Stratigraphy, petrography, and correlation of late Cenozoic middle Atlantic Coastal Plain deposits: Implications for late-stage passive-margin geologic evolution

FRANK J. PAZZAGLIA

Department of Geosciences, Pennsylvania State University, University Park, Pennsylvania 16802

ABSTRACT

Regional chronostratigraphic correlation of middle Atlantic upper Coastal Plain and Fall Zone fluvial deposits, long thought to represent the updip, proximal facies of a well-known post-Oligocene marine sequence in the Salisbury Embayment (Ward and Strickland, 1985; Figs. 1, 2 and 3), have challenged geologists for more than a century (Lewis, 1880; McGee, 1888; Bascom and others, 1902, 1909; Shattuck, 1901, 1906; Shattuck and others, 1902; Salisbury and Knapp, 1917; Barrell, 1920; Bascom, 1921; Knopf, 1924; Cooke, 1930, 1958; Campbell, 1931; Campbell and Bascom, 1933; MacClintock and Richards, 1936; Darton, 1939; Hack, 1955; Schlee, 1957; Owens and Denny, 1979a; Owens and Minard, 1979; Ramsey, 1992). Successful regional chronostratigraphic correlation of these deposits will provide insight on the tectonic (Owens, 1970; Hack, 1982; reviewed in Gardner, 1989; Prowell, 1988), isostatic (Pazzaglia and Gardner, in press [a]), eustatic (Hon and others, 1987), and fluvial processes driving late Cenozoic middle Atlantic passive-margin geologic evolution. Fluvial deposits at the head of Chesapeake Bay (Fig. 1) occur as a crucial link between previously described, late Cenozoic stratigraphy in New Jersey (Owens and Sohl, 1969; Owens and Minard, 1975, 1979; Owens and others, 1989; Sugarman and others, 1991, 1993), on the Delmarva Peninsula (Jordan, 1964; Owens and Denny, 1979a; Owens and Minard, 1979; Benson and others, 1985; Mixon, 1985; Andres, 1986; Groot and others, 1990; Ramsey and Schenck, 1990), in southern Maryland and Virginia (Owens and Denny, 1978, 1979b; Ward and Blackwelder, 1980; Blackwelder, 1981; Cronin and others, 1984; Ward and Strickland, 1985; Ramsey, 1988; Mixon and others, 1989a, 1989b; Ward and Powars, 1989, 1991; McCartan, 1989a, 1989b), and central Appalachian Piedmont (lower Susquehanna River; Pazzaglia and Gardner, 1992 and in press [b]).

INTRODUCTION

The origin, age, and correlation of middle Atlantic upper Coastal Plain and Fall Zone (Fig. 1) fluvial deposits, long thought to represent proximal facies of a well-known post-Oligocene marine sequence in the Salisbury Embayment synthesized from petrographic-lithostratigraphic deposits and from a review and compilation of previous research; (2) present a generalized reconstruction of post-Oligocene Salisbury Embayment depositional history; and (3) speculate on the influence of late Cenozoic isostasy, tectonics, and eustasy on the stratigraphic framework of these passive-margin deposits. There exists a plethora of geologic processes, including subaerial weathering, diagenesis, recycling of older sediments, and hydraulic sorting during transport; these limit the usefulness of petrographic techniques for the middle Atlantic Coastal Plain (Krajewski, 1977; Soller and Owens, 1991). The purpose of the petrographic analyses therefore is not to present and draw conclusions from a statistically valid data set, but to systematically compile a semi-quantitative data set to observe general petrographic similarities, differences, and trends.

Selected deposits include (1) upper Coastal Plain and Fall Zone fluvial deposits at the head of Chesapeake Bay; (2) fluvial, marine, and estuarine deposits of the Delmarva Peninsula; and to a lesser extent, (3) fluvial and marine deposits in New Jersey, Maryland, and Virginia (Fig. 1). Petrographic analyses were conducted on 56 sand samples collected at localities named in Figure 4 and elsewhere. The bulk samples were washed and sieved with a 230-mesh screen to remove clay and fine silt. The washed samples were then separated by size, retaining a split of the 250 μm size fraction for light-mineral analysis and splits of the 149 and 88 μm fractions for heavy-mineral analysis. Alkaline and potassium feldspars were identified with the aid of Rodzonic Acid and Sodium Cobaltinitrate stain, respectively. The heavy minerals were separated from the 149 and 88 μm size frac-
grains, thus minimizing sorting effects (Folk, 1968).

REGIONAL GEOLOGIC SETTING

Salisbury Embayment Marine Stratigraphy

A thick sequence of Cretaceous to Quaternary, gently southeast-dipping, predominantly marine unconsolidated sediments underlies the middle Atlantic Coastal Plain in a tectonically active, arcuate-shaped basin called the "Salisbury Embayment" (Owens, 1970; Brown and others, 1972; Mixon and Newell, 1982; Newell and Rader, 1982; Mixon and Powars, 1984; Newell, 1985; Owens and Gohn, 1985; Ward and Strickland, 1985; McCartan, 1989c; Ward and Powars, 1989, 1991; Figs. 1 and 3). Late Oligocene to late Pliocene, unconformity-bound formations with well-established biostratigraphic ages (Ward and Strickland, 1985; Ward and Powars, 1991) of the Chesapeake Group (Old Church, Calvert, Choptank, St. Marys, Eastover, Yorktown, and Chowan River Formations) overlie a regional Oligocene unconformity (Posag, 1985; Figs. 2 and 3). Below this unconformity, Cretaceous through Eocene, predominantly fluvial deltaic and marine deposits dip basinward more steeply than do the Chesapeake Group. The Kirkwood and Cohanseym Formations of New Jersey (Owens and others, 1989; Sugarman and others, 1991, 1993) and the informal lithostratigraphic Bethany and Manokin formations of Delmarva (Andres, 1986) are stratigraphic equivalents to the Calvert, Choptank, and Eastover Formations, respectively (Fig. 2), but not considered part of the Chesapeake Group.

Chesapeake Group deposition reflects complex interactions between basin tectonics and/or isostasy and eustasy. Detailed dating demonstrates that marine deposition began in New Jersey in the early Miocene (Burdigalian) with the Kirkwood Formation (Owens and others, 1989; Sugarman and others, 1993) and steadily progressed southwestward into Virginia, ceasing with the Chowan River Formation during the late Pliocene (Ward, 1984). This southwest shift may reflect uplift of the northern Appalachians and regional down-to-the-southwest tectonic tilting (Newell and Rader, 1982) of the Salisbury Embayment.

Upper Coastal Plain and Fall Zone Fluvial Stratigraphy

Upper Coastal Plain and Fall Zone fluvial deposits, characterized by thin (<30-m),

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Figure 1. Location of study area. Har. = Harrisburg; NYC = New York City; Phil. = Philadelphia; Bal. = Baltimore; Wash. D.C. = Washington, D.C.; HR = Hudson River; DR = Delaware River; SKR = Schuylkill River; SR = Susquehanna River; PR = Potomac River; RR = Rappahannock River; JR = James River. Cross sections are illustrated in Figure 2.

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poorly dated sequences of sand, gravel, and silty clay, occur as inset fluvial terraces and regional alluvial sheets that unconformably overlie Piedmont bedrock or upper Coastal Plain marine deposits. In New Jersey, these fluviol deposits are represented by the quartzose, late Miocene? Beacon Hill Formation that is 60–110 m (deposit base) above mean sea level (amsl) followed by the quartzo-feldspathic late Miocene? Bridgeton Formation (25–45 m amsl), the feldspathic late Miocene? Pensauken Formation (15–30 m amsl), and the lithic-rich Trenton Gravels (Van Sciver Lake beds and Spring Lake beds of Owens and Minard, 1979) below 12 m amsl.
Figure 3. Generalized stratigraphy of the Salisbury Embayment. The three cross sections are constructed to illustrate which down-dip marine units are potential chronostratigraphic equivalents to fluvial deposits at the head of Chesapeake Bay. Notice how post-Oligocene sediments, including the fluvial deposits above the Oligocene unconformity, dip basinward less steeply than older units. Cross section A′A′ compiled from Brown and others (1972), Cleave and others (1968), and Mixon and others (1989). Cross section B′B′ compiled from Owens and Denny (1979a), Owens and Minard (1979), Mixon (1985), Benson and others (1985, 1990), and Ramsey and Schenck (1990). Cross section C′C′ compiled from Kummel (1950), Owens and Minard (1975, 1979), Owens and others (1989). See Figure 1 for cross-section locations.

Figure 4. Generalized geologic map of a portion of Cecil and Harford Counties, Maryland (modified from Owens, 1969, and Conant, 1990). Fault location and relative offset is inferred from elevation of Pensauken Formation and Turkey Point beds. See Figure 1 inset for location.
Older fluvial deposits such as the Bryn Mawr Formation often form a resistant caprock that locally preserves underlying Coastal Plain sediments resulting in an inversion of topography and the geomorphic expression of the fluvial unit as an upland gravel.

RESULTS

Bryn Mawr Formation

The Bryn Mawr Formation (Lewis, 1880; Bascom and others, 1909; Bascom, 1924), a quartzose, sandy gravel, occurs along the Pennsylvania, Delaware, and northeastern Maryland Fall Zone and upper Coastal Plain at elevations greater than 60 m asl. Basinward, across the Elk Neck Peninsula, the base of the deposit drops in elevation from about 120–60 m asl, truncating first the Potomac Group and then the Matawan and Monmouth Group (Conant, 1990; Figs. 2, 4, and 5). The formation has a highly variable thickness, ranging from about 2 m in interfluvie regions to 30 m adjacent to the Susquehanna River.

Quarrying operations in Cecil County, Maryland (York Pit, Fig. 4) expose the Bryn Mawr Formation as 10–30 m of well-rounded, well-sorted, open-framework, sandy gravel and pebbly sand interbedded with sand, silty sand, and clayey silt beds unconformably overlying the Cretaceous Potomac Group (Figs. 3, 6). Exposures extending for 1–2 km were observed over a period of 2.5 yr and 200–500 m of highwall retreat. The entire deposit is deeply weathered with high-chroma, dark red and red colors (2.5 YR) in the upper 1 to 5 m, reflecting the accumulation of illuviated clay in thick argillic horizons, and ironstone (pedogenic, lateritic?) occurring at the deposit base. Paleoflows obtained from trough cross-stratification throughout Cecil County, Maryland, exhibit a south and southeast orientation, and all of the channels that have been observed in the York pit have a northwest to southeast orientation (Fig. 4).

Three informally defined members contain six distinct lithofacies. Member 1 contains lithofacies 1, a 1- to 5-m-thick, moderately well sorted, planar and trough cross-stratified, quartz pebbly sand exhibiting a deeply weathered, high-chroma red and dark red (10R), truncated argillic horizon. Member 1 occurs only at the base of the formation where it unconformably overlies the Potomac Group and is truncated and unconformably overlain by Members 2 and 3.

Member 2, composed of lithofacies 2, 3, and 4 (Fig. 6), represents the thickest and most extensive part of the Bryn Mawr Formation, underlying the topographically highest interfluvies on the Fall Zone and upper Coastal Plain. Lithofacies 2 is a well-sorted, open-framework, white and yellow gravel (average clast size of 4–10 cm) and pebbly sand with 1- to 5-m-thick topset-foreset cross-sets that occurs at or near the base of member 2. Lithofacies 3 is a very well-sorted, white and yellow, medium-bedded, wavy and trough cross-stratified medium sand interbedded with finely laminated, wavy cross-stratified, gray, silty clay; lithofacies 4 is a thin to thickly bedded, extremely well sorted, red, white, or yellow, pebbly sand with trough cross-stratification and imbricated pebbles on horizontal bedding surfaces. All unconformably overlie lithofacies 2.

Member 3, unconformably inset into members 1 and 2, contains two lithofacies (lithofacies 5 and 6; Fig. 6). Lithofacies 5 is a red, thin to thickly bedded, poorly to moderately well-sorted sand and pebbly sand interbedded with thick, gray, laminated and massive, silty clay and clayey silt exhibiting clayey rip-up clasts. The dominant sedimentary structures in lithofacies 5 occur as planar and trough cross-stratification in 10- to 50-cm-thick sets. In contrast, lithofacies 6, a thick to very thickly bedded, medium-sorted, trough cross-stratified, red and yellow, pebbly sand and sandy gravel exhibits gently east-dipping 0.5- to 5-m-thick epsilon cross-stratification in large, northwest- to southeast-oriented channels at least 0.25 to 0.5 km wide. Individual epsilon cross-beds fine-up.

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Figure 5. Schematic cross section of upper Coastal Plain fluvial stratigraphy in Cecil and Harford Counties, Maryland. Note the break in vertical scale between 45 and 90 m.
from coarse and medium pebbly sand to medium and fine, trough cross-stratified sand, which is then abruptly conformably overlain by white and gray, finely laminated, clayey silt 0.5–2 m thick. In places, the lamination is destroyed by what may be poorly preserved bioturbation.

These lithofacies suggest deposition in a braided alluvial plain of the lower Susquehanna River. Although poorly preserved, member 1 probably represents transverse gravel bars and dunes in shallow braided alluvial channels in the more proximal portions of the braid plain (Collinson, 1986; Wescott and Ethington, 1990). In contrast, member 2 represents prograding distributary mouth bars (lithofacies 2) and braided-stream and shoreline deposits (lithofacies 4; Wescott and Ethington, 1990; Elliot, 1986) similar to those described for the more distal portions of braid plains or fan-deltas (Wescott and Ethington, 1990), as well as estuarine, crevasse splay, and levee deposits (lithofacies 3) typical of interdistributary regions (Elliot, 1974; Miall, 1984). Member 3 represents relatively deep braided channels with lateral accretion surfaces (lithofacies 6) (Allen, 1963; Collinson, 1986; Wescott and Ethington, 1990) juxtaposed with abandoned channel fills of fine-grained overbank and lacustrine sediments, crevasse splay, and levee deposits (lithofacies 5) (Walker and Cant, 1984). During Bryn Mawr Formation deposition, the braided plain probably extended across most of the northern Chesapeake region and across what is now the northern Delmarva Peninsula (Fig. 1).

The age of the Bryn Mawr Formation is poorly known; however, a recent, fortuitous palynoflora recovered from the Brandywine Formation, a similar quartzose upland gravel of the Potomac River (Hack, 1955; Schlee, 1957) suggests a late Miocene (McCcartan and others, 1990) to early Pliocene (Groot and others, 1990) age and warm-temperate climate. Extensive efforts to extract similar palynofloras from fine-grained beds of the Bryn Mawr Formation have been largely unsuccessful. A sample collected from Woodlawn, Maryland (Fig. 4) contains the sparse and very poorly preserved remains of a post-Cretaceous palynoflora within a matrix of dark, carbonized woody fragments (F. Pazzaglia, unpub. data) consistent with those found in the Chesapeake Group (Groot and others, 1990; Groot, 1991). Additionally, a single leaf impression recovered from lithofacies 5 (York Pit) and a petrified log cemented by hematite, found in the Petrillo Pit in the center of the Elk Neck Peninsula (Fig. 4), have been tentatively identified as *Pterocarya* and *Taxodium*, respectively (A. Traverse, 1992, personal commun.). Both strongly resemble the well-preserved and abundant *Pterocarya* and *Taxodium* fossils recovered from the Brandywine Formation (McCcartan and others, 1990).

**Petrography.** The pebble-size fraction of the Bryn Mawr Formation is nearly 98% vein quartz and quartzite with ~1% white chert and <1% Appalachian sandstone and Piedmont crystalline lithologies (Table 1). Labile clasts are completely weathered and so friable that they crumble when handled. Petrography of the sand-size fraction light-mineral suite is dominated by quartz with only a minor amount (<5%) of weathered lithic fragments and kaolinite “ghosts” after feldspar.
TABLE 1. CLAST COUNTS OF SELECTED UPPER COASTAL PLAIN, FALL ZONE, AND PIEDMONT FLUVIAL DEPOSITS

<table>
<thead>
<tr>
<th>Age</th>
<th>Name</th>
<th>Vein quartz</th>
<th>Meta quartz</th>
<th>Quartzite</th>
<th>Sandstone siliciclast short limestone</th>
<th>Limestone</th>
<th>Ironstone granite grains</th>
</tr>
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<tbody>
<tr>
<td>Q</td>
<td>QT4-Suegannah R.</td>
<td>3.5</td>
<td>--</td>
<td>10</td>
<td>74</td>
<td>3.5</td>
<td>--</td>
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<tr>
<td></td>
<td>QT4-Suegannah R. (Peltier, 1949)</td>
<td>2</td>
<td>12</td>
<td>31</td>
<td>81</td>
<td>5</td>
<td>5</td>
</tr>
<tr>
<td>Q</td>
<td>QT2-Suegannah R.</td>
<td>1</td>
<td>2</td>
<td>31</td>
<td>64</td>
<td>2</td>
<td>2</td>
</tr>
<tr>
<td>and</td>
<td>QT2-Suegannah R. (Peltier, 1949)</td>
<td>5</td>
<td>17</td>
<td>17</td>
<td>75.5</td>
<td>2.5</td>
<td>2</td>
</tr>
<tr>
<td>Q17-Conod Farm</td>
<td>15</td>
<td>40.5</td>
<td>30.5</td>
<td>12</td>
<td>2</td>
<td></td>
<td>2</td>
</tr>
<tr>
<td>P</td>
<td>Pensauken/Columbia-Turkey Point</td>
<td>18.5</td>
<td>35.0</td>
<td>16</td>
<td>22.5</td>
<td>3.5</td>
<td>3</td>
</tr>
<tr>
<td></td>
<td>Pensauken/Columbia-Turkey Point</td>
<td>7</td>
<td>30</td>
<td>10</td>
<td>54</td>
<td>9</td>
<td>9</td>
</tr>
<tr>
<td></td>
<td>Columbia (average)-Delmarva</td>
<td>46</td>
<td>--</td>
<td>16</td>
<td>12</td>
<td></td>
<td>2</td>
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<tr>
<td>Q17-Braid Creek</td>
<td>16</td>
<td>16</td>
<td>41</td>
<td>24</td>
<td>3</td>
<td></td>
<td>3</td>
</tr>
<tr>
<td>Q17-Suegannah River (combined)</td>
<td>25</td>
<td>12</td>
<td>37.5</td>
<td>25</td>
<td>&lt;1</td>
<td></td>
<td>1</td>
</tr>
<tr>
<td>P</td>
<td>Perryville-Mountain Hill</td>
<td>12</td>
<td>45</td>
<td>41</td>
<td>1</td>
<td>1</td>
<td>1</td>
</tr>
<tr>
<td></td>
<td>Perryville-Elk Neck</td>
<td>25</td>
<td>62</td>
<td>0</td>
<td>5</td>
<td>8</td>
<td>8</td>
</tr>
<tr>
<td></td>
<td>Perryville-Rikon</td>
<td>28</td>
<td>59</td>
<td>0</td>
<td>4</td>
<td>9</td>
<td>9</td>
</tr>
<tr>
<td></td>
<td>Perryville-Rt. 40 Pit</td>
<td>10</td>
<td>15</td>
<td>31.5</td>
<td>31.5</td>
<td>12</td>
<td>12</td>
</tr>
<tr>
<td>M</td>
<td>Bloom Castle-Virginia (Ramsey, 1988)</td>
<td>64</td>
<td>--</td>
<td>24</td>
<td>9</td>
<td>3</td>
<td>3</td>
</tr>
<tr>
<td></td>
<td>Bridgeton-Morrisville, PA</td>
<td>5</td>
<td>8.5</td>
<td>18</td>
<td>66</td>
<td>2.5</td>
<td>2.5</td>
</tr>
<tr>
<td></td>
<td>Bridgeton-Jamesburg, New Jersey</td>
<td>20</td>
<td>49</td>
<td>19</td>
<td>7</td>
<td>5</td>
<td>5</td>
</tr>
</tbody>
</table>

T2g-Britton Farm, Lancaster County, PA
T2g-Kirk Farm, Lancaster County, PA
Bryn Mawr-Elk, PA (2.4 cm)
Bryn Mawr-Elk, PA (4.10 cm)
Bryn Mawr-Elk Neck
Bryn Mawr-Mountain, Harford County, Maryland, (Owens, 1969)
Bryn Mawr-Carver, Harford County, Maryland, (Owens, 1969)
Bryn Mawr-Aberdeen, Harford County, Maryland, (Owens, 1969)
Bryn Mawr-Webster, Harford County, Maryland, (Owens, 1969)
Bryn Mawr-Rays Hill, Cecil County, Maryland, (Owens, 1969)
Beacon Hill-Beacon Hill, New Jersey
Brandwine-#07, Herring Run, D.C. (Schlee, 1967)
Brandwine-#10, northern peninsula (Schlee, 1967)
Brandwine-#05, central peninsula (Schlee, 1967)
Brandwine-#07, southern peninsula (Schlee, 1967)
Brandwine-average-Maryland, (Schlee, 1967)

K
Potomac Group-Cecil County, Maryland

Note: clast size = 2-4 cm in diameter unless otherwise indicated.

*Q = Quaternary; P = Pliocene; M = Miocene.

See Figures 1 and 4 for locations.

- denotes that the relative percent of metamorphic quartz was not reported. Total non-sedimentary quartz percentage is reported in vein quartz column.

(7). Quartz grains show relatively high proportions of the monocrystalline unstrained variety with respect to the strained and polycrystalline variety. The non-opaque heavy mineral assemblage is a very mature suite of zircon, tourmaline, rutile, and leucoxene (Fig. 7), typically with less than 20% metamorphic component, predominantly staurolite and sillimanite, and less than 1% labile component. Leucoclime and ilmenite/leucoxene and less than 5% magnetite (Fig. 7) comprise the opaque heavy mineral suite. X-ray diffraction identifies the major clay minerals as kaolinite and kaolinite polymorphs such as dickite. Conant (1990) reported the clay minerology as 65% kaolinite, 5%-10% illite, 20%-25% gibbsite, and smaller amounts of mixed-layer clay and montmorillonite.

Perryville Formation

Relatively flat geomorphic surfaces adjacent to streams, at the mouth of the Susquehanna River, and fringing the shoreline of Cecil and Harford Counties, Maryland, 30 to 45 m asml are underlain by heterolithic fluvial sands and gravels (Miller, 1920; Conant, 1990; Figs. 4 and 5). Herein these deposits comprise an informal lithostratigraphic unit called the Perryville formation. The Perryville formation is a 1- to 6-m-thick, heterolithic, but quartz dominated, well-rounded, moderately well to poorly sorted, pebbly sand. The deposit occasionally exhibits horizontal to slightly inclined bedding surfaces, fining-up bedding and shallow trough cross-stratification in 10- to 20-cm-high sets. Paleoflow directions vary widely across Cecil and Harford Counties but generally indicate a transport direction subparallel to the current drainage orientation (Fig. 4).

Perryville formation deposition occurred as thin stratified and interbedded, at streams (Conant, 1990; Owens, 1969), pediment gravels in interfinger regions, and alluvial-fan deposits at stream mouths. The petrography strongly reflects a local source with the more lithic and labile-rich deposits invariably located near Fall Zone streams and the Susquehanna River. The abundance of reworked, rounded ironstones in both the pebble- and sand-size fractions of the Perryville formation strongly implicates the Bryn Mawr formation as a major source.

Dateable material has not been recovered from the Perryville formation, and so its age is largely conjectural. The Perryville gravels are stratigraphically younger than the Bryn Mawr Formation and older than the Pensauken, Talbot, and Kent Island formations, which are inset at lower elevations (Figs. 4 and 5).

Petrography. The petrography of the Perryville formation varies with location. Point counts of the coarse pebble-size fraction (Table 1) show predominantly quartz and quartzite, but up to 40% labile clasts, including ironstone clasts, Piedmont quartzite schists, gneisses, and gabbros, and Valley and Ridge quartzites, sandstones, and red siltstones. The labile clasts become even more easily when struck with a hammer and exhibit weathering rinds ranging from 1 to 3 cm; however, they are not nearly as weathered as the Bryn Mawr Formation labiles. The petrography of the sand-size light minerals shows primarily quartz with as much as 15% feldspar and lithic fragments dominated by reworked ironstone (Fig. 7). Zircon, tourmaline, and rutile with ~20% to 30% sillimanite, kyanite, and staurolite, and up to 10% labile minerals, such as hornblende, garnet, and epidote, comprise the non-opaque heavy-mineral suite (Fig. 7). The opaque heavy-mineral suite exhibits relatively high amounts of leucoxene, ilmenite/leucoxene and hematite with respect to ililit and typically less than 10% magnetite (Fig. 7). Both the heavy- and light-mineral suites of the Perryville formation are similar to those of the Bryn Mawr Formation, differing mainly in the presence of more labile components in the former.

Pensauken Formation

The Pensauken Formation (Owens and Denny, 1979a; Owens and Minard, 1979; Columbia Formation of McGee, 1888; Jordan, 1964), a 5- to 30-m-thick, poorly sorted, feldspathic, boulder- and lithic-rich, fluvial channel-sand deposit, mantles most of the northern Delmarva Peninsula (Conant, 1990; Ramsey and Schenck, 1990). The top of the formation varies from ~18 to 30 m in elevation, defining a distinct, widespread geomorphic surface; locally, in drill cores. The base
Figure 7. Petrography of selected (a) upper Coastal Plain fluvial deposits and (b) lower Coastal Plain marine deposits. The samples are arranged in their correct stratigraphic position but do not reflect data from a single locality. Rather, the data base reflects samples from across the entire Salisbury Embayment. As a result, relative thickness on the vertical axis does not in any way reflect the relative or true thickness of a formation, but rather the number of samples analyzed for that particular formation. Samples analyzed in this study are marked with a dot on the plots; those from published sources do not have a dot. Modern = data from the Chesapeake estuary, Delaware estuary, and Susquehanna River. Glacial = data from pre-Woodfordian glacial deposits in New Jersey. 1 = includes data from Owens and others (1974); 2, 3 = includes data from Owens (1969) and Owens and others (1983); 4 = data from Owens and Glaser (1989); 5 = includes data from Jordan (1964), Owens and others (1983), Owens and Glaser (1989), and Leggett (1992); 6 = includes data from Owens and others (1983); 7 = includes data from Owens (1969) and Owens and others (1983). 8 = includes data from Jordan (1964), Owens and others (1983), and Leggett (1992); 9, 10 = data from McCartan (1989d); 11 = data from Owens and others (1989); 12 = data from Owens and others (1983), Benson and others (1985), McCartan (1989d); 13 = data from Owens and others (1989); 14 = data from Owens and others (1989); McCartan (1989d). 15 = zircon, tourmaline, rutile, ssak = staurolite, sillimanite, andalusite, kyanite; hegap = hornblende, epidote, garnet, actinolite, tremolite, pyroxene; 16 = Qms/Qu = ratio of monocrystalline non-strained quartz (straight extinction) to polycrystalline and strained quartz (undulatory extinction). Bethany and Manokin formation data include data from Leggett (1992). Grain sizes investigated in this study are 250 \( \mu \text{m} \) for the light fraction and 88–149 \( \mu \text{m} \) for the heavy fraction; grain sizes investigated in all Owens and McCartan references are 74–177 \( \mu \text{m} \) for both the light and heavy mineral fractions; grain sizes investigated by Jordan are 62–500 \( \mu \text{m} \) for both the light and heavy mineral fractions. Dashed horizontal line between Condon Farm and Turkey Point in plot denotes different geographic locations of the Turkey Point bed (TPb) samples.

lies below sea level (Conant, 1990). The Pensauken Formation is locally inset into the Perryville formation along the east side of the Elk Neck Peninsula (Figs. 4 and 5).

Three lithofacies, all contained in the exposures at Turkey Point, Maryland (Fig. 4), have been described for the Pensauken Formation (Fig. 8). Lithofacies 1, which is 5 m of red, medium- to coarse-grained, moderately well sorted, medium to thickly bedded, trough cross-stratified, feldspathic, pebbly sand with a coarse, channel-bottom gravel lag, unconformably overlies the Potomac Group. The troughs are developed on low-relief, northeast-striking, northwest-dipping surfaces of point bars and indicate a southwest paleoflow (Fig. 4). At least three 15-cm-thick, laminated, clayey silt beds with a single-pebble-thick lag base mark distinct unconformities through the otherwise sandy facies.
Lithofacies 1 is conformably overlain by lithofacies 2, which is composed of 6 m of white, pebbly, feldspathic sand with 20- to 40-cm-thick, tabular and trough cross-sets and 1-m-thick dunes on 1- to 5-m-thick point and longitudinal bars. Cross-stratification in lithofacies 2 commonly exhibits reactