992), Paleozoic faults (labeled), and onshore buried and exposed early Mesozoic rift basins (shaded). The SFZ is linear and dips steeply to the northwest (see text) while the other regional faults are curved and dip southeastward, suggesting a cross-cutting relationship. Faults and basins taken from Mixon and others (1989, 1992), Benson (1992), Geologic Map of Virginia (1993), and Wilkes (1993). CML, Cameron's-Martic line; HMZ, Hylas mylonite zone; LM, Lakeside mylonite zone; SML, Spotsylvania magnetic lineament (SML is interpreted to be a northeast extension of the Lakeside mylonite zone, Mixon and others, 1992). Onshore Triassic basins: CB, Culpeper; GB, Gettysburg; NB, Newark; RB, Richmond; TB, Taylorsville; TOB, Toana; FB, Farmville; BB, Buena; QB, Queen Anne; GWB, Greenwood; BRB, Bridgeville; other unnamed basins also included (from Benson, 1992). Rivers from south to north: JR, James; RR, Rappahannock; PR, Potomac; SR, Susquehanna; DR, Delaware. Teeth and hachures along thrust faults and basin border faults denote direction of dip. The I-64 seismic-reflection profile of Pratt and others (1988) is denoted with a dashed line. W, Washington, DC.

though sparse, the drill-hole (Thompson, 1978, 1979, 1981), seismic-reflection (McKenna and others, 1999a, 1999b), and seismicity (Sbar and others, 1975) data to the northeast (Figures 1 and 9) suggest that the Stafford fault zone continues northeastward to at least the Delaware River bend and probably beyond. I, therefore, conclude that the Stafford fault zone is a long linear fault system between north-central Virginia and Long Island and that late Miocene to Pleistocene, up-to-the-west displacement on the fault zone produced at least some of the Fall Line incision. The linear, steeply-dipping nature of the Stafford fault zone, which is characteristic of many strike-slip faults (Sylvester, 1988), strongly suggests that it is an oblique strike-slip fault zone. This conclusion is supported by counterclockwise oblique fold axes subsidiary to the mapped Stafford fault zone (Mixon and others, 1992), which indicate a dextral component of slip.

Could the Stafford Fault Zone be a Secondary Manifestation of Flexural Isostatic Deformation?

Pazzaglia and Gardner (1994) postulated that the high angle faulting near the inner (west) edge of the Salisbury embayment might be a secondary manifestation of flexural isostatic deformation in the uppermost crust. However, several observations argue against such an origin for the Stafford fault zone. First, a convex-up flexural hinge zone favors bending moment (normal) faulting, not oblique strike-slip displacement (Quinlan, 1984) like that associated with the Stafford fault zone (Mixon and Newell, 1977). Secondly, the age of the Stafford fault zone is at least early Cretaceous (110 Ma, Mixon and others, 1992) whereas the flexural hinge zone modeled by Pazzaglia and Gardner (1994) developed within only the last 15 million years. Third, the RBT, which is likely a geomorphic expression of the Stafford fault zone, extends northeastward well beyond the flexural hinge zone proposed by Pazzaglia and Gardner (1994) along the west edge of the South New Jersey arch (Figure 1). Finally, the RBT is linear whereas the trend of a flexural hinge zone would reflect the nonlinear shape of the embayment's western border (Figure 1). I, therefore, conclude that the Stafford fault zone was not produced by flexural isostatic deformation.

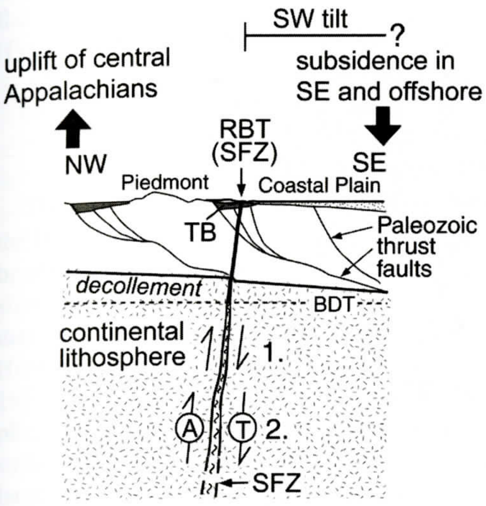
Crosscutting Relationship of the Stafford Fault Zone with Paleozoic and Triassic Terranes and Structures, and its Possible Basement Origin

At least three characteristics of the Stafford fault zone distinguish it from the Paleozoic thrust faults and early Mesozoic basin border faults and suggest that it crosscuts these structures. The first of these characteristics is that it dips steeply to the northwest whereas the nearby thrust faults and basin border faults in the Piedmont and Coastal Plain between central Virginia and New Jersey dip mostly southeastward, becoming listric at depth (Ratcliffe and others, 1986; Pratt and others, 1988; Root, 1988; Milici and others, 1991; Benson, 1992; Nowroozi and James, 1994; Owens and others, 1998; LeTourneau, 1999; Plank and others, 2000) (Figure 10). The inferred steep northwest dip of the Stafford fault zone is from geologic studies of the mapped Stafford fault zone in northern Virginia (e.g., Mixon and others, 1992) and the focal mechanism study of the 1973 Wilmington, Delaware, earthquake (Sbar and others, 1975). Although figure 10 only shows the larger regional thrust faults, other shorter thrust faults and the allochthonous terranes also generally dip southeastward near the Stafford fault. The bedrock geologic map of Plank and others (2000), for example, shows that the Paleozoic thrust faults and terranes just west of the Stafford fault in Delaware dip southeastward. Just southwest of the mapped Stafford fault, the Lakeside mylonite zone and its northeast extension, the Spotsylvania magnetic lineament (SML), dip less than 60° to the southeast (e.g., Mixon and others, 1992) (Figure 10). The regional I-64 seismic-reflection profile (Figure 10) also reveals the southeast dip of the regional thrust faults (Pratt and others, 1988).

The second distinguishing characteristic of the Stafford fault zone is the lack of basalt intrusives along its trend. This observation indicates that it did not form during early Mesozoic rifting.

The third distinguishing characteristic of the Stafford fault zone is its linearity, which contrasts sharply with the curved trends of the Paleozoic faults and Triassic basin borders. Such linearity suggests that it formed as a strike-slip fault by a compressional stress field oriented obliquely to the fault zone.

These characteristics strongly suggest that the Stafford fault zone cross-cuts the Paleozoic faults and Triassic basins, indicating that either it formed since the end of early Mesozoic rifting or that it is a reactivated pre-Alleghanian basement fault that has fractured the overlying terranes (Figure 11). It is unlikely, though, that late Mesozoic or Cenozoic compressional strain could have produced such a new 450-km-long



SCENARIOS FOR DISPLACEMENT ON THE STAFFORD FAULT ZONE:

1. Up-to-the-west displacement on the SFZ from isostatic uplift to the west and subsidence to the southeast in Salisbury embayment and offshore.
2. Oblique, up-to-the-west dextral displacement on SFZ from obliquely-oriented maximum horizontal compressive stress field.

Figure 11. Conceptual northwest-southeast-oriented cross-section (not to scale) across Salisbury embayment (SE) and Piedmont illustrating the steeply-dipping nature of the Stafford fault zone (SFZ) and the two possible forces possibly causing west-side-up and dextral displacements on it (see text). Dextral component of displacement denoted by 'A' (away from observer) and 'T' (toward observer) within circles. Solid arrows above cross-section denote isostatic uplift of the central Appalachians to the west and subsidence to the southeast in the Salisbury embayment and offshore, thus possibly contributing to the up-to-the-west displacement on the Stafford fault southwest of the Delaware River. Also shown is the location of the southwest tilt on the southern flank of the South New Jersey arch and in the Salisbury embayment (southeast side of the Stafford fault) that caused the major rivers to migrate to the southwest. BDT, brittle-ductile transition zone; RBT, river bend trend that defines the Stafford fault location; TB, Triassic basin (gray patterns).

strike-slip fault across the allochthonous terranes because dextral reactivation of the numerous northeast-trending Alleghanian faults throughout the Coastal Plain and Piedmont provinces could have accommodated the regional strain in the upper crust. Another possible cause of the crosscutting relationship is that the Stafford fault zone is a pre-Alleghanian strike-slip basement fault whose reactivation at depth sometime since the end of early Mesozoic rifting has fractured the overlying allochthonous terranes (Figure 11). This basement origin is analogous to the Precambrian basement fault beneath the décollement to the west associated with the New York-Alabama lineament (NYAL) (King and Zietz, 1978; Johnston and others, 1985), except for the fracturing of the overlying allochthonous terranes by the Stafford fault zone. Unfortunately, unlike the sedimentary rocks of the Valley and Ridge province along the NYAL, the relatively high magnetization and greater thickness (7 to 15 km) of the Piedmont allochthonous terranes to the east makes detection of the Stafford fault zone beneath the terranes difficult using magnetics and gravity techniques.

Possible Causes of Up-to-the-West Displacement on the Stafford Fault Zone

There are at least two possible causes of the up-to-the-west displacement on the Stafford fault zone, both of which are likely superimposed. One likely cause is oblique, up-to-the-west dextral displacement on the fault zone produced by its oblique orientation relative to the northeast–southwest-oriented maximum horizontal compressive stress field in the mid-Atlantic states area (Zoback and Zoback, 1991) (Figures 1 and 11). Although the stress field shown on figure 1 represents only the present stress orientation, the dextral component of displacement documented from the detailed trench study across the Dumfries segment of the Stafford fault zone in Virginia (page 438 of Mixon and Newell, 1977) supports a similarly oriented stress field at least back to the early Pleistocene to produce oblique movement on

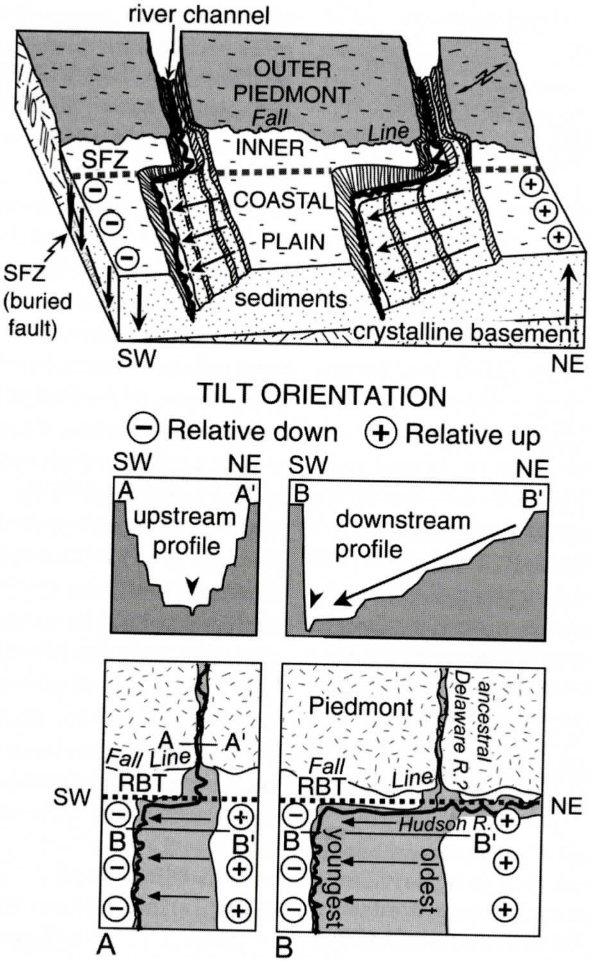


Figure 12. Conceptual model demonstrating how termination of regional uplift associated with South New Jersey arch against the Stafford fault zone (SFZ, defined by river bend trend (RBT)) produced the southwest river bends and an abrupt change in cross-valley shape of larger rivers along the inner (western) Coastal Plain. The steeply-dipping SFZ decouples the upper crust to the west from that to the east. Consequently, regional down-to-the-southwest tilting stops to the west against the SFZ, causing river migration to the southwest downstream from the fault (map A, lower left). Continued cross-valley tilting and river migration downstream results in a down-to-the-southwest cross-valley tilt with unpaired slip-off terraces on the northeast side of the valley. The amount of river migration is greater in valleys toward the northeast, suggesting that the primary source of the tilt is from uplift to the northeast at the South New Jersey arch. In contrast, absence of cross-valley tilting northwest of the SFZ allows rivers upstream to meander freely and form symmetric cross-valley shapes with paired terraces. Map B in the lower right shows how some larger ancient rivers, such as a paleo-Hudson, temporarily flowed from the northeast along the inner Coastal Plain, then turned abruptly southeastward down the Delaware paleovalley where they migrated to the southwest. Stafford fault zone (SFZ) projected to the surface above the Coastal Plain sediments denoted by gray dashed line in upper diagram.

the fault zone. Another possible cause of the up-to-the-west displacement on the Stafford fault zone is that it may be accommodating some of the crustal strain caused by isostatic uplift of the Appalachians to the west and sediment loading in the Salisbury embayment to the east (Figure 11). This conclusion is supported by the steep, up-to-the-west reverse fault plane solution of the 1973 Wilmington, Delaware, earthquake (Figure 9A).

Tectonic Origin of the Major River Bends

The prominent 300-km-long alignment of the Potomac, Susquehanna, and Delaware river bends and their proximity to the Fall Line incision prompted many investigators to speculate that they are fault-related (e.g., McGee, 1888; Higgins and others, 1974; Mixon and Newell, 1977). Although the linearity of the RBT implies that the underlying Stafford fault zone is a strike-slip fault system, there is no evidence for large dextral strike-slip displacement during the Cenozoic that could have produced the 35- to 100-km-offsets of the major rivers, nor could such large displacements have occurred because of the very low Cenozoic slip rates on faults in the eastern U.S. (Prowell, 1988).

Marple (1998) and Marple and Talwani (2000) provided a tectonic model that could explain these river bends without strike-slip displacement. They postulated that the down-to-the-southwest cross-valley tilt of the Cape Fear and Pee Dee River valleys in the Coastal Plain was caused by termination of regional tilting on the southern flank of the Cape Fear arch to the west against the East Coast fault system (Figures 5B and 23 of Marple and Talwani, 2000).

Larger river valleys traversing the mid-Atlantic Fall Line display similar morphological changes where they traverse the RBT (Figures 5 and 6). Perhaps the best example of such a deflection is the southwest migration of the Potomac River (e.g., Schlee, 1957). Not all of these major rivers, however, necessarily were directly deflected by cross-valley migration. The bend of the present Delaware River, for example, may have simply occupied an ancient valley of the paleo-Hudson River (Stanford, 1993). The paleo-Hudson River did, though migrate laterally southwestward at least 55 km just downstream of the RBT (Owens and Minard, 1979) (Figure 12). The Bridgeton deposits, and thus the southwest river migration, may have extended much farther to the northeast, but would have been removed by stream erosion in the Rancocas Creek valley (Figure 6 cross-section) and by streams draining the outer Coastal Plain to the southeast. This abrupt change to cross-valley migration of ancient rivers downstream of the RBT strongly suggests the presence of a steep, deep-crustal fault system that has decoupled the crust east of the RBT from that to the west (Figures 11 and 12). Because of this decoupling, regional down-to-the-southwest tilting in the embayment during Miocene to late Pliocene time abruptly stopped to the west along the RBT. This termination of tilting also implies that the vertical displacement on the Stafford fault zone should increase to the southwest, which is supported by the increase in maximum vertical offset along the fault at the base of the Coastal Plain sediments from about 45 m near the Delaware River bend (Thompson, 1978, 1979, 1981) to about 75 m near the Potomac River bend (Mixon and others, 1992) (Figure 1). However, the decoupling and down-to-the-southwest cross-valley tilt do not require coseismic slip on the Stafford fault zone to allow cross-valley tilting east of the RBT.

Although the magnitude of such a tilt is very gentle (<0.2 m/km), it is clear that it caused the major rivers to migrate southwestward. Thus, it demonstrates how even subtle tectonic deformation can greatly impact alluvial rivers.

Source of Regional Southwest Tilt

Based on the proximity of the South New Jersey arch to the northeast end of the river bends (Figure 1), the most likely cause of the southwest tilting in the Salisbury embayment is uplift at the arch, thus causing the rivers on its southern flank to migrate away from it. Several observations support this mechanism. First, the sizes of the river bends and distance of river migration to the southwest decrease progressively away from the arch (Figure 1). Sedimentologic studies, for example, northeast of the Delaware estuary (Owens and Minard, 1979) indicate that an ancient Hudson River migrated laterally at least 70 km to the southwest across the New Jersey Coastal Plain during late Miocene to Pliocene time, leaving behind Bridgeton Formation (fluvial) sediments that are progressively older to the northeast (Figure 6). The Potomac River, in contrast, not including migration along its southwest-convex curve southeast of the RBT, has migrated less than 35 km since mid-Miocene time (Schlee, 1957). Secondly, the right-stepping bend of the Potomac River is opposite that expected if differential subsidence in the embayment had caused the tilt because the axis of the embayment lies northeast of the Potomac River (Figure 1). The Cape Fear and Pee Dee rivers display similar characteristics on the southern flank of the Cape Fear arch in the Carolinas (Marple and Talwani, 2000). The Cape Fear River, which is much closer to the arch axis, has migrated much farther to the southwest than the Pee Dee River since mid-Pliocene time, producing a large right-stepping bend in the river and broad, unpaired terraces to the northeast.

Another mechanism that I investigated for the cause of the cross-valley tilting that produced some of the river deflections was glacial loading and unloading north of the embayment. This process likely did not produce the regional tilt in the Salisbury embayment though because the tilt is limited to the area southeast of the RBT; glacial loading and unloading effects likely would have also caused the area northwest of the RBT to be tilted southwestward. Furthermore, the regional tilt appears to have ended before the Pleistocene glaciations (see next section).

Timing of Southwest Tilt

Based on sedimentologic and geomorphic data (e.g., Pazzaglia, 1993; Figure 6), the southwest tilt in the Salisbury embayment and postulated uplift at the South New Jersey arch likely occurred during Miocene to late Pliocene time. Cessation of the regional tilt during late Pliocene time is inferred from the late Pliocene northeast migration of the paleo-Hudson River as it crossed the Coastal Plain between the modern Chesapeake and Delaware bays, leaving behind the late Pliocene-age Pensauken fluvial sediments in the subsurface that are progressively younger to the northeast (Owens and Minard, 1979). Nor is there evidence for significant regional tilting in the embayment since the Pliocene. First, the youngest (Pleistocene) terraces in the lower Delaware River are paired on both sides of the valley (Figure 6). Secondly, although the late Pleistocene paleochannels of the Susquehanna River migrated southwestward Delmarva Peninsula (Colman and others, 1990), this migration did not produce a river bend upstream in the inner Coastal Plain and, therefore, is not associated with the right-stepping bend of the Susquehanna River (Figure 1). Likewise, there is no indication of cross-valley tilting along the Potomac valley during late Pleistocene time, except for the lower (Pleistocene) terraces inside its southwest-convex curve southeast of the RBT (Figure 6). These younger terraces are probably related to formation of the curve by tilting of a large fault block as described in the previous section, rather than by regional southwest tilting. Thus, most regional southwest tilting in the embayment apparently stopped during late Pliocene time.

Origin of Potomac River Curve and Cross-valley Tilting in the Patuxent Valley

Another conspicuous feature along the inner Coastal Plain of the mid-Atlantic is the southwest-convex curve in the Potomac River valley just downstream of the RBT (Figure 1). The southwest cross-valley tilt and progressively younger terraces downslope toward the channel (Figure 6) indicate that the curve formed by local downcutting and channel migration to the southwest. Possible causes of the southwest migration include a local change in sediment erodibility and tectonic tilting. There is no change in sediment erodibility nor southwest dip of bedding (McCartan, 1989), however, that could have caused the river migration. A more likely cause is southwest tilting of a fault-block beneath the Coastal Plain based on the curve's location between the Stafford fault zone on the west and the Port Royal fault to the east (Hack, 1982; Mixon and others, 1992) (Figure 1). Ages of terrace sediments inside the Potomac River curve (Figure 6) indicate that the tectonic deformation and river migration occurred from late Miocene to at least late Pleistocene time.

The local cross-valley tilting of the Patuxent River valley downstream of the RBT (Figure 5) suggests that it too is associated with southwest tilting of a fault block between the Stafford fault zone on the west and an undocumented fault on the east. The middle to late Pleistocene age of the Patuxent terrace (Glaser and Hansen, 1973; Glaser, 1984) indicates that the causative cross-valley tilting occurred during middle to late Pleistocene time.

Seismicity Along the Stafford Fault Zone and Its Seismic Potential

Much seismicity in the eastern U.S. is attributed to reactivation of Triassic basin faults and Paleozoic shear zones (e.g., Ebel and Kafka, 1991), such as that near Cameron's–Martic Line and the Newark Triassic basin (Figure 9). However, the alignment of historical seismicity and northeast-trending isoseismals of the 1971 Fredericksburg, 1973 Wilmington and 1884 New York City earthquakes with the river bend trend (RBT) between Virginia and New York City (Figure 9) suggests that this seismicity is associated with displacement on the Stafford fault zone. This seismicity trend is oblique to basin trends and parallels the long axis of isoseismals of historical earthquakes (Figure 9). The coincidence of the northeast end of the Stafford fault zone (as extended northeast herein) with the area of rapid horizontal ductile shear strain striking north to northeast in the lower crust near New York City (Zoback and others, 1985; Prescott and others, 1987) makes it a candidate for the cause of this strain and the source of the three largest New York City earthquakes in 1737, 1783, and 1884. Seismicity near New York City might also be related to stresses produced by the intersection of the Stafford fault zone and northwest-trending cross-faults, such as the Dobbs Ferry fault of Seeber and Dawers (1989). Confirmation of these hypotheses, however, will require more seismicity, geodetic, and seismic-reflection data along the RBT.

SUMMARY

Geomorphic, geologic, and geophysical data suggest that late Cenozoic, up-to-the-west displacement on the Stafford fault zone, possibly combined with other processes such as flexural deformation and sea level lowering, caused the incision along the Potomac, Susquehanna, and Delaware rivers at the U.S. mid-Atlantic Fall Line. This up-to-the-west displacement likely involved two processes: 1) oblique, up-to-the-west dextral displacement on the fault driven by the northeast–southwest-oriented horizontal compressive stress field in the mid-Atlantic area and 2) up-to-the-west displacement caused by isostatic uplift of the Appalachians to the west and sediment loading in the Salisbury embayment to the east. The large river bends, in contrast, were produced by abrupt termination of down-to-the-southwest tilting in the Salisbury embayment to the west against the Stafford fault zone during Miocene to late Pliocene time. The source of the tilt is likely from uplift at the South New Jersey arch northeast of the river bends. The abrupt northwestern limit of the southwest tilt and the linear nature of the Stafford fault zone strongly suggest that it is a steeply dipping, deep-crustal oblique strike-slip fault that has decoupled the crust southeast of the fault from that to the northwest. Historical seismicity along most of the river bend trend (RBT) (Figure 9) suggests that deformation is occurringon the Stafford fault zone extended herein.

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X-RAY POWDER DIFFRACTION EVIDENCE FOR SHOCKED QUARTZ IN AN UPPER EOCENE SAND DEPOSIT, WARREN COUNTY, GEORGIA, U.S.A.

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ABSTRACT

Shocked quartz grains collected from a transgressive lag at the sequence boundary between the upper Eocene Twiggs Clay and the kaolin-rich middle Eocene Huber Formation in the Georgia Coastal Plain were examined by X-ray powder diffraction (XRD) for evidence of crystal defect line broadening. The deposition of the sand was approximately correlative with the Chesapeake Bay impact (35 Ma), and thus the sand is a logical place to look for impact debris, especially because similar-age tektites occur in south-central Georgia. Sample comparisons were made between three groups of fine to medium sand-size quartz separates: (1) quartz grains that display single or multiple sets of deformation features, which are petrographically similar to planar deformation features (PDFs) induced by shock loading, (2) quartz grains that were randomly selected from the sand unit, and, as a control, (3) quartz grains from the Crow Creek Member of the Pierre Shale (Upper Cretaceous Nebraska) that display obvious sets of PDFs formed in association with the Manson impact. Alluvial quartz grains from crystalline rocks of the Georgia Piedmont were used an instrument standard. Fourier numerical analysis of the quartz (*100*) peaks shows that grains exhibiting PDFs have a higher defect density than grains from the random sampling. The mean coherent scattering domain size of the Eocene sands with PDFs is 960Å, which is about half the size of those in a random sampling. The presence of quartz with PDFs and defect densities higher than the surrounding grains is consistent with the notion that the deposit contains remnants of material ejected by the Chesapeake Bay impact. These XRD findings, taken with the petrographic evidence for PDFs in quartz grains provide the first evidence for in-place preservation of upper Eocene impact deposits in Georgia Coastal Plain strata.

INTRODUCTION

Previous investigations designed to search for an impact horizon in the southeastern US Coastal Plain associated with the Chesapeake Bay impact (Albin, 1997; Albin and others, 2000; Horwath, 1990; Zwart, 1978) have fallen short, in part because they were designed to look for geochemically unstable microtektites. A second factor includes limiting the hunt to units that are isochronous with the impact event. Albin and Wampler (1996) were the first to make inference that evidence for impact ejecta could be found in Eocene Dry Branch Formation (Twiggs Clay member). In this study, we focus on characterization of the more geochemically resilient quartzose material as a recorder of impact events in Eocene age sediments.

In east-central Georgia, the upper Eocene Twiggs Clay overlies the middle Eocene kaolin-rich Huber Formation. The contact between the two units is a sequence boundary and transgressive surface overlain by a patchy coarse-grained sand layer as thick as 10 cm. Harris and others, (2002) determined that approximately 3-5% of fine- to medium-grained quartz examined from the layer contain optical planar elements (spacing and orientation) similar to planar deformation features (PDFs). Most of the quartz grains display one set of planar features, but some exhibit two or three intersecting sets (Harris 2003). We interpret these linear optical features are PDFs and propose that the quartz was shocked by bolide impact. We hypothesize that quartz grains exhibiting PDFs have a higher crystal defect density than quartz grains that do not display PDFs. The coherent X–ray scattering dimension (CSD_{hkl}) is defined as the *d*–spacing for an *hkl* plane multiplied by the number of defect-free translations perpendicular to the plane. A single coherent scattering domain is easily understood as a set of perfectly aligned cards bounded by faults. Larger mineral grains can be visualized as a mosaic of domains that are characterized the mean of all the CSD_{hkl} lengths. Shocked-quartz grains should therefore have measurably smaller mean CSD_{hkl} lengths than optically clear quartz grains.

Two major concerns arise when X-ray powder diffraction (XRD) is used to analyze ejecta for identification of shocked quartz. The first concerns the frequency of shocked quartz in sediments that preserve a record of distal ejecta. Shocked grains are often rare in siliciclastic sediments, because allochthonous non-shocked quartz grains may dilute the system (e.g., Exmore Breccia, Poag and Poppe, 1998). The presence of only a few non-shocked grains in an XRD experiment overwhelms the XRD pattern because of contributions from their large CSDs. XRD peak intensities (I) (*i.e.*, shapes) are proportional to the square of the number (N) of defect-free scattering domains. This is illustrated in the theoretical intensity function, which is the basis for the calculation of continuous XRD patterns (Moore and Reynolds 1997):

$$I(\theta) = LpG^2\Phi \tag{1}$$

where

$$\Phi(\theta) = \frac{\sin^2(2\pi ND\sin\theta/\lambda)}{\sin^2(2\pi D\sin\theta/\lambda)} \tag{2}$$

and Lp = Lorentz-polarization factor for random powders, G = layer scattering factor, Φ = the interference function, D = interplanar d-spacing, Θ = the Bragg angle, and λ = wavelength of X–ray radiation.

A solution to this problem is to handpick shocked grains and conduct XRD analysis on minimal quantities (8 to 11 grains ~ 2 mg) of powdered specimen. One potentially detrimental effect of having small sample quantity is a lack of statistical representation of coherently diffracting crystal domains. This is partially compensated by using long counting times for data collection and by repeated data collection after remounting of the sample. Additional consequences of a thin sample are that transparency effects become minimized and that higher order reflections are diminished (Hurst and others, 1997).

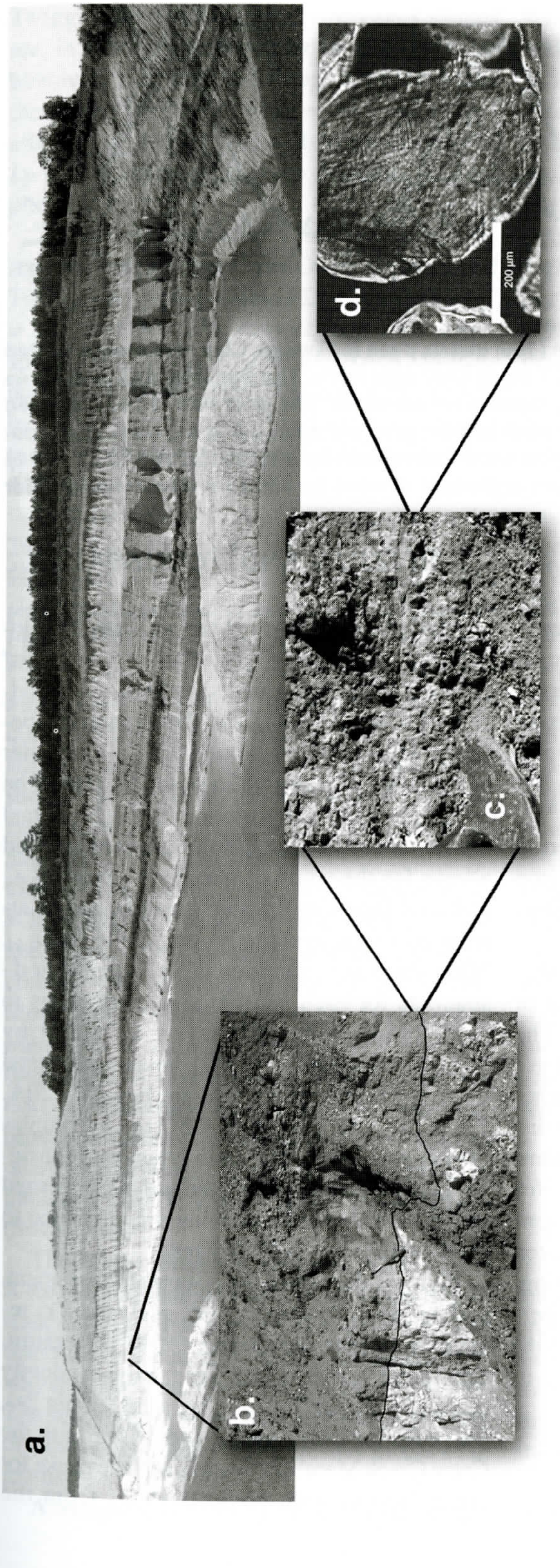
The second concern arises because XRD instrument optics significantly distort peaks and a mere comparison of observed profiles is not enough to know the line broadening contribution due to the distribution of CSDs. CSDs are therefore evaluated by correcting for instrumental line broadening and assuming that grains without linear features are defect free (*i.e.,* the size of CSDs = ∞). The correction of widths and shapes of X-ray powder diffraction lines is accomplished by the numerical Fourier-analysis of Stokes (1948). It is assumed that the observed intensity function $h(x)$, is related to the instrument broadening function $g(x)$ and the CSD broadening function $f(x)$,

$$h(x) = \sum f(y)g(x-y)\delta y \tag{3}$$

where x and y are auxiliary variables that are typically measured in units of °2Θ. Eq 3 constitutes the set (a) of linear simultaneous equations in which the intensity values for each °2Θ of $h(x)$ and $g(x)$ are known, and the values of $f(x)$ are unknown. By considering the values of °2Θ in the range of $-a/2$ to $+a/2$, where $h(x)$ and $g(x)$ fall to zero, it is possible to obtain the Fourier series of all the functions. In practice this results in a range that includes 60 equations or ±0.3 °2Θ in increments of 0.01°.

METHODS

Samples from the Twiggs Clay, sand layer, and Huber Formation were collected in the J.M. Huber Corp. Purvis Mine located near the Pur-



vis School on the west side of Georgia highway 17, approximately 8 km north of Wrens, Georgia (Fig. 1a,b). The underlying Huber Formation is ~ 97 percent kaolinite, with trace amounts of silt-sized quartz. The quartz in the overlying Twiggs Clay is very-fine sand and silt-sized and is a minor component. Consequently, optical selection of quartz with and without PDFs from the Twiggs and Huber deposits was not possible. Only the sand layer yielded sufficient quantities of fine to medium sand-sized quartz, for which grains with and without PDF could be clearly selected. We note here that the overlying sand units of the Irwinton Sand and Tobacco Road Sand (Elzea-Kogel and others, 2000) were specifically examined and did not yield grains with planar fabrics.

Control quartz grains used in this study included (1) alluvial grains collected from Beech Creek, located in the metamorphic Appalachian Piedmont near Athens, Georgia, and (2) grains collected from the known Manson impact ejecta in the Crow Creek Member of the Cretaceous Pierre Shale, Nebraska (provided by Sarah Chadima, South Dakota Geological Survey); (Witzke and others, 1996).

Figure1. Site of sampling near Wrens, Georgia. (a) Panoramic north-east-south view of the Purvis mine, operated by the J.M. Huber Corp. Visible are the middle Eocene Huber Formation, overlain by the marine upper Eocene Twiggs Clay, which in turn is overlain by migrating channel-fill sands, of the Irwinton and Tobacco Road Sands. (b) Sequence boundary (high lighted by line) showing the position of the thin (0-10 cm) sand lag deposit. (c) Close-up view of lag deposit (hammer for scale). (d) Cross-polarized photomicrograph of quartz grain from the lag layer, exhibiting PDFs.

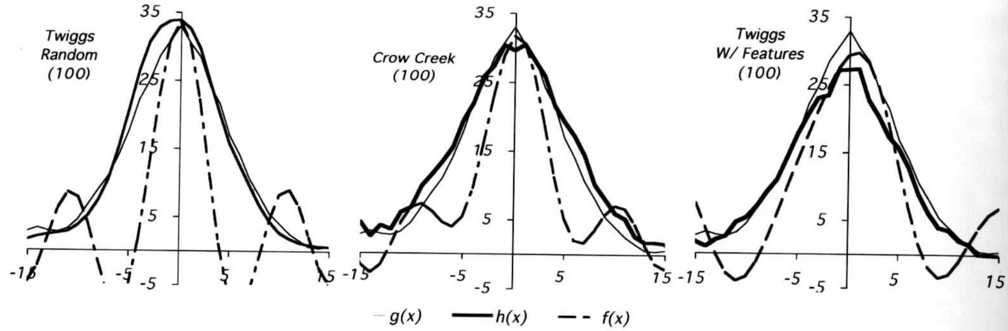


Figure 2. X-ray diffraction line profiles of the (*100*) quartz peak for Twiggs Clay randomly selected grains, Crow Creek Member with PDFs, and Twiggs Clay with PDFs. The thin line, *f(x)*, in all three comparisons is the Beech Creek, Georgia, instrument standard. The bold line is the observed profile, *h(x)*, and the dashed line is the true peak profile, *g(x)*, unfolded using the numerical Fourier analysis of Stokes (1948). CSD_{100} lengths were determined by a least-squares fit of *g(x)* to equation 2 and then multiplying the resultant optimized *N* value by 4.2725 Å (the d–spacing for the (*100*) reflection).

Samples were hand ground using a zirconium mortar and pestle with alcohol. The powdered material was transferred to a zero-background plate; a large (30 x 30 mm) single quartz crystal cut and polished 6° off the *c*-axis, which has the effect of minimizing substrate scattering effects. XRD analyses were conducted using a Scintag diffractometer, with Co Kα radiation, a 250 mm goniometer circle, 2°/4° primary and scattering slits, 0.5°/0.3° scattering and receiving slits, 40 kV and 40 mA, a step size of 0.01°, and a count time of 10 s per step. The quartz (100) and (101) reflections (indexed in the $P3_221$ space group) were used in this study because they provide sufficient signal to noise to give reliable counting statistics. The use of thin sample thicknesses precludes the use of the higher order reflections (peaks at higher angles) because of inadequate signal to noise response. For each sample, three to four pattern replicates (after remounting) were corrected for displacement and averaged to minimize orientation. Averaging also reduced noise generated from the instrument and Compton scattering. All Fourier transform calculations were performed using the equations of Stokes (1948) and Klug and Alexander, (1974) facilitated in an Excel spreadsheet.

RESULTS AND DISCUSSION

As discussed above, coherent scattering domains are most easily visualized as mosaics of crystallites that are bounded by faults (Schnieder and others, 1984). The disorientation between adjacent domains need be only a matter of fractions of a degree in order for one to be considered separate from another (Moore and Reynolds 1997). The Fourier analysis method allows for the study of true peak shapes (*i.e.*, unfolded functions), whose breadths can be attributed to the average length of the CSD_{hkl}. Microstrain (*i.e.*, variation of the interplanar d–spacings for each *hkl* set) also contributes to peak broadening. However, as noted by Schneider and others, (1984), strain values tend to be constant when CSDs are beyond 220Å. Given the large size of the CSD in our samples (discussed below), the strain effect was considered to be constant among the samples analyzed and differences in broadening are attributed to differences in mean CSD sizes.

Figure 2 illustrates the observed, instrument, and unfolded functions, *h(x)*, *g(x)*, and *f(x)* respectively, for the quartz (100) reflection. Shown are the peaks for quartz grains randomly selected from the Twiggs Clay sand lag (fig 2a), the peaks for quartz grains with PDFs from the Crow Creek Member (fig. 2b), and the peaks for quartz grains with possible PDFs from the

Twiggs Clay sand lag (fig. 2c). Peaks shapes are, in theory, influenced by *Lp* and *G* (Eq.1), however, over this small region of 2Θ space these functions are nearly constant and do not affect peak shape (Klug and Alexander 1974). *Lp* and *G* corrections were therefore not applied.

The average number of *(100)* domains in the random Twiggs Clay, Crow Creek Member, and Twiggs Clay with possible PDFs are $N = 460$, $N = 330$, and $N = 225$, respectively. This corresponds to CSD_{100} lengths of 1970Å, 1410Å, and 960 Å, respectively. *N* values were derived by a least-squares curve fit of the data sets *f(x)* to the interference function, Eq 2 (Raner 1998). All fits have an R^2 of 0.99.

Relative comparison of quartz grains in the Twiggs Clay lag deposit reveals that grains with PDFs have twice the defect density in the *(100)* planes compared to grains that do not exhibit PDFs. The CSD_{100} length of shocked Twiggs Clay quartz is seven times smaller than the mean CSD lengths of laboratory annealed quartz (Schneider and others, 1984). It is six times larger than quartz experimentally shocked at 15 GPa and about the same as quartz experimentally shocked at 1.5 GPa (Schneider and others, 1984). At first this seems contradictory to experimental work, which suggests that at least 10 GPa is needed to create PDFs (Grieve and others, 1996). However, making such direct comparisons between laboratory and natural shock events is difficult, because natural shock events can be marine (wet) impacts or terrestrial (dry) impacts, and the target can be either porous (sedimentary) or non-porous (metamorphic and igneous). Moreover, the shocked quartz domains can anneal over long periods of geologic time at earth surface temperatures (Sokur, 1999).

The *(101)* reflections were also examined for X-ray line broadening and in all cases a constant CSD_{101} of about ~1500Å ($N = 450$) was determined. We have no clear explanation as to why the *(100)* and *(101)* shapes do not co-vary. The reason may be that preferential directions of planar dislocations are present in the PDF grains.

A recognized difficulty in the Fourier analysis is choice of background correction (Young and others, 1967). The tails of the diffraction peaks determine the Fourier coefficients. Our study suffers from less than ideal signal-to-noise ratio because the data comes from the analysis of only eight to eleven grains. Obtaining larger sample mass would allow for both better counting statistics and for analysis of higher order reflections. However, as noted in our introduction and by Schneider and others, (1984), XRD of thousands of grains is a method biased by grains with large mean CSD sizes.

The second factor of directional defects may be possible. Goltrant and others, (1992) discussed the fact that shock deformation in quartz is a strongly heterogeneous process. Critical sets of PDFs identified optically commonly include the *(001)*, *(102)*, and *(103)* planes and less commonly *(101)* and *(104)* planes. Is it important to note that CSDs for only the *(100)* and *(101)* can be studied by XRD. XRD intensities for planes seen optically, with exception to the *(101)*, are too low to determine a mean CSD_{hkl} value. The distinction amongst these various factors may only be resolved by high-resolution methods of electron diffraction and transmission electron imaging.

Our results show that XRD analysis of selected grains serves as a screening tool to support the recognition of shocked quartz. XRD units are common to most geology departments and the sample preparation and data collection do not require any special tools (except a zero-background substrate). The correlation of mean CSD lengths to actual shock deformation processes is possible with careful XRD analysis.

CONCLUSIONS

Bulk XRD methods that analyze large populations of quartz grains with variable CSD lengths are not capable of distinguishing the presence of shocked quartz. However, the XRD technique of analyzing selected grains on a zero-background plate provides an easy way to screen quartz grains suspected to have been shocked by hypervelocity impacts. Analysis of quartz grains exhibiting PDFs from a basal lag in the upper Eocene Twiggs Clay near Wrens, Georgia is consistent with the notion that the deposit contains ejecta from the Chesapeake Bay impact.

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ORIGIN OF GULF COAST OPAL-CT: A STUDY OF THE CLAYSTONE-RICH TALLAHATTA FORMATION IN MISSISSIPPI

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ABSTRACT

The origin of Tertiary sedimentary opaline materials in the southeastern United States has been attributed to either biogenic or volcanogenic processes. Biogenic processes involve deposition of silica-rich tests of radiolarians and diatoms, with subsequent diagenetic alteration of opal-A to opal-CT. Volcanogenic processes create opal-CT during alteration of volcanic ash, often accompanied by growth of zeolite minerals. The Lower Eocene Tallahatta Formation, which contains abundant opal-CT, was examined to resolve a dominantly biogenic or volcanogenic origin for this formation in eastern Mississippi. Nine cores spanning the entire Tallahatta thickness near Meridian, Mississippi, along with nearby outcrops, were used to generate 577 samples. X-ray diffraction analysis (XRD), scanning electron microscopy (SEM), chemical analyses and petrographic investigations were used to characterize the mineralogy, texture, chemistry and fossil abundance of opal-CT-rich samples of the Tallahatta Formation. Based on quantified XRD data, about 80 percent of the samples consisted of opal-CT and the zeolite clinoptilolite, with lesser amounts of quartz, smectite group clays, and muscovite. Glauconite was observed in thin section. Nearly 75 percent of the Tallahatta claystones in the Meridian area had clinoptilolite as a dominant mineral. Few fossils were observed, either in thin section or SEM. No opal-A was found using XRD, suggesting that opal-A may not have been a precursor to opal-CT in the claystones. This opal-CT-clinoptilolite-smectite assemblage, along with the general lack of fossils, is indicative of a dominantly volcanogenic rather than dominantly biogenic origin for the opal-CT in the study area. Opaline claystones of the Tallahatta Formation in Alabama and eastward are reported to have numerous fossils and much less clinoptilolite, suggesting a greater biogenic input to the origin of opal-CT in those areas.

INTRODUCTION

Opaline materials have been described in numerous Tertiary sedimentary units in the southeastern United States (e.g., Brindley, 1957; Heron and others, 1965; Wermund and Moiola, 1966; Reynolds, 1966, 1970; Heron, 1969; Carver, 1972, 1980; Pollard and Weaver, 1973; Wise and Weaver, 1973; Weaver and Wise, 1974; Weaver and Beck, 1977; Laws and Thayer, 1992). Opaline claystones of the Lower Eocene Tallahatta Formation are present throughout the eastern Gulf and Atlantic coastal plain from Mississippi to Georgia and South Carolina. Two theories have been proposed for the origin of opal-CT beds in the Tallahatta Formation and equivalent strata, one that the original silica was of biogenic origin (Cunningham, 1894; Wise and Weaver, 1973; Weaver and Wise, 1974; Carver, 1980; Laws and Thayer, 1992), whereas the other proposes a volcanogenic origin (Grim, 1936; Reynolds, 1966, 1970; Heron, 1969; Gibson and Towe, 1971;

Matson and Pessagno, 1971). Study of samples from both core and outcrop of opal-CT-bearing units in the Tallahatta Formation near Meridian, Mississippi, was conducted to evaluate potential origins of the silica enrichment.

Opaline Silica Phases

Opal and opaline silica can be grouped as opal-A, opal-CT and opal-C (Jones and Segnit, 1971). Opal-A is a highly disordered, near amorphous silica polymorph precipitated from solution either organically or inorganically. Marine silica-secreting organisms such as diatoms, radiolarians and sponges can extract silica from seawater at concentrations well below saturation levels with respect to amorphous silica. Inorganically precipitated opal-A forms when a solution becomes saturated with respect to silica, causing silicic acid polymer to precipitate. These polymers then grow by the Ostwald ripening process (in which larger particles with lower solubilities grow at the expense of smaller particles with higher solubilities) to form colloidal sols and gels (Iler, 1979). Opal-A produces an X-ray diffraction (XRD) pattern of a single diffuse band extending from about 19° 2θ to 25° 2θ. Comparative XRD patterns of various opal forms and quartz are given in Murata and Larson (1975), Calvert (1983) and Williams and others (1985).

Opal-CT is a disordered variety of low cristobalite that contains randomly interstratified tridymite (Flörke, 1955; Jones and Segnit, 1971; Flörke and others, 1975). This material is a transition phase between opal-A and microcrystalline quartz, and can form from opal-A as a solution-reprecipitation process (Carr and Fyfe, 1958; Mizutani, 1966; Kastner and others, 1977; Kastner, 1981; Williams and Crerar, 1985). Opal-CT was originally referred to as lussatite, (Mallard, 1890; Flörke, 1955), and later as cristobalite, α-cristobalite, or disordered cristobalite. After classification by Jones and Segnit (1971), the term opal-CT has been widely accepted in the literature. Opal-CT has an XRD pattern characterized by two broad reflections, one within 22.0° to 21.4° 2θ, corresponding to crystal lattice spacings (d-spacings) of 4.04Å to 4.14Å, and another at 36.9° 2θ (2.44Å d-spacing). Porcellanite is a term often used to describe sediments and rocks composed primarily of opal-CT (Calvert, 1977).

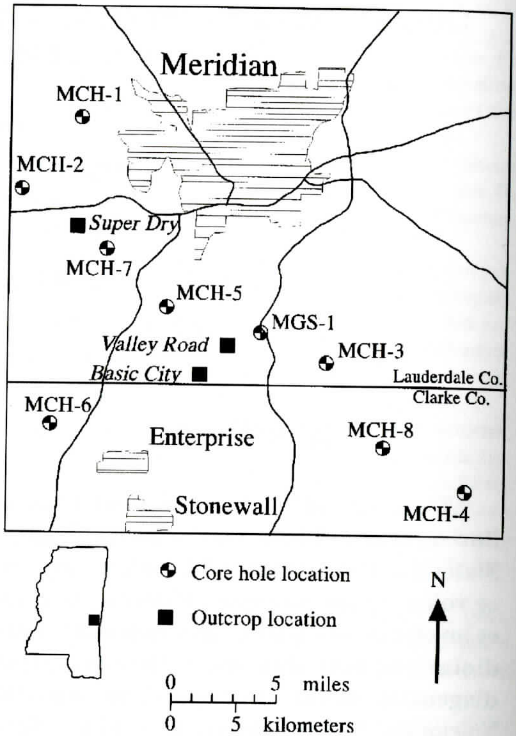

Figure 1. Location map of the study area.

Opal-C closely resembles cristobalite and is recognized by major XRD reflections occurring as sharp symmetrical peaks centered at 22.2° 2θ (4.04Å), 28.5° 2θ (3.13Å), 31.5° 2θ (2.84Å) and 35.9° 2θ (2.50 Å). Jones and Segnit (1971) believe this type of natural hydrous silica is commonly associated with lava flows. However, reports of authigenic well-ordered opal-C suggest that high temperature may not be required for the formation of this phase of hydrous silica (Ijima and others, 1980; Ijima and Tada, 1981).

GEOLOGIC SETTING

The study area covers about 660 square kilometers within Clarke and Lauderdale Counties of Mississippi (Figure 1). This area lies within the coastal plain of the Gulf of Mexico and ad-

Epoch		Group	Formations	
Eocene	Middle	Claiborne	Winona Formation	
	Lower		Tallahatta Formation	Neshoba Sand Member
				Basic City Shale Member
			Meridian Sand	

Figure 2. Stratigraphic nomenclature of the study area near Meridian, Mississippi. Nomenclature follows that of Dockery (1996).

jacent to the Mississippi Embayment, a northward extension of these Cretaceous and Tertiary sediments. Tallahatta sediments have been assigned to the Lower Eocene in recent work by Bybell and Gibson (1985) and Dockery (1996). The study area was selected based on the excellent exposures of the Tallahatta Formation, and the availability of nine cores taken through the entire formation.

In the study area the Tallahatta Formation lies unconformably on the Lower Eocene Meridian Sand, the lowermost stratigraphic unit of the Claiborne Group, and beneath the Middle Eocene Winona Formation (Figure 2). The contact between the Tallahatta Formation and overlying Winona Formation is conformable and marked by a gradational change from clay or mud to glauconitic sand (Roquemore, 1984). The contact between the Meridian Sand and Tallahatta Formation is easily recognized due to the sharp change in lithology from medium-grained Meridian sand to Tallahatta claystone.

Tallahatta Stratigraphy

The name Tallahatta first appeared in an article by Dall (1898). The term Tallahatta (named after the Tallahatta Hills in Choctaw County, Alabama) has generally been used in Mississippi since its introduction by Johnson in 1905 to replace Hilgard's term of "Siliceous Claiborne" (Thomas, 1942). Lowe (1919) initially subdivided the Tallahatta Formation in Mississippi into the "Winona Sand" and "Basic Claystone." The Winona is presently classified in Mississippi as a separate formation (Dockery, 1996). Thomas (1942) introduced the Neshoba Sand into the stratigraphic sequence and recognized it and the Basic City Shale as the upper and lower members, respectively, of the Tallahatta Formation in central and southeastern Mississippi. In northern Mississippi, the Tallahatta Formation is undifferentiated.

Regionally, the lithologic character of the Tallahatta Formation is quite variable along strike in a northwest-southeast direction. The Basic City Shale Member is well developed throughout eastern Mississippi, but gives way in central and western Mississippi to the Neshoba Sand Member. The shale facies in the Basic City Shale Member changes westward to sand and quartzite (Lowe, 1919; Merrill and others, 1985).

The Basic City Shale Member of the Tallahatta Formation consists mainly of grey to dark grey, glauconitic claystone. It can be divided into four units (Kabir, 1998). The lowermost unit consists of glauconitic claystone (clinoptilolite and smectite clay) and sandstone. The second unit is clinoptilolite-rich claystone, the third unit is opal-CT-rich claystone, and the uppermost unit consists of claystone (mainly clinoptilolite with smectite clay), siltstone, mudstone and sandstone. The thickness of the Basic City Shale Member averages 45 to 50 meters along strike.

PREVIOUS INVESTIGATIONS OF OPAL-CT IN THIS REGION

The origin of opal-CT in the southeastern part of the United States has been attributed to either volcanic or biologic processes. Cunningham (1894) described what later was identified as the Tallahatta Formation in Alabama as a product of silica-rich marine biogenic deposition of radiolarians and diatoms. Grim (1936) describes this same material as being of volcanic origin. Reynolds (1966, 1970) argues that opal-CT along with zeolite and montmorillonite in this area is the result of weathering of volcanic ash. Heron (1969) also indicates that zeolite and opaline silica were derived from volcanic ash, part of which originated in Mexico and the western United States. Gibson and Towe (1971) further proposed that such ash provides siliceous source material for Eocene Horizon A radiolarian cherts of the Caribbean and North Atlantic. These deep-sea cherts are time equivalent to some of the opaline coastal plain deposits of the southeastern United States (Gartner, 1970). Mattson and Pessagno (1971) found evidence of a volcanic origin for Horizon A cherts, possibly as a product of dacitic and andesitic volcanic activity in Caribbean islands. Wise and Weaver (1973) and Weaver and Wise (1974) examined several Lower to Middle Eocene formations in South Carolina and Alabama that consist of opaline sediments and concluded that these sediments were deposited by biogenic activity based on the relative abundance of fossils and lack of zeolite. Carver (1980) studied the Congaree Formation of South Carolina, considered by Wise and Weaver (1973) to be time equivalent to the Tallahatta Formation of Mississippi and Alabama, and concluded the opaline silica was of biogenic origin. Laws and Thayer (1992) have also supported the idea of a biogenic origin for the Tallahatta Formation based on scanning electron microscopy, X-ray diffractometry and petrographic study of samples from Alabama.

ANALYTICAL PROCEDURES

Outcrop sections and cores containing the Tallahatta Formation were analyzed to determine if the opal-CT is dominantly biologic or volcanic in origin. Stratigraphic sections were compiled from cores through the entire Tallahatta Formation and from outcrops to determine the vertical distribution of sediments on the basis of gross lithology. X-ray diffraction analysis (XRD), scanning electron microscopy (SEM), chemical analyses and petrographic investigations were used to determine the mineralogical, microfabric, chemical and fossil characteristics of opal-CT-rich samples.

Nine cores, each approximately 50 meters long, were used in this study. Samples of the core were collected only in the claystone lithology for purposes of XRD, SEM, petrography and chemical analyses. Claystone samples were collected at 60-cm intervals, except at variations in lithology (such as change in color, presence of sand lenses, mica or glauconite, or due to chertification), where samples were collected at 30-cm intervals. Field samples were collected from three exposures, the Super Dry Mining site, Basic City Railway section and Valley Road section (Figure 1).

Because opal-CT is microcrystalline, the most reliable way to identify its presence is through XRD. A total of 577 samples were examined. All samples were ground to less than 63 microns. A Phillips-Norelco diffractometer with copper K_α radiation at 40 kilovolts and 20 milliamps was used. The scanning rate was 1° (2θ) per minute from 4° to 40° 2θ. Using X-ray diffractograms, opal-CT along with other associated minerals such as clinoptilolite, smectite group clays, and quartz were quantified using internal standards (Kabir, 1998). Such information allows major mineralogy to be assigned throughout all the cores. From the quantified XRD data, thirteen samples having the highest percentage of opal-CT were selected for chemical analyses and petrographic examination. Petrographic slides from Roquemore (1984) along the Valley Road outcrops were also examined.

Scanning electron microscopy (SEM) was done on 30 samples to identify small minerals and examine the distribution of these minerals within pores of the claystone. Elemental analyses were obtained for individual minerals using an energy dispersive X-ray (EDX) system attached to the SEM. Ten of the 30 samples were selected for SEM examination on the basis of the highest opal-CT percentage from drill holes, plus two from the Super Dry Mining site and one from the Basic City Railway section (Figure 1). These thirteen samples were also used for chemical and optical analysis. The other seventeen samples for SEM examination were selected on the basis of different proportions of opal-CT and clinoptilolite. Samples were cemented to aluminum stubs with silver paint, coated with a thin film of gold-palladium alloy

Table 1. Whole rock chemical analysis results in weight percent. Sample numbers MCH and MGS refer to core hole numbers shown in Figure 1. MFS are field samples collected at the three outcrop locations in Figure 1.

	MCH 1-22	MCH 2-35	MCH 3-23	MCH 4-24	MCH 5-25	MCH 6-54	MCH 7-26	MCH 8-31	MGS 1-14	MGS 1-31	MFS 1	MFS 13	MFS 23
SiO_2	84.33	80.95	79.50	80.40	79.60	80.10	78.20	80.40	70.70	75.69	87.77	88.50	79.70
Al_2O_3	3.60	4.37	5.29	5.03	5.05	4.66	5.31	4.78	8.18	7.98	3.58	3.31	6.89
CaO	0.70	0.77	0.66	0.71	0.83	1.29	1.06	0.96	0.87	0.31	0.24	0.17	0.24
MgO	0.53	0.77	0.94	0.84	0.90	0.74	0.98	0.84	1.55	1.19	0.57	0.40	0.96
Fe_2O_3	1.74	1.83	2.40	1.96	2.06	2.06	2.15	1.59	3.24	3.76	1.10	0.81	1.99
MnO	0.01	< 0.01	0.01	< 0.01	0.01	< 0.01	< 0.01	< 0.01	0.01	< 0.01	< 0.01	< 0.01	< 0.01
K_2O	0.93	0.85	1.17	1.18	1.01	0.94	1.18	0.94	1.54	1.02	0.66	0.42	1.10
Na_2O	0.17	0.16	0.16	0.14	0.14	0.55	0.16	0.12	0.16	0.12	0.11	0.11	0.12
P_2O_5	0.01	0.01	0.04	0.05	0.03	0.05	0.04	0.03	0.05	0.06	0.03	< 0.01	0.03
TiO_2	0.39	0.28	0.48	0.47	0.38	0.42	0.36	0.24	0.58	0.57	0.20	0.17	0.38
Cr_2O_3	0.03	0.02	0.03	0.03	0.01	0.04	0.01	0.01	0.01	0.03	0.01	0.01	0.02
LOI	6.21	7.95	7.49	7.12	7.85	7.08	8.16	7.90	10.90	8.76	5.40	5.48	8.23
Total	98.65	97.96	98.17	97.93	97.86	97.93	97.61	97.81	97.87	99.49	99.67	99.38	99.66
$+H_2O$	1.84	2.18	2.33	2.48	2.80	1.81	2.30	1.99	3.83	2.76	1.44	1.39	2.41
$-H_2O$	3.25	4.23	4.16	4.40	3.66	3.63	4.07	4.44	5.39	5.17	3.24	3.53	5.04

and examined using a Joel JSM-6100 instrument.

Whole rock chemical analyses were performed on the 13 samples to check the quantified XRD results, to confirm the necessary constituents were available for formation of opaline materials and clinoptilolite, and to compare with the available data for opal-CT elsewhere in the Tallahatta Formation (Hastings and McVay, 1963; Laws and Thayer, 1992). Results of the whole rock analyses are listed in Table 1. Chemical analyses were performed by Chemex Labs of Reno, Nevada using inductively coupled plasma emission spectroscopy (ICP-AES). Detection limit for all elements is 0.01 weight percent.

RESULTS

X-ray diffraction data show the two main claystone facies minerals of the Tallahatta Formation are opal-CT and clinoptilolite. Quantified XRD data indicate one or both of these minerals comprise about 80 percent of all samples (Kabir, 1998). Other identified minerals were quartz, smectite group clays, and muscovite. The classification of Jones and Segnit (1971) was followed for the XRD identification of opaline silica. The 4Å cristobalite d_{101} peak (major peak) for opaline silica identified in 373 of the 577 samples is in the range of 4.09Å to 4.14Å (Table 2). The adjacent tridymite peak (Figure 3) confirms the opaline silica as opal-CT. Seventy six percent of the 373 samples containing detectable opal-CT have d-spacing of 4.11Å to 4.12Å (Table 2), indicating strongly disordered opal-CT (Graetsch, 1994) is the most common phase of opal present. Opal-C, which was recognized by Jones and Segnit (1971) as a sharp, symmetrical peak at 22.2° 2θ (4.04Å), was not found in any of the samples. Opal-A, which can be recognized by a single diffuse XRD band centered around 21.6° 2θ (Jones and Segnit, 1971), also was not found.

SEM photomicrographs indicate the opal-CT is typically massive and structureless. It can also be formed within cavities as spherical crystal aggregates (Figure 4) similar to those described by Flörke and others (1975). These growths are 3 to 10 μm in diameter, and are described by

Table 2. Results of XRD analyses of opal-CT samples given in relative abundance for particular d_{101}-spacings. MCH and MGS designations refer to core hole numbers shown in Figure 1. MFS are field samples collected at the three outcrop locations shown in Figure 1.

Sample group	Samples for XRD	Samples with opal-CT	d_{101}-spacing (Å)					
			4.09	4.10	4.11	4.12	4.13	4.14
MCH-1	73	29	0	11	12	5	1	0
MCH-2	70	44	0	6	24	13	1	0
MCH-3	51	31	0	2	11	15	2	1
MCH-4	44	31	0	1	18	11	0	1
MCH-5	65	45	0	7	15	14	9	0
MCH-6	72	50	1	1	21	12	14	1
MCH-7	73	49	0	1	21	12	14	1
MCH-8	56	33	1	3	15	14	0	0
MGS-1	36	27	1	1	8	14	3	0
MFS	37	34	0	3	11	16	4	0
Total	577	373	3	36	156	126	48	4

Calvert (1977) as being composed of bladed crystals 30 to 50 nm thick. Wise and Kelts (1972) referred to these types of spherical aggregates as "lepispheres."

Clinoptilolite is the other common mineral in the study area. Quantified XRD results indicate that 73 percent of all Tallahatta claystones cored in the Meridian area had clinoptilolite as the dominant mineral present, consistent with the results of Spencer (1983). Clinoptilolite is block or coffin shaped. It can also be formed as flakes by dissolution of opal-CT lepispheric blades to which alkali and alkaline earth elements were added (Riech, 1979). Thin sheets of authigenic smectite clay often coat clinoptilolite crystals.

The high abundance of SiO_2 (70 to 90 percent) confirms that the quantified XRD method successfully characterized the major mineralogy. Samples were chosen for chemical analyses based on XRD determination that they were primarily opal-CT, and the SiO_2 content for all samples was above 70 percent, with most near 80 percent SiO_2 (Table 1). Similar levels of

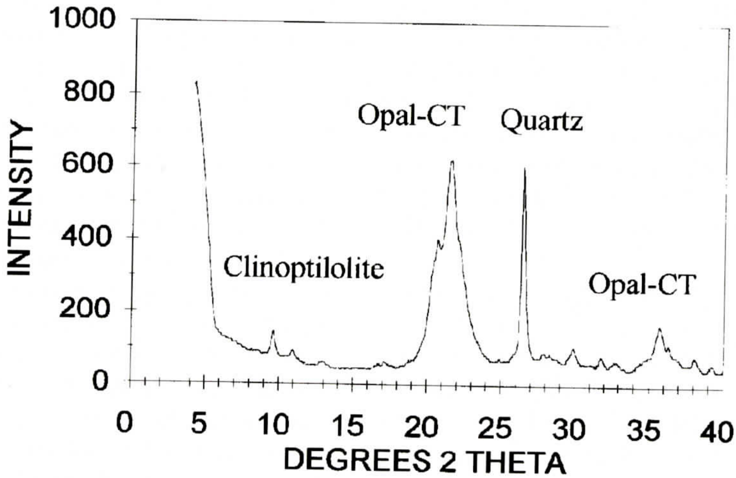

Figure 3. Example XRD pattern of sample MCH 6-54 showing opal-CT with clinoptilolite and quartz.

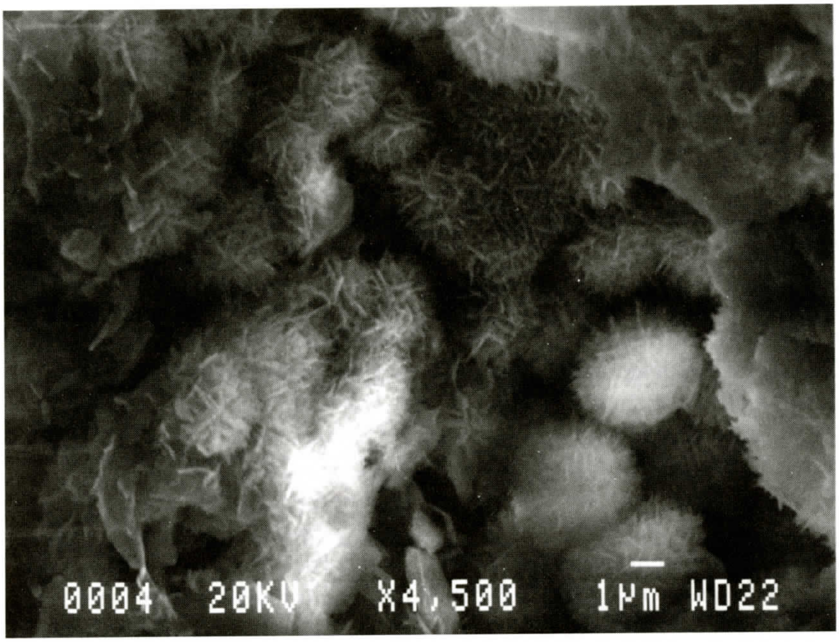

Figure 4. SEM image of opal-CT lepisphere in cavity of sample MGS1-31.

SiO_2 were found by Hastings and McVay (1963) and Laws and Thayer (1992) elsewhere in the Tallahatta Formation.

There is a strong correlation between Al_2O_3 and Fe_2O_3, which are probably contained in aluminosilicate minerals identified by XRD or visually-identified in the core, such as clinoptilolite, smectite clay or muscovite. Higher amounts of Al_2O_3 are associated with lesser amounts of SiO_2.

The other common mineral-forming cations, represented by the oxides CaO, MgO, and K_2O are present at levels commonly in the range of 0.5 percent to 1.5 percent. Na_2O is much less abundant, generally below 0.2 percent. Titanium is present at low levels, possibly substituting for silicon in silicate minerals or as trace occurrences of titanium-bearing minerals such as rutile.

Thirty-three thin sections were examined, 20 from the opal-CT and zeolite section along the Valley Road outcrop sections (Figure 1) and one from each of the 13 samples used for chemical analysis. Samples consist of a fine-grained massive combination of isotropic and anisotropic minerals. Angular detrital grains of silt and very fine sand-sized quartz are the largest anisotropic mineral grains present. The groundmass, where anisotropic, appears to consist of less than 5-micron-sized tabular grains, probably zeolite and phyllosilicates. Green ellipsoidal to spherical glauconite grains, 20 to 50 microns in diameter, were observed in all slides and occasionally found replacing the inner parts of larger fossils. Fossils constitute well less than 1 percent of the thin sections. Whole fossils or fossil fragments are rarely larger than 100 microns, and most are less than 50 microns. No mineral grains commonly ascribed to traditional volcaniclastic sediments, such as feldspars, biotite or euhedral quartz crystals (Carozzi, 1993) were observed in thin section.

DISCUSSION

Thin zones of opal-CT (porcellanite) formed under low-temperature conditions have been found in selected deep-ocean drilling sites near Antarctica (Bohrmann and others, 1990; Botz and Bohrmann, 1991; Bohrmann and others, 1994; Shipboard Scientific Party, 1999). These zones are relatively young (Pliocene to Pleistocene) and have been formed in water depths of 1.5 to 3 km at temperatures between 0° and

4° C (Botz and Bohrmann, 1991). Host sediments are composed of exceedingly pure (95 to 99 percent) biogenic opal-A lacking detectable clay minerals and quartz. The resulting porcellanites consist almost entirely of opal-CT (Botz and Bohrmann, 1991; Shipboard Scientific Party, 1999). The presence of even small amounts of detrital minerals, such as clays, will cause a reduction in the rate of opal-A to opal-CT transformation (Isaacs, 1982; Williams and others, 1985), thus explaining why these young opal-CT occurrences are restricted to several deep-ocean cores around Antarctica. Nonetheless, these deposits indicate that something more than temperature increases must drive the opal-A to opal-CT conversion (Botz and Bohrmann, 1991).

The Tallahatta Formation in the study area consists of an assemblage of opal-CT, clinoptilolite, smectite group minerals and quartz. This collection of minerals indicates the original sediment is unlikely to have been the high-purity opal-A material seemingly required to produce early diagenetic opal-CT under the low-temperature conditions described by Bohrmann and others (1994).

The sequence of change from opal-A contained in diatoms to opal-CT found in diatomite is well established. According to Hein and others (1978), in the first 300 to 400 meters of burial diatom frustules are fragmented, which increases the surface area available for the associated mild dissolution. By 600 meters, the dissolution of opal-A is advanced and opal-CT precipitates abundantly between 600 to 700 meters depth. The Monterey Formation in California has a typical burial depth for the transformation of diatomaceous opal-A to opal-CT of 700 meters (Murata and Larson, 1975). The temperature required for the transformation of opal-A to opal-CT is 35° to 55° C (Calvert, 1983). Because the depth of burial is relative to the geothermal gradient, Hein and others (1978) suggest 500 meters of diatomaceous sediment may be required to obtain sufficient temperature at the base of the siliceous section. Murata and others (1977) estimate that the complete transformation of opal-A to disordered opal-CT occurs at a temperature of about 50° C and at a burial depth of 900 to 1000 meters. Nähr and others (1998) show a transition from opal-A to opal-CT in deep sea drilling results at depths of 340 to 370 meters. Isotopic temperatures determined by these authors from authigenic clinoptilolite range from about 25° to 65° C. All of these studies, along with Isaacs (1982) and Williams and others (1985) indicate several hundred meters of burial and mildly elevated temperatures are required for the conversion of opal-A to opal-CT when deposited with other types of sediments containing detrital minerals.

It appears that the Tallahatta Formation never achieved burial to depths great enough for transformation of opal-A to opal-CT. After studying the Eocene Congaree Formation of South Carolina, which is time equivalent to the Tallahatta Formation in Mississippi, Carver (1980) concluded that this unit was never buried to depths greater than 100 meters. This depth of burial was too shallow to allow biogenic opal-A to completely transform into opal-CT, and thus some opal-A should be detectable using XRD in the Tallahatta sediments. Because no opal-A was detected in any of the 577 samples examined using XRD, we suggest non-biogenic (volcanic) silica sources may well have been a major contributor to formation of the opal-CT present in the Tallahatta Formation in eastern Mississippi. The opal-A reported by Laws and Thayer (1992) in the Tallahatta Formation of Alabama was inferred based on the abundance of diatoms observed.

Brasier (1980) suggests over six million diatom frustules are required for one cubic centimeter volume of diatomite (opal-A). Murata and Larson (1975) indicate opal-A is present in the upper parts of the Miocene Monterey shales they studied to document the biogenic opal-A to opal-CT transition. With increasing diagenesis the structural disorder decreases according to the reaction path: non-crystalline opal (opal-A) $\rightarrow$ opal-CT $\rightarrow$ opal-C $\rightarrow$ quartz in sea floor deposits of silica (Calvert, 1977; Williams and others, 1985). Out of 577 samples analyzed from the Tallahatta Formation, not a single sample contained any opal-A and the d-spacings (mostly 4.11 to 4.12Å) signify opal-CT. In volcanically-derived deposits of silica, opal-C and

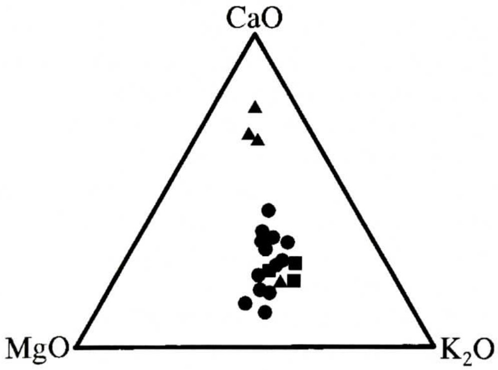

Figure 5. Ternary plot of CaO, MgO and K_2O from whole rock chemical analysis results. Circles are results listed in Table 2 of this study, squares are data from Hastings and McVay (1963), and triangles are data from Laws and Thayer (1992).

opal-CT seem to be formed directly without the initial development of opal-A (Iijima, 1978; Flörke and others, 1991). Because all silica-rich microorganisms initially precipitated opal-A before further change, biogenic material does not appear to be dominant in the formation of opaline material in the study area. The predominance of biogenic silica in marine sediments (Calvert, 1983; Williams and Crerar, 1985) does not preclude the occurrence of opaline sediments primarily of non-biogenic (volcanic) origin.

Based on the quantified XRD and chemical analyses, the samples chosen for thin section examination should have the highest percentage of opal-CT present. If the opal-CT is solely of biogenic origin, then the greatest abundance of fossils could be expected to be present in these zones. Weaver and Wise (1974) indicate that fossils were most abundant in their samples that had 60 percent to 90 percent SiO_2, and that siliceous microfossils were observed in 90 percent of the samples they examined. The general lack of fossils, particularly in the abundances noted by Wise and Weaver (1973), Weaver and Wise (1974), or Laws and Thayer (1992) in their samples from Alabama, indicates either the dissolution-reprecipitation process involving opal-A was more complete and effectively eliminated fossils or that opal-CT around Meridian, Mississippi probably had a different origin.

The CaO content reported in the upper Tallahatta samples collected by Laws and Thayer (1992) ranges from 3.39 to 5.57 weight percent, well above levels of CaO reported in Hastings and McVay (1963) or the current study (Table 1). This is easily illustrated by comparing the relative abundances of CaO, MgO and K_2O within the three studies (Figure 5). Laws and Thayer (1992) attribute the elevated CaO to the presence of calcite skeletal fragments occurring within the clay lenses and partings. The relatively low CaO in the samples from the Meridian study area is consistent with the lack of fossils observed in thin sections and SEM.

Eocene opaline silica found in Alabama, Georgia and South Carolina as reported by Wise and Weaver (1973), Weaver and Wise (1974), Carver (1980), and Laws and Thayer (1992) lacks appreciable zeolites. They concluded that the opal-CT in those areas is mainly biogenic in origin. Laws and Thayer (1992) state that zeolitic beds in the Tallahatta Formation of Alabama are volumetrically minor components and the presence of zeolite is insignificant to assume the whole formation was volcanogenic in origin. Quantitative XRD data (Kabir, 1998) indicates the zeolite clinoptilolite is a major constituent of claystone in the study area, where three-fourths of the claystones have clinoptilolite as the dominant mineral present. Both Wise and Weaver (1973) and Laws and Thayer (1992) acknowledge a volcanic origin for the zeolite minerals found in the Tallahatta Formation. Authigenic clinoptilolite has been observed in other marine sediments at about 5 volume percent (Nähr and others, 1998).

Clinoptilolite has long been known to be formed during alteration of volcanic ash (Hay, 1966; Hay and Sheppard, 2001). Broxton and others (1987) document extensive diagenetic development of clinoptilolite and smectite group clays in nonwelded glassy layers of ash flow tuffs in southern Nevada. Reynolds (1966, 1970) has shown that volcanic ash of rhyolitic composition contains the necessary ions to form clinoptilolite. Kizaki (1965) observed cores of volcanic ash shards where the outer rim is smectite clay and the inner part is clinoptilolite. It has been suggested (Kerr, 1931; Kerr and Cameroon, 1936; Bramlette and Posnjak, 1932) that zeolite minerals such as clinoptilolite are intermediate components formed from the glass that is then altered to smectite clay.

Numerous authors have noted the presence of volcanically-derived materials in Eocene rocks of the Gulf Coast region. Evidence of Eocene volcanism has been reported by Hunter and Davies (1979) in a wide area ranging from Mississippi to the Veracruz basin in Mexico. Hunter and Davies (1979) further mentioned that rocks of the Claiborne Stage (Figure 2), which include the Tallahatta Formation, record volcanic detritus (tuff and volcanic fragments) in the Gulf Coast. Senkayi and others (1987) describe clinoptilolite associated with opal-CT and kaolinite formed during weathering of interbedded tuff and lignite in Late Eocene sediments of southern Texas. Heron (1969) suggested that continental volcanoes contributing siliceous ash formed the clinoptilolite of the Eocene mudrocks in South Carolina. He further claimed that the ash originated in Mexico and the western United States and was carried to the east by wind and water currents similar to the present Gulf Stream. After examining the Universiad Formation of Eocene-Miocene age in Cuba, Gibson and Towe (1971) found the clinoptilolite-opal/cristobalite-montmorillonite suite indicative of altered volcanic glass and assumed that the actual volcanic center must have been distant, likely south of the United States. Matson and Pessagno (1971) mentioned dacitic volcanism occurred elsewhere in the Caribbean from Late Cretaceous to Early Tertiary, distributing volcanogenic material to the north and east of the source region.

The presence of smectite clay in Tallahatta claystone, even though a minor constituent, is important in determining the origin of opaline materials. Reynolds (1970) considers smectite clay, where associated with cristobalite and zeolite, to be formed from volcanic material. Iijima (1978) and Barbieri and others (1981) conclude that the clinoptilolite, opal-CT and smectite clay association indicates an origin by alteration of volcanic glass in marine sediments that underwent burial diagenesis. Roquemore (1984) describes a one-meter bed of smectite clay that grades upward into glauconite sand in the uppermost Tallahatta Formation from the Valley Road section (Figure 1), a mineralogic sequence consistent with bentonite clays found in Upper Cretaceous and Tertiary formations in northeast and central Mississippi (Grim and Güven, 1978). We suggest that the mineral assemblage present in the Tallahatta Formation in eastern Mississippi, which includes clinoptilolite, opal-CT and smectite group clays, is indicative of a predominantly volcanic origin for the opaline material.

CONCLUSIONS

The Tallahatta Formation in the vicinity of Meridian, Mississippi, consists of glauconitic claystone with significant amounts of opal-CT, clinoptilolite, subordinate amounts of smectite group clays and muscovite. Petrographic study of the claystone indicates a massive opaline structure with few recognizable fossils. Eocene units in the study area show no evidence they were buried to depths sufficient to convert opaline diatomite (opal-A) to opal-CT. The lack of opal-A in the Tallahatta claystone of eastern Mississippi argues that the precursor of opal-CT was not opal-A, thus the opal-CT most likely was from non-biogenic, i.e. volcanic, sources. The Tallahatta Formation eastward from Alabama to South Carolina changes to a more fossiliferous opal-CT with a significant decrease in clinoptilolite, more likely the result of biogenic precipitation with only a minor volcanogenic component. The opal-CT-clinoptilolite-smectite mineral assemblage of the Tallahatta Formation in eastern Mississippi supports an origin from the alteration of volcanic ash in a shallow marine environment.

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REVISED STRATIGRAPHY AND NOMENCLATURE FOR THE UPPER HINTON FORMATION (UPPER MISSISSIPPIAN) BASED ON RECOGNITION OF REGIONAL MARINE ZONES, SOUTHERN WEST VIRGINIA

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ABSTRACT

The Hinton Formation is a lithologically variable succession of dominantly siliciclastic strata known to comprise intercalated nonmarine and marine facies. Previous attempts to subdivide the formation into traceable members have been flawed by excessive naming and miscorrelation of localized units. The present study abandons all previous nomenclature for subdivisions of the upper part of the formation (above the Avis Limestone Member), and proposes a new stratigraphic framework that recognizes only two formal units, herein named the Fivemile and Eads Mill members. Given that both of these units are characterized by the presence of marine fossils, they both represent significant marine transgressions in the central Appalachian basin. The revised stratigraphy also reveals a general cyclothemic architecture in the upper Hinton succession much like that found in Pennsylvanian formations in the basin. The new stratigraphic framework should provide a sound basis for more in depth stratigraphic and sedimentological analysis of the upper Hinton Formation.

INTRODUCTION

The Upper Mississippian Mauch Chunk Group of southern West Virginia is a southeast-thickening clastic wedge consisting of paralic facies deposited in the central Appalachian foreland basin during the onset of the Late Paleozoic ice age (Donaldson and Shumaker, 1981; Englund and Thomas, 1990; Miller and Eriksson, 2000). Although high-frequency transgressive-regressive cycles (i.e., cyclothems) occur in Upper Mississippian as well as Pennsylvanian successions in North America (Weller, 1930; Wanless and Shepard, 1936; Veevers and Powell, 1987), relatively few studies (Whisonant and Scolaro, 1980; Schalla, 1984; Miller and Eriksson, 1999, 2000) have been published on the nature and origin of depositional cycles in the Mauch Chunk Group. Studies of this sort have been hindered by the lack of a practical stratigraphic framework that reflects major marine transgressive events in the otherwise nonmarine succession. Although the group consists of four well-established formations (Bluefield Formation, Hinton Formation, Princeton Sandstone, and Bluestone Formation; see Figure 1), member-rank subdivisions have remained problematic for stratigraphic analysis.

In this paper, the lithostratigraphy of the upper part of the Hinton Formation is reviewed and revised. Previous studies (Reger, 1926; Englund and Thomas, 1990; Miller and Eriksson, 2000) have suggested the existence of major transgressive events in the upper Hinton, but have not delineated regional marine zones in the outcrop belt. The present study shows that the upper Hinton Formation comprises two regional marine zones, herein named the Fivemile and Eads Mill members. The lithology of these units is described and they are assigned composite stratotypes in accordance with the North American Stratigraphic Code (North American Commission on Stratigraphic Nomenclature,

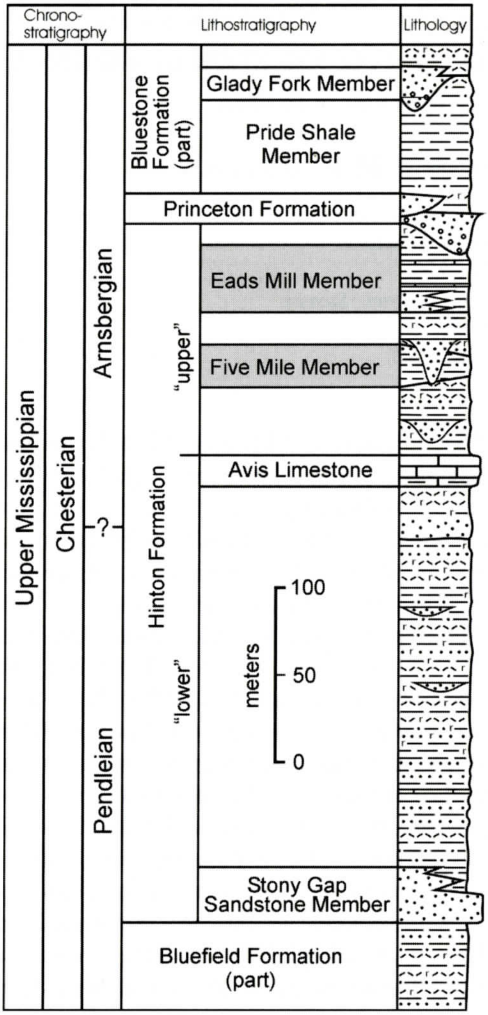

Figure 1. General stratigraphy of the Hinton Formation in southern West Virginia with newly described marine members indicated (shaded gray).

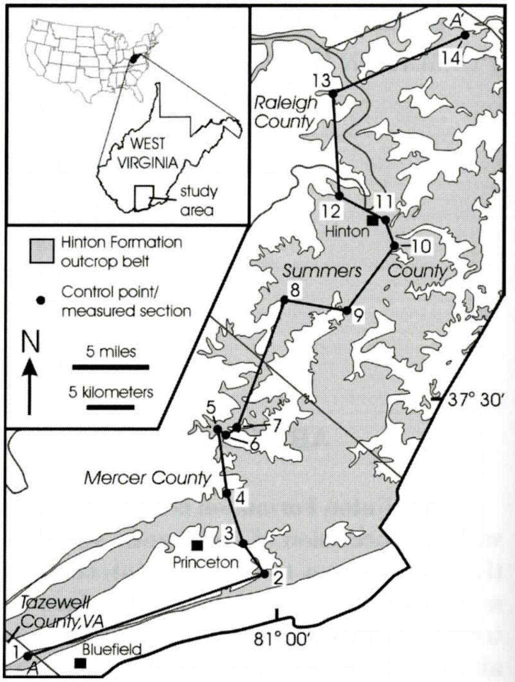

Figure 2. Study area showing outcrop of Hinton Formation (shaded), location of control points, and line of section shown in Figure 3. Composite stratotype for Fivemile Member comprises section 2, 8, and 9. Composite stratotype for Eads Mill Member comprises sections 5 and 6. Sections are (1) VA Route 102; (2) U.S. Route 460; (3) I-77; (4) Mercer Springs Road; (5) Eads Mill; (6) Brush Creek Canyon; (7) White Oak Creek; (8) Suck Creek; (9) True Road; (10) Mt. Zion Church Road; (11) Elk Knob; (12) Madam's Creek; (13) Farleys Creek; and (14) Fisher Creek.

1983). Also, the significance of the Fivemile and Eads Mill members for cyclothem analysis of the upper Hinton Formation is discussed briefly.

METHODS

Data for this study were collected from high-quality outcrop exposures throughout Mercer, Summers and Raleigh counties in West Virginia and Tazewell County, Virginia (Figure 2). The lithologically distinctive Avis Limestone of Reger (1926) was used as the primary stratigraphic datum. Well-exposed sections were described and used to construct a stratigraphic cross section (Figure 3) that illustrates the lateral continuity of the new marine members. Measured sections are on file and available from the West Virginia Geological and Economic Survey (see Figure 2 for locations and section names).

Previous studies have recognized and named individual marine fossiliferous beds within the upper Hinton (e.g., Reger, 1926). In the present study, these beds were integrated into genetically related packages of facies that record regional transgressions (i.e., marine zones). The

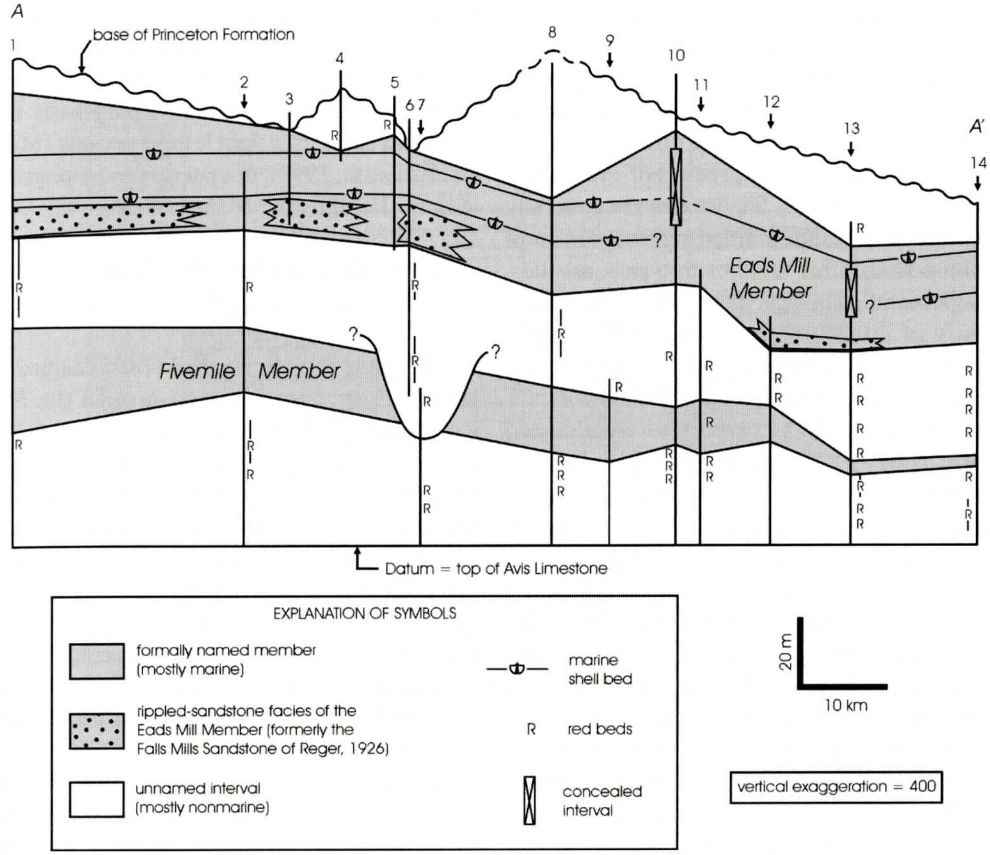

Figure 3. Stratigraphic cross section of upper Hinton Formation illustrating regional continuity and lithological character of Fivemile and Eads Mill members. See Figure 2 for location of cross section.

"marine zone" concept used herein has been used previously to revise and refine the lithostratigraphic framework of the Middle Pennsylvanian Kanawha Formation of West Virginia (Blake and others, 1994; Martino, 1994, 1996). As envisioned by Martino (1994), marine zones comprise a mosaic of marine and marginal-marine facies.

Since marine zones are lithologically distinctive units, they can be designated as formal lithostratigraphic units. In accordance with the North American Stratigraphic Code (North American Commission on Stratigraphic Nomenclature, 1983), this study has used general fossil content in conjunction with other lithic characteristics to define members in the study interval.

STRATIGRAPHY, MEGAFAUNA, AND DEPOSITIONAL ENVIRONMENTS

The upper Hinton Formation, up to 150 meters thick, extends from the top of the Avis Limestone of Reger (1926) to the base of the Princeton Sandstone (Figure 1). Where the Princeton is absent, the upper contact is placed at the top of the highest red bed subjacent to the Pride Shale Member of the Bluestone Formation (Campbell, 1896; Beuthin and others, 2000; Blake and others, 2000). Brachiopods (Gordon and Henry, 1981; Henry and Gordon, 1992) and conodonts (Stamm, 1997) support correlation with the upper Chesterian Series of the mid-continent, and megaflora (Blake and others, 2002) support correlation with the lower

Namurian (Arnsbergian) of western Europe. Lithologically, the upper Hinton Formation consists of a highly variable succession of mudstone, sandstone, and limestone with limited occurrences of coal. Mudstones are predominantly red and green, but gray mudstones are also found. Sandstones are generally very fine- to medium-grained and range in composition from lithic arenites to quartz arenites. Conglomeratic lenses within sandstones consist mostly of intraformational clasts. Sandstones are generally of limited areal extent and are not mappable on a 1:24000 scale (Beuthin and others, 2000; Blake and others, 2000; Beuthin and Blake, 2001; Blake and others, 2001; Beuthin and Blake, 2002a; Blake and Beuthin, 2002). Limestones include marine shell beds, thinner epifaunal bivalve-dominated bioherms, and argillaceous lime mudstones with a sparse nonmarine fauna. Calcareous glaebules also occur within red mudstones (caliches). "Coal" beds are generally thin (cm-scale) and impure, and grade laterally into carbonaceous shale. These carbonaceous shale beds commonly contain ostracodes and bivalves (Beuthin and Blake, 2002b). Carbonaceous rocks are associated commonly with gray mudstone that contain lycopsid roots and stems (Beeler, 1999).

The megafauna of the upper Hinton Formation, as first described in Reger (1926), comprises brachiopods, bryozoans, bivalves, gastropods, ostracodes, trilobites, and pelmatazoans. Cooper (1961) also reported cephalopods from beds near the top of the formation (upper part of the Eads Mill Member of the present study). In their biostratigraphic assessment of Upper Mississippian brachiopods in the central Appalachian basin, Henry and Gordon (1992) reported 15 genera from the upper Hinton Formation. In a taxonomic study of bivalves from the Mauch Chunk Group, Hoare (1993) reported 19 genera from the upper Hinton Formation, all from strata assigned to the Fivemile and Eads Mill members of the present study.

Deposition occurred in fluvial, coastal, and shallow-marine environments (Englund and others, 1981; Englund and Thomas, 1990; Miller and Eriksson, 2000; Beuthin, 2001; Huffman and Beuthin, 2001; Portner and Beuthin, 2001; Beuthin and Blake, 2002b). Climate was relatively dry, ranging from seasonal to semi-arid (Cecil, 1990; Miller and Eriksson, 1999; Beuthin and Blake, 2002b). Coaly beds may represent either transient humid periods (Miller and Eriksson, 1999), or waterlogged topographic lows (Beuthin and Blake, 2002b).

ABANDONMENT OF PREVIOUS NOMENCLATURE

Campbell and Mendenhall (1896) named the Hinton Formation for exposures in the New River Gorge around Hinton, West Virginia, but did not recognize any subdivisions other than the Stony Gap Sandstone at the base of the formation. Commonly used member-rank nomenclature is derived largely from stratigraphic schemes developed by Reger (1926) and Englund (1968) (see Figure 4). For reasons articulated below, both of these schemes created more stratigraphic confusion than clarity, and therefore should be abandoned, except for the Avis Limestone of Reger (1926).

Reger (1926) conducted the first detailed stratigraphic study of Upper Mississippian rocks. He recognized 22 members in the upper Hinton Formation and designated type sections for most of them (Figure 4). However, recent geologic mapping at the 1:24000 scale has found that, generally, these members are localized units that have little stratigraphic utility beyond their type locality (Beuthin and others, 2000; Blake and others, 2000; Beuthin and Blake, 2001; Blake and others, 2001; Beuthin and Blake, 2002a; Blake and Beuthin, 2002). Reger's stratigraphy is also flawed as a result of numerous miscorrelations. For example, Reger applied the name "Pluto Coal" to a carbonaceous shale near the top of the Hinton Formation; however, the type Pluto Coal of Krebs and Teets (1916) actually occurs in the Bluestone Formation (Meissner, 1981). Reger also miscorrelated other units, as shown later in the present study. In addition, Reger (1926) assigned the same geographic name to several different, discontinuous units (e.g., Tallery Sandstone, Tallery Limestone, Tallery Coal). This practice is inconsistent with modern strati-

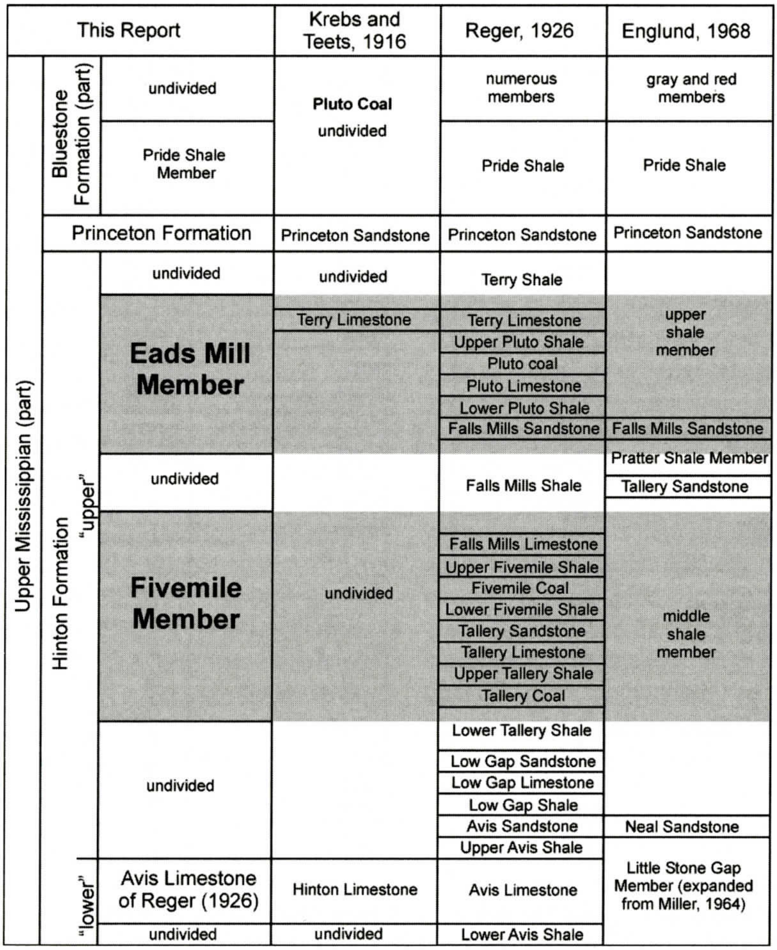

Figure 4. Correlation chart for the upper Hinton Formation comparing previous stratigraphic nomenclature with members defined in the present study.

graphic procedure.

Englund (1968) developed a six-member scheme for mapping purposes (Figure 4). The name "Neal Sandstone" was proposed as a substitute for the Avis Sandstone of Reger (1926). However, the type Neal Sandstone (near Bluefield, Virginia) is about 58 km southwest of the type Avis Sandstone (near Hinton, West Virginia), and Englund (1968) did not demonstrate that the two units are correlative. Also, the Neal Sandstone is absent in the Virginia Route 102 section of this study (section 1 on Figures 2 and 3), which is about 200 meters west of the type section. Hence, the Neal Sandstone appears to be a localized facies, rather than a mappable unit. Englund also created confusion by apparently (but not explicitly) redefining the Tallery Sandstone of Reger (1926). The type Tallery is a shaly sandstone that occurs below Reger's (1926) Fivemile interval. However, Englund (1968) applied the name "Tallery Sandstone" to a resistant, quartzose sandstone that occurs above Reger's (1926) Fivemile interval (Figure 4). This confusion has been compounded by recent studies which have applied the name "Tallery" to any massive, quartzose sandstone in the upper Hinton Formation (Englund and Thomas, 1990; Miller and Eriksson, 1999, 2000; Sullivan and Allen, 2001).

In essence, we find that existing member-rank nomenclatural schemes for the upper Hinton Formation (Reger, 1926; Englund, 1968)

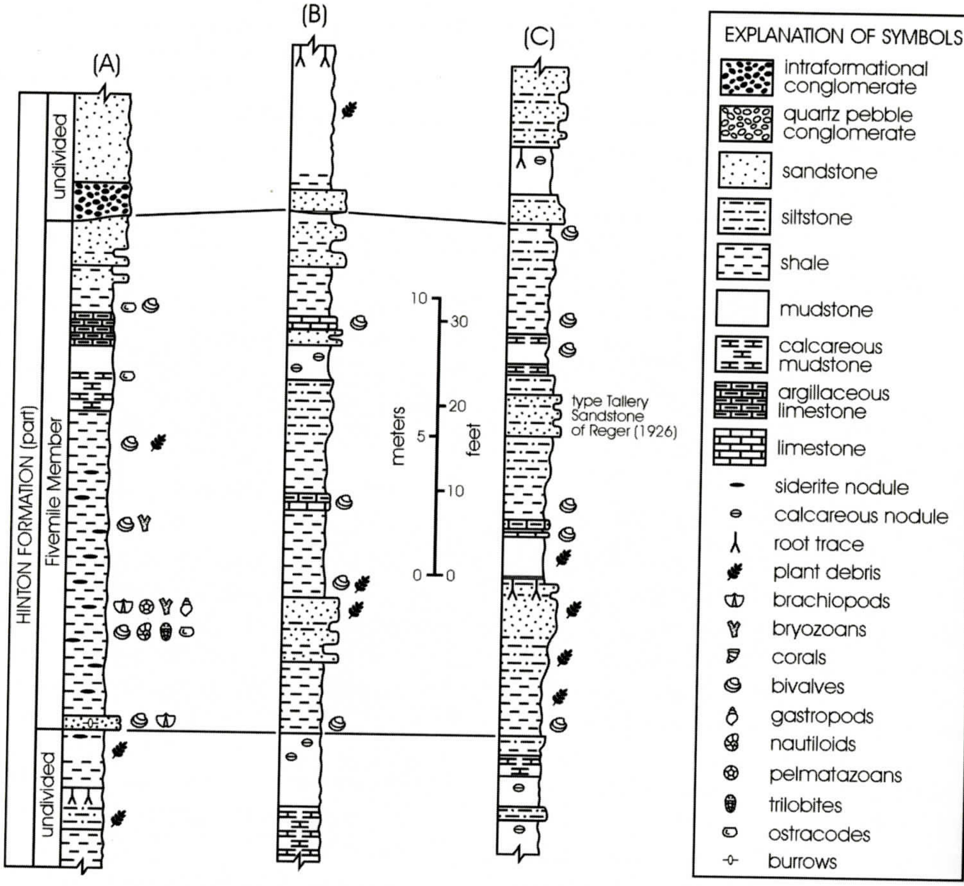

Figure 5. Geological columns illustrating composite stratotype of Fivemile Member. (A) U.S. Route 460 section (section 2 on Figures 2 and 3); (B) Suck Creek section (section 8 on figures 2 and 3); (C) True Road section (location 9 on Figures 2 and 3).

are confusing, impractical, and prone to misuse. Hence, we abandon these schemes.

NEW STRATIGRAPHIC FRAMEWORK

The Fivemile Member (Redefined)

A succession of strata, including Reger's (1926) Upper Fivemile Shale, is redefined and revised as the Fivemile Member in the present study (Figures 1 and 4). The base of the Fivemile Member ranges from 20 to 43 meters above the top of the Avis Limestone (Figure 3). The member, ranging from 3 to 27 m thick, consists mostly of fossiliferous shale in and around its type area (eastern Mercer County). In other parts of the study area, the member is more heterogeneous and comprises varying proportions of mudstone, siltstone, and sandstone, as well as fossiliferous shale. In the present study, a composite reference section is designated that characterizes the lithologic variability of the unit (Figure 5).

In the type area of Reger's (1926) Upper Fivemile Shale (along Fivemile Creek in eastern Mercer County), the redefined Fivemile Member is well exposed in a roadcut along the westbound lane of U.S. Route 460, approximately 5.2 km east of the junction of Route 460 and Interstate 77. This locality (section 2 in Figures 2 and 3), which is better exposed than Reger's (1926, p. 337) still extant Princeton-Narrows road section, 0.8 km to the west, serves as the principal reference section for the Fivemile Member. In the Route 460 section, the member is 18.6 meters thick (Figure 5A), and comprises strata that Reger (1926) assigned to the Upper Fivemile Shale, the Falls Mills Limestone, the Falls Mills Shale and the Falls Mills Sandstone in his Princeton-Narrows road section (his units 3-8). The lower contact of the Fivemile Member is sharp and is recognized at the base of a thick bed of bioturbated, very fine-grained sandstone that contains broken brachiopods, bivalves, and unidentifiable shell material. This sandstone overlies a seven-meter-thick composite bedset of carbonaceous shale and green to gray mudstone, siltstone and shale (the Lower Fivemile Shale and Coal of Reger, 1926). Plant debris is common in this succession, and some of the mudstone and siltstone beds contain root traces. The fossiliferous sandstone at the base of the Fivemile Member is overlain by 11 meters of gray fossiliferous shale with siderite nodules (the Upper Fivemile Shale of Reger, 1926). This shale carries a diverse marine fauna consisting of corals, brachiopods, bryozoans, trilobites, ostracodes, conularids, cephalopods, bivalves, gastropods, and pelmatazoans. The fossiliferous shale is overlain by a bedset of unfossiliferous calcareous mudstone and argillaceous limestone (the Falls Mills Limestone of Reger, 1926). The topmost strata of the member consist of a coarsening-upward succession (CUS) of gray shale, silty shale and sandstone. The shale in this CUS, which equates to Reger's (1926) Falls Mills Shale, contains bivalves and ostracodes at the base. The sandstone is very fine-grained and laminated to ripple bedded. The Fivemile Member is sharply overlain by a 17-meter-thick complex of heterolithic channel-fills composed of interbedded sandstone, siltstone, and mudstone. Most of the beds are green to gray, but some of the mudstone beds are red and contain root traces. A thick bed of intraformational conglomerate at the base of this complex partially truncates the sandstone that caps the Fivemile Member. The Falls Mills Sandstone in Reger's Princeton-Narrows road section appears to equate to the lower part of this channel complex and the sandstone at the top of the Fivemile Member.

Throughout eastern Mercer County, the Fivemile Member retains the shale-dominated character exhibited in the Route 460 section, but carries a less diverse fauna dominated by myalinid and pectinid bivalves. The basal contact generally is sharp and is marked by a change from red, green, and gray beds containing plant debris and/or root traces to gray and green beds with common bivalves. Typically, the member is overlain by a scour-based, cross-bedded sandstone or intraformational conglomerate with overlying sandstone, much like in the Route 460 section. The erosion surface associated with these sandstone and conglomerate beds exhibits several meters of relief. About 13 kilometers north of its type area, the Fivemile Member has been completely removed by the channel-fill sandstone complex (sections 6 and 7 of Figures 2 and 3). Elsewhere, the Fivemile Member has been observed to grade up into a greenish-gray, rooted mudstone that is, in turn, truncated by an erosion surface.

Beyond eastern Mercer County, the Fivemile Member is more heterogenous as shown by supplementary references sections in Summers County (Figure 5B,C). Green to gray, fossiliferous shale remains a distinguishing lithological character; however, green to gray siltstone and sandstone, and thin to medium beds of gray limestone consisting of densely packed shell material (dominated by bivalves) are also typical. Locally in Summers County, a few beds of green and red mudstone are also found. Characteristically, the low-diversity megafauna is dominated by bivalves (myalinids and pectinids are common) and ostracodes. Rare occurrences of cephalopods and conularids have been noted. As in the type area, the lower contact is sharp; gray to green, fossiliferous shale at the base of the member typically overlies a mudstone-dominated facies succession that exhibits varying degrees of paleopedogenic development (common pedogenic features include root traces, pedogenic slickensides, and calcareous nodules). Typically, the basal fossiliferous shale is part of a meter-scale CUS capped by ripple-bedded sandstone. At most localities, the Fivemile Member is overlain by a scour-based, cross-bedded or ripple-bedded sandstone that fines up into mudstone that contains plant fossils or is paleopedogenically overprinted. At a few localities the upper contact is less distinct, with the member grading up into a succession of interbedded greenish-gray mudstone, siltstone, and sandstones with varying degrees of paleopedogenic overprint.

The True Road section of the Fivemile Member (section 9 of Figures 2 and 3; Figure 5C) was measured also by Reger (1926, p. 267-269). The member there generally comprises his units 16-21, and parts of units 15 and 22 (upper part of the Lower Tallery Shale through the lower part of the Falls Mills Shale). Of particular note, the type Tallery Sandstone (Reger's unit 18) occurs as sandstone facies within the Fivemile Member (Figure 5C). Also, the type Tallery Coal (Reger's unit 21) occurs as a thin, localized bed of carbonaceous shale (not coal) at the top of the coarsening upward cycle in the basal part of the member (Beuthin and Blake, 2002b).

Eads Mill Member (New Name)

In the present report, a marine zone near the top of the Hinton Formation is recognized and formally named the Eads Mill Member (Figures 1 and 4). The base of the Eads Mill ranges from 55 to 90 meters above the top of the Avis Limestone (Figure 3). This member comprises most, if not all of Englund's (1968) upper shale member, as well as Reger's (1926) Falls Mills Sandstone, and the upper part of his Falls Mills Shale (Figure 4). Although the Falls Mills Sandstone has been mapped as a member of the Hinton Formation (Englund, 1968), the present study regards this unit as an informal facies within the lower part of the Eads Mill Member; herein, it is referred to as the "rippled-sandstone facies" of the Eads Mill Member. Throughout the area, the lower contact of the member is typically marked by a sharp transition from rooted mudstone to dark gray or black shale containing ostracodes, bivalves, or both. At most localities, the Princeton Sandstone directly (and unconformably) overlies the Eads Mill Member (Figure 3). However, at some localities, the member grades up into an unnamed, heterogenous interval of Hinton strata that comprises sandstone, siltstone, mudstone, shale, and impure coal. Gray and green beds are common in this interval, but red beds are also abundant at some localities. The Eads Mill Member, ranging from 28 to 42 meters thick, consists mostly of gray to greenish-gray shale with interbedded shelly limestone. Shale beds are sparsely to moderately fossiliferous, and typically carry marine fauna dominated by brachiopods and bivalves. Shelly limestone beds, which are up to one meter thick, comprise a diverse fauna of brachiopods, bryozoans, corals, bivalves, cephalopods, gastropods, ostracodes, and pelmatazoans. Two of these shell beds are traceable across the study area, and serve as informal key beds within the member (Figure 3). Where the rippled-sandstone facies is present, the lower shell bed occurs about 2 meters above the sandstone. The upper shell bed typically occurs 9-11 meters above the lower shell bed. Locally, beds of green or gray mudstone, siltstone, and sandstone also are found (including the rippled-sandstone facies, which is described more fully below). The rippled-sandstone facies is prominently developed in the southern part of the study area (Mercer County, West Virginia and Tazewell County, Virginia), but is only locally present in the northern part of the area (Summers and Raleigh counties, West Virginia) (Figure 3). No red beds have been observed in this member.

The type section of the member is located along Mercer County Route 3 (Eads Mill Road) above the Bluestone River (section 5 on Figures 2 and 3). The name derives from the nearby rural settlement of Eads Mill. The type Eads Mill Member, 32 meters thick, consists of gray shale with interbedded fossiliferous limestone and calcareous mudstone (Figure 6A). A medium-thick bed of ostracode-bearing carbonaceous shale at the base of the member sharply overlies a rooted, greenish-gray mudstone. The top contact, also sharp, marks the base of a 2.6-meter-thick, greenish-gray sandstone. The Hinton-Princeton contact, which is an irregular erosional surface, occurs about five meters above the top of the Eads Mill Member. The rippled sand-

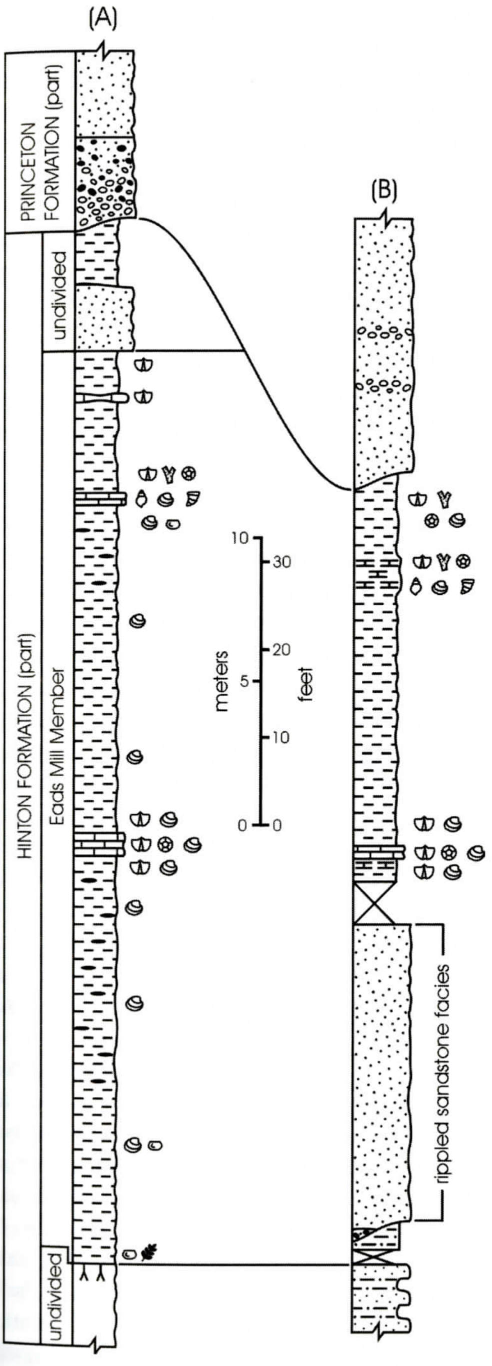

Figure 6. Geological columns illustrating composite stratotype of Eads Mill Member. (A) Eads Mill section (section 5 on Figures 2 and 3); (B) Brush Creek Canyon section (section 6 on Figures 2 and 3). Explanation of symbols on Figure 5.

stone facies is absent in the type section; however, it is present in the type area, locally forming a prominent cliff along the valley walls of the Bluestone River and its tributary, Brush Creek (Figure 6B). Outcrop tracing of the rippled sandstone facies indicates that it grades laterally into the lower part of the type Eads Mill Member. Reger (1926) did not measure the Eads Mill section, so there is no direct means of determining which of his members correlate with the Eads Mill Member. The type Eads Mill Member comprises the upper part of the Falls Mills Shale, the Lower Pluto Shale, the Pluto Limestone, the Pluto Shale, the Terry Limestone, and the lower part of the Terry Shale of Reger's (1926, p. 295-298) composite section for the Hinton Formation (Figure 4).

The rippled sandstone facies, as much as 10.3 meters thick, consists of gray, ripple-bedded sandstone, and some cross-bedded sandstone with minor interbeds of silty shale and siltstone. Where the facies is present, its basal contact is invariably sharp and relatively flat, with less than one meter of relief. Typically, the base occurs 40-100 centimeters above the base of the Eads Mill Member. The upper contact is intercalated and marks a facies change from sandstone to shale. Locally, portions of the rippled-sandstone facies are bioturbated.

Unnamed Intervals

Stratigraphic intervals in the upper Hinton Formation that occur directly below the Fivemile Member, between the Fivemile and Eads Mill members, and above the Eads Mill Member (Figure 3) remain undifferentiated and unnamed because they each lack inherently distinctive lithologic characteristics. In general, each of these intervals consists mostly of red and green mudstone with interbedded sandstone, calcareous mudstone/argillaceous limestone, and scattered carbonaceous shale or impure coal. Red beds, which often contain root traces or other pedogenic features, generally are more abundant in the northern part of the study area than in the southern part (Figure 3). In contrast to the Fivemile and Eads Mill members, these intervals contains numerous large channel fills that range from sandy to heterolithic. Plant debris is the most common fossil material in the unnamed intervals. Although these intervals are predominantly nonmarine in character, some thin (<50 centimeters), bivalve- and ostracode-bearing gray shales occur locally. Also, the Avis Limestone is directly overlain by a several meters of gray to green shale that contains bivalves and brachiopods (the Upper Avis Shale of Reger, 1926). Because this shale is genetically part of the Avis marine zone, it has not been split out as a separate member in the present study.

DISCUSSION AND CONCLUSIONS

As noted by Cooper (1961, p. 68), "few formations have the amazing lithologic range of the Hinton succession which includes clean sands, red beds, black beds high in organic matter, channeling sands, even-bedded sandstones, and shell limestones. The succession is so variable vertically and horizontally that the naming of individual units, most of which are probably not very persistent, is of little value." Indeed, previous attempts at subdividing the formation into member-rank units have suffered from the naming and miscorrelation of units that are localized facies. As discussed above, the type Tallery Sandstone (Reger, 1926) occurs within the Fivemile Member of the present study. However, in the type area of the Fivemile Member, Reger (1926) miscorrelated a sandstone below the Fivemile Member with the type Tallery. Englund (1968) further confused the issue by assigning the name "Tallery" to a quartzose sandstone above the Fivemile Member.

Despite the complex lithologic variations in the Hinton Formation and the problems that have plagued attempts at subdividing the formation, the present study demonstrates that members can be objectively defined and traced in the upper Hinton Formation. The success of this study derives largely from use of the marine-zone concept that previously helped to resolve lithostratigraphic complexities in Pennsylvanian rocks of the central Appalachian basin (Blake and others, 1994; Martino, 1994, 1996). Historically, the Avis Limestone has been regarded as the only significant transgression in the Hinton Formation, largely because it is the only thick and mappable marine carbonate unit in a dominantly siliciclastic succession. As shown in this study, siliciclastic-dominated units also record regional marine events. Furthermore, as Miller and Eriksson (2000) recently suggested, the traditional view of the Mauch Chunk Group of southern West Virginia as a nonmarine redbed succession is inaccurate, because significant parts of the succession consist of shallow-marine to marginal-marine deposits. Indeed, transgressive-regressive cyclicity, similar to that in younger Pennsylvanian rocks in the area, appears to be a basic stratigraphic feature of the upper Hinton Formation. For example, the Middle Pennsylvanian Kanawha Formation consists of "coal-bearing sequences punctuated by numerous transgressive intervals" (Martino, 1996, p. 218). By analogy, the upper Hinton Formation consists of red bed-bearing sequences punctuated by two transgressive intervals.

Lithologic and faunal differences suggest that the Eads Mill Member records a more pronounced transgression than the Fivemile Member. The shale-dominated Eads Mill Member, which carries a diverse marine fauna, records offshore marine conditions throughout the area during peak transgression. Widespread shelly limestone beds within the member also suggest periods of sediment starvation across the basin (Miller and Eriksson, 2000; Huffman and Beuthin, 2001). In contrast, the Fivemile Member records marine conditions only in the southeastern study area; elsewhere, it carries a restricted marine fauna and contains abundant sandstone beds suggestive of nearshore deposition. Also, the Fivemile thins northward across the area, whereas the Eads Mill does not display such a pattern of thinning (Figure 3). Compared to the Eads Mill Member, the Fivemile Member may record a lesser change in relative sea level, a greater sediment supply during sea-level rise, or both.

Previous studies have suggested the existence of transgressive-regressive cycles in the Upper Mississippian clastic wedge that are similar in scale to Pennsylvanian cyclothems in the central Appalachian basin (Whisonant and Scolaro, 1980; Englund and others, 1981; Englund and others, 1986; Englund and Thomas, 1990). In the northern part of the basin, transgressive-regressive cycles consisting of intertonguing marine carbonate units and nonmarine siliciclastic units occur at a variety of scales (Brezinski, 1989; Kammer and Lake, 2001). A recent sequence stratigraphic model for Upper Mississippian cyclothems in the central Appalachian basin envisions the presence of six unconformity-bounded paralic sequences in the combined Avis Limestone-upper Hinton interval (Miller and Eriksson, 1999, 2000). Although this model suggests five transgressions in the upper Hinton Formation, the present study recognizes only two such transgressions. This discrepancy apparently relates to different approaches; the present study is based on detailed geologic mapping and sedimentological logging of numerous high-quality outcrop sections, whereas Miller and Eriksson relied primarily on the sequence stratigraphic interpretation of numerous gamma-ray logs combined with limited outcrop data. Also, the present study has integrated individual beds containing marine or brackish fauna into larger transgressive packages, in a manner similar to other recent studies of Carboniferous paralic cyclothems (Chesnut, 1994; Martino, 1994; Davies and Gibling, 2003). In contrast, Miller and Eriksson (2000, p. 221) apparently viewed individual marine-influenced beds, as well as "hot spikes" on gamma-ray logs, as separate transgressive events of regional significance.

The revised lithostratigraphic framework also provides some insights into the sequence stratigraphy of the upper Hinton Formation. Local truncation of the Fivemile Member (section 6 on Figure 3) suggests the existence of at least one paleovalley as envisioned by Miller and Eriksson (1999, 2000). On the other hand, neither the Tallery nor the Falls Mills sandstones of Reger (1926) appear to be incised valley-fills as inferred by Miller and Eriksson (1999, 2000); rather, these sandstones are facies within the Fivemile and Eads Mill members, respectively (Figures 5C and 6B).

Despite discrepancies between the present study and the work of Miller and Eriksson (1999, 2000), both studies agree that the upper Hinton Formation consists of allocyclically controlled genetic packages of strata. Although basinal tectonics involving either peripheral bulge migration or differential activation of basement blocks cannot be ruled out (Tankard, 1986; Ettensohn, 1994), the theory of glacio-eustatic control of Upper Paleozoic cyclothems (Wanless and Shepard, 1936; Veevers and Powell, 1987) provides a simple explanation for the Fivemile and Eads Mill transgressions. However, further study is needed to more definitely resolve the origin of these marine incursions. As noted by Miall (1997, p. 183), "the only reliable test of a eustatic control is precise correlation." Future work should be directed toward integrating outcrop and subsurface data, and making interbasinal comparisons and correlations.

To date, inconsistent application of poorly defined nomenclature for stratigraphic subdivisions of the Hinton Formation has produced more confusion than insight into the nature and origin of this succession of strata. However, the delineation of the Fivemile and Eads Mill members should provide a sound lithostratigraphic framework for future work. Both of these formally designated units are also genetic packages of strata (i.e., marine zones) that support a general but objectively based, cyclothemic architecture in the upper Hinton succession much like that found in Pennsylvanian-aged formations in the basin (Busch and Rollins, 1984; Blake and others, 1994; Chesnut, 1994; Martino, 1996). Future sedimentological and paleontological investigations can be integrated with this lithostratigraphic architecture to yield new insights, as well as evaluate existing hypotheses. Over time, the marine-zone concept may prove practical for refining lithostratigraphy in other parts of Upper Mississippian clastic wedge in the central Appalachian basin. In any case, resolving basic lithostratigraphic relationships should foster greater success in interpreting this succession as a record of environmental change related to onset of the Late Paleozoic ice age and the assembly of Pangea.

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Comatulid Crinoids from the Castle Hayne Limestone (Eocene), Southeastern North Carolina

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ABSTRACT

Seven comatulid crinoids, *Palaeantedon caroliniana*, *Microcrinus conoideus*, *Hertha plana*, *Himerometra bassleri*, *Amphorometra parva*, *Glenotremites carentonensis*, and *Placometra* n. sp., are identified from the Martin Marietta Quarry near Castle Hayne, New Hanover County, North Carolina. Identification of *Hertha plana*, *Amphorometra parva* and *Glenotremites carentonensis*, though possibly reworked from sediments below the Eocene Castle Hayne Limestone from North Carolina, extends the paleobiogeographic range of European species to southeastern North America. Extension of the paleobiogeographic ranges of these three species has implications for timing Tethyan influence upon distribution of comatulid taxa.

INTRODUCTION

The middle Eocene Castle Hayne Limestone in North Carolina is well known for its abundant and varied invertebrate fauna (Emmons, 1858; Canu & Bassler, 1920; Kellum, 1926; Richards, 1950; Jones, 1983; Carter *et al.*, 1988; Blow & Manning, 1996). While much work has been done on echinoids found within the Castle Hayne Limestone (Cooke, 1959; Kier, 1980), other echinoderm taxa such as ophiuroids, asteroids, and crinoids have been collected by several workers but not published upon. This may be due in part to the small size of the non-echinoid echinoderms and the difficulty of isolating, recognizing, and identifying these faunal elements.

With the exception of the description of *Microcrinus conoideus* by Emmons (1858), and a brief mention of *Microcrinus conoideus* and *Democrinus* sp. by Baum (1977), crinoids from the Castle Hayne Limestone have been virtually overlooked. Careful examination of prepared bryozoan-echinoid calcirudite from the Martin Marietta Quarry near Castle Hayne, New Hanover County, North Carolina has yielded numerous centrodorsals and brachials from several comatulid crinoid species.

In all, at least seven genera are represented. Examination of monographs by Gislén (1924, 1934) and Rasmussen (1961), as well as species descriptions from Rasmussen (1978), Strimple and Mapes (1984), Oyen (1995) and Jagt (1999), has led to identification of the genera *Amphorometra* Gislén, 1924; *Hertha* von Hagenow, 1840; *Himerometra* Clark, 1907; *Microcrinus* Emmons, 1858; *Palaeantedon* Gislén, 1924; *Glenotremites* Goldfuss, 1829, and *Placometra* Gislén, 1924. While Eocene comatulids have been reported throughout the

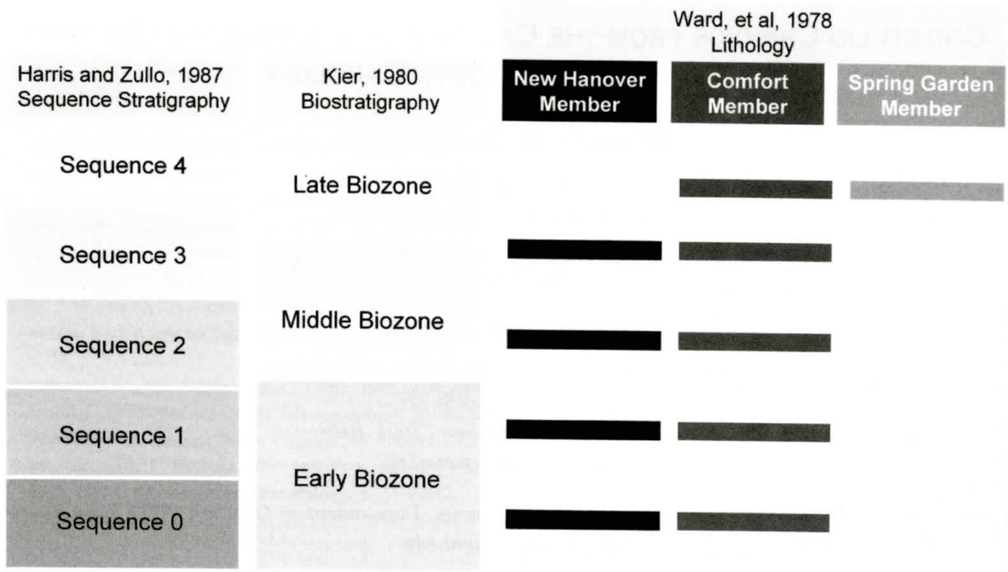

Figure 1. The correlation between sequence stratigraphy (Harris and Zullo, 1987), lithostratigraphy (Ward, *et al.*, 1978), and biostratigraphy (Kier, 1980) of the Castle Hayne Limestone. Ward, *et al.*'s, (1978) New Hanover, Comfort, and Spring Garden Members are time transgressive and not found in all depositional sequences. They essentially form a "fining upward" lithology in each depositional sequence in which they are found.

southeastern United States (Emmons, 1858; Gislén, 1924; Strimple and Mapes, 1984; Oyen 1995, 2002; and Oyen and Perrault, 1997), sediments from the Martin Marietta Quarry, near Castle Hayne, New Hanover County, North Carolina contain the greatest diversity reported to date.

GEOLOGIC SETTING

The middle Eocene Castle Hayne Limestone extends as a 16-32 km wide outcrop belt from Brunswick and New Hanover counties, north through east central Pender County, through western Onslow, Jones, and Craven counties, and into southeastern Pitt County (Otte, 1986, Figure 1) in southeastern North Carolina. Bounded by unconformities above and below, the formation is typically overlain by Oligocene and younger rocks, and underlain by Paleocene and Cretaceous rocks. The age of the Castle Hayne Limestone ranges from mid-Lutetian through Priabonian (Harris and Laws, 1997).

The stratigraphic relationship of the Castle Hayne Limestone has been interpreted by several workers, including Baum *et al.*, 1978; Ward *et al.*, 1978; Kier, 1980; Zullo and Harris, 1986, 1987; and Harris and Zullo, 1987. Harris and Zullo (1987), using sequence stratigraphy, divided the Castle Hayne Limestone into five sequences of deposition (0 – 4) (see Figure 1). Sequences bounded by regional unconformities reflect changes in sea-level and depositional environment. A complete lithologic section would consist of a phosphate-pebble biomicrudite base, overlain by biosparrudite which, in turn, is overlain by biomicrudite that grades into biosparrudite. The complete lithostratigraphic section is rarely represented at a single locality and sequences are typically represented by different lithologies at various localities where the Castle Hayne Limestone is exposed.

Ward *et al.* (1978: Figure 1) divided the formation into three members; New Hanover, Comfort, and Spring Garden. The lowest unit (New Hanover Member) is a slightly arenitic, micritic, phosphatic lithocalcirudite, middle unit (Comfort Member) is a gray to cream colored, bryozoan-echinoid calcirudite, that grades to a fine calcarenite, and the uppermost unit

(Spring Garden Member) is a tan to gray arenaceous, molluscan-mold biocalcirudite. All units typically represent different time sequences at different exposures.

Using primarily echinoids, Kier (1980) determined that the Castle Hayne Limestone could be divided temporally into three informal biozones (early, middle, and late). These biozones overlap somewhat sequences defined by Harris and Zullo (1987), yet they provide a reasonable biostratigraphic interpretation of the Castle Hayne Limestone.

Because each depositional sequence is composed of similar sets of lithofacies, Zullo and Harris (1987, Figure 2) performed a correlation using megafossils contained within the Castle Hayne Limestone. The echinoid genera *Protoscutella* and *Periarchus* were of particular use since they form an evolutionary sequence throughout the time range of the Castle Hayne Limestone (Kier, 1980). In particular, the oldest species, *Protoscutella mississippiensis*, is restricted to Sequence 1, *P. conradi* is restricted to Sequence 2, *P. plana* is restricted to the base of Sequence 3, and *Periarchus lyelli* (the youngest species) ranges throughout the remainder of Sequence 3 and through Sequence 4 (Zullo and Harris, 1987). These data support the use of echinoid biostratigraphy to temporally determine depositional sequence.

Comatulid specimens collected from the Martin Marietta Quarry in New Hanover County were found within spoil piles in the southwestern portion of the quarry where most recent excavations have taken place. The location of the spoil piles, coupled with the presence of copious specimens of *Periarchus lyelli*, indicates that the comatulids were probably originally contained within sequences 3 and/or 4 of the Castle Hayne Limestone (Harris, pers. comm.). Sequence 3 of the Castle Hayne Limestone is correlated with the Cross Formation in South Carolina and the Gosport Sand and lower Moodys Branch Formation in Alabama. Sequence 4 of the Castle Hayne Limestone correlates with the Harleyville and Parkers Ferry Members of the Cooper Formation in South Carolina, and the upper Moodys Branch Formation, Cocoa Sand, Pachuta Marl, Shubuta Clay, and Red Bluff Formation in Alabama (Zullo and Harris, 1987).

DEPOSITIONAL ENVIRONMENT

Interpretation of the depositional environment in which the Castle Hayne Limestone was formed is complicated by several factors, including the presence of formation outliers throughout the southern coastal plain, abruptly changing thickness of the formation from one exposure to another, and the isolation of outcrops (Otte, 1986). Stratigraphic analysis performed by Gibson (1970), Jones (1983), Otte (1986), Zullo and Harris (1986), and Harris and Laws (1997) has produced a fairly clear picture of the history and environment in which the carbonates were deposited. Sea level rise during the middle Eocene, coupled with a productive, relatively warm-water, environment, allowed for the development of Castle Hayne Limestone (Gibson, 1970; Otte, 1986; Harris and Laws, 1997). The depositional basin was formed by differential movement of fault-bounded crustal blocks, relative movement of which also controlled thickness and distribution of carbonate lithofacies of the Castle Hayne Limestone (Jones, 1983). The middle Eocene sea floor of North Carolina was composed of broad bands of inner shelf (0 - 15 m water depth), middle shelf (15 - 50 m water depth), and outer shelf (50 - 100 m water depth). Depositional environments formed an open, relatively warm-water embayment that paralleled the present-day shoreline (Jones, 1983; Otte, 1986) and outer shelf deposits most likely correspond to lithofacies in which the comatulid fauna was found.

MATERIALS AND METHODS

Cream colored, bryozoan-rich matrix was collected from abundant spoil piles in the Martin Marietta Quarry, Castle Hayne, New Hanover Co., North Carolina. All matrix was screened through ¼ inch mesh to remove large clasts and fossil fragments. Remaining matrix was then screened through 1/32 inch screen to remove silt and small fragments and screened material was then washed in tap water to re-

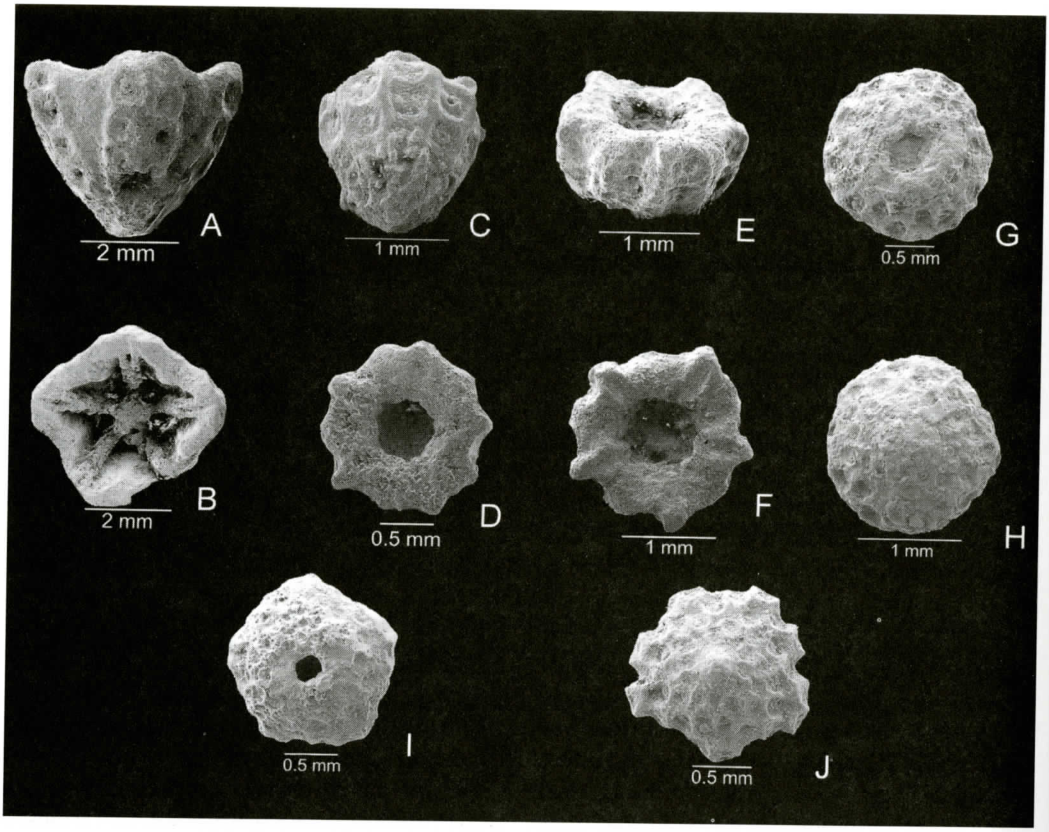

Figure 2. All images are of centrodorsals. A, B, *Microcrinus conoideus*, NCSM 8225, lateral and aboral views. C, D, *Amphorometra parva*, NCSM 8229, lateral and aboral views. E, F, *Placometra veniti* n. sp., holotype NCSM 8231, lateral and aboral views. G, *Himerometra bassleri*, NCSM 8228, dorsal view. H, *Palaeantedon caroliniana*, NCSM 8224, dorsal view. I, *Glenotremites carentonensis*, NCSM 8230, dorsal view. J, *Hertha plana*, NCSM 8226, dorsal view.

move remaining silt or clay. Once thoroughly dried, prepared matrix was examined under magnification in order to isolate and remove comatulid specimens. All figured specimens were mounted and sputter-coated with gold/paladium using a Anatech Hummer V sputter coater. Once coated, all figured specimens were examined and photographed using a Philips XL 30 ESEM TMP scanning electron microscope (NSF support award # DBI-0098534).

SYSTEMATIC PALEONTOLOGY

Repository: Figured specimens are reposited at the North Carolina State Museum of Natural Sciences (NCSM).

Order COMATULIDA Clark, 1908

Superfamily ANTEDONACEA Norman, 1865

Family ANTEDONIDAE Clark, 1909

Subfamily ANTEDONINAE Norman, 1865

Genus *Palaeantedon* Gislén, 1924

Palaeantedon caroliniana Gislén, 1934

(Figure 2, H)

Description. — This species is represented by a somewhat worn centrodorsal. Centrodorsal slightly arched, discoidal, dorsally rounded, with no dorsal scar or depression. No cirrus free area. Cirrus sockets small, numerous, closely spaced, without distinct ornament and do not form distinct columns. Centrodorsal cavity moderate. Ventral surface with indistinct basal furrows.

Discussion. — This is the first recorded occurrence of this species from North Carolina. This species has also been recorded from Eocene formations in South Carolina and Louisiana.

Figured specimen. — NCSM 8224.

Occurrence. — Castle Hayne Limestone, Eocene, Martin Marietta Quarry, near Castle Hayne, New Hanover County, NC.

Subfamily ZENOMETRINAE Clark, 1909

Genus *Microcrinus* Emmons, 1858

Microcrinus conoideus Emmons, 1858

(Figure 2, A-B)

Description. — This species is represented by a somewhat worn centrodorsal; conical, broken dorsally on one radial face such that four most dorsal cirrus sockets on that face are missing. Cirrus sockets arranged into distinct columns of two sockets separated by well defined interradial furrows. Centrodorsal cavity star-shaped, deep and wide, with interradial septa.

Discussion. — *Microcrinus conoideus* was first described from the Eocene of Craven County, NC by Emmons (1858). The type specimen appears to be missing. Baum (1977) noted the occurrence of this species in his unpublished thesis, but did not provide a figure.

Figured specimen. — NCSM 8225.

Occurrence. — Castle Hayne Limestone, Eocene, Martin Marietta Quarry, near Castle Hayne, New Hanover County, NC.

Subfamily HELIOMETRINAE Clark, 1909

Genus *Hertha* von Hagenow, 1840

Hertha plana (Nielsen, 1913)

(Figure 2, J)

Description. — This species is represented by a well preserved centrodorsal. Centrodorsal slightly arched, somewhat discoidal, with a sharp irregular ventral edge. Dorsal area rounded to knob-like, cirrus free, without dorsal star. Cirrus sockets small, without ornament, numerous without forming distinct columns. Centrodorsal cavity somewhat narrow. Ventral surface without basal furrows.

Discussion. — This is the first recorded occurrence of this species from North Carolina and also the first recorded occurrence of this genus in North America. This species is also known from the Danian (Lower Tertiary) of Denmark and the Netherlands.

Figured specimen. — NCSM 8226.

Occurrence. — Specimen is possibly reworked from the Rocky Point Member of the Peedee Formation, which uncomformably lies below the Castle Hayne Limestone, Eocene, Martin Marietta Quarry, near Castle Hayne, New Hanover County, NC.

Superfamily MARIAMETRACEA Clark, 1909

Family HIMEROMETRIDAE Clark, 1908

Genus *Himerometra* Clark, 1907

Himerometra bassleri Gislén, 1934

(Figure 2, G)

Description. — This species is represented by a fairly well preserved centrodorsal. Centrodorsal low discoidal with depressed dorsal area. Cirrus sockets without distinct ornament, numerous, and arranged into 3 irregular marginal circles. Ventral surface with basal furrows.

Discussion. — This is the first recorded occurrence of this species from North Carolina. This species is also known from the Eocene Moodys Branch Formation, Mississippi, Eocene Yazoo Formation, Alabama, Eocene lower Ocala Limestone, Florida, and the Eocene Dry Branch Formation, South Carolina.

Figured specimen. — NCSM 8228.

Occurrence. — Castle Hayne Limestone, Eocene, Martin Marietta Quarry, near Castle Hayne, New Hanover County, NC.

Superfamily TROPIOMETRACEA Clark, 1908

Family CONOMETRIDAE Gislén, 1924

Genus *Amphorometra* Gislén, 1924

Amphorometra parva Gislén, 1925

(Figure 2, C-D)

Description. — This species is represented by a well preserved centrodorsal. Centrodorsal high conical. No dorsal star or depression. Cirrus sockets arranged in columns. Centrodorsal cavity moderate. Ventral side shows worn basal furrows.

Discussion. — Baum (1977) recorded this genus from North Carolina in his unpublished thesis, but did not provide a figure. This is the first figured specimen of this genus from North Carolina. This species is also known from the Upper Cretaceous of England.

Figured specimen. — NCSM 8229.

Occurrence. — Specimen is possibly reworked from the Rocky Point Member of the Peedee Formation, which uncomformably lies below the Castle Hayne Limestone, Eocene, Martin Marietta Quarry, near Castle Hayne, New Hanover County, NC.

Family PTERICOMIDAE Rasmussen, 1978

Genus *Placometra* Gislén, 1924

Placometra veniti, new species

(Figure 2, E-F)

Diagnosis. — Centrodorsal truncated conical, with large, flattened dorsal side. No radial dorsal star. Centrodorsal cavity with overhanging edge. Sockets sub-circular, very large, without distinct ornament, approximately three on each radial side. Cirrus sockets separated by pronounced interradial ridges.

Description. — Dimensions: Centrodorsal height: 0.7 mm. Centrodorsal outside diameter: 2.0 mm. Centrodorsal inside diameter: 0.9 mm.

Centrodorsal discoidal, sub-pentagonal, rough and dorsally flattened. Large, deep cirrus sockets, without sculpture, arranged in 10 rows, approximately 3 sockets in each row. Cirrus sockets without well defined sculpture. Nerve lumen for cirrus sockets sub-rectangular in outline and well defined.

Radial sectors of ventral face form shallow clefts. Centrodorsal cavity with pronounced overhanging ledge. Basal furrows narrow and deep. Sides of centrodorsal divided into 5 radial areas by ridges that extend from basal furrows on ventral surface to the flattened dorsal area. Two cirrus socket rows between adjacent radial ridges. Area between each pair of cirrus sockets forms a slightly raised ridge that extends from outer edge of the ventral face to dorsal area. Dorsal surface flat. Dorsal star not present.

Discussion. — *Placometra veniti* differs from the Turonian species *P. laticirra* (Carpenter, 1880) in having more regularly spaced and shaped cirrus sockets. *Placometra veniti* differs from the Turonian species *P. scutata* (Gislén, 1924) in size, number, and shape of cirrus sockets per radial side. The centrodorsal of the Late Cretaceous species *P. mortenseni* (Gislén, 1924) is sub-pentagonal, whereas the centrodorsal of *P. veniti* is conical.

Holotype. — NCSM 8231.

Etymology. — This species is named in honor of Edward Venit who first introduced us to the crinoid fauna of Castle Hayne.

Occurrence. — Possibly re-worked from sediments below the Castle Hayne Limestone, Eocene, Martin Marietta Quarry, near Castle Hayne, New Hanover County, NC.

Superfamily NOTOCRINACEA Mortensen, 1918

Family NOTOCRINIDAE Mortensen, 1918

Genus *Glenotremites* Goldfuss, 1829

Glenotremites carentonensis de Loriol, 1894

(Figure 2, I)

Description. — This species is represented by a somewhat worn, bryozoan-encrusted centrodorsal. Centrodorsal low pentagonal. Dorsal area worn to a circular pit. Dorsal area flattened and somewhat cirrus free. Cirrus sockets worn, small, and irregularly arranged. Centrodorsal cavity narrow.

Discussion. — This is the first recorded occurrence of this species from North Carolina. This taxon is also known from the Upper Cretaceous of France.

Figured specimen. — NCSM 8230.

Occurrence. — Specimen is possibly reworked from the Rocky Point Member of the Peedee Formation, which uncomformably lies below the Castle Hayne Limestone, Eocene, Martin Marietta Quarry, near Castle Hayne, New Hanover County, NC.

DISCUSSION

The diverse comatulid crinoid fauna, along with other abundant and varied invertebrate faunas, from the Castle Hayne Limestone of New Hanover County, North Carolina suggests a productive carbonate ramp environment of deposition. The abundance and diversity of these invertebrate faunas is not unexpected given the nature of Eocene Tethyan paleoenvironments. What is surprising is the paleogeographic distribution of some comatulid species identified from sediments from Castle Hayne. Identification of *Glenotremites carentonensis*, *Hertha plana*, and *Amphorometra parva*, previously primarily known as European, from North Carolina effectively extends the paleogeographic range of these species to include North America.

Furthermore, the presence of three comatulid species with European, Late Cretaceous affinities raises an interesting spatial issue. Three possibilities exists: (1) these species are reworked from Late Cretaceous sediments that unconformably underlie the Castle Hayne Limestone, (2) the taxa are part of the Castle Hayne fauna and have not been reworked from Cretaceous sediments, and (3) the taxa represent both in situ and reworked specimens. If any of the taxa represent specimens reworked from Late Cretaceous sediments then the Tethyan influence was felt much earlier than previously thought. More work needs to be performed in order to determine if any of the three taxa exist within the Late Cretaceous sediments found at the Castle Hayne Locality.

Oyen (1995) postulated that European comatulid crinoids show a wide geographic distribution because of oceanic currents operating within the Tethyan Seaway. Oyen (1995) based his argument of paleo-circulation patterns on *Himerometra* and called for further investigation of North American comatulid crinoids.

With three of the species found in North Carolina having European affinities, Oyen's Tethyan distribution pattern is supported and suggests further need for investigation of southeastern crinoid faunas.

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THE STREAM NET AS AN INDICATOR OF CRYPTIC SYSTEMATIC FRACTURING IN LOUISIANA

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ERRATA

1. At the end of the fourth line of the second paragraph of the **Abstract**, on p. 1, "|" should read, "≈."

2. In the caption of Figure 3 on p. 6, fifth line from the bottom, "Φ^2" should read, "χ^2."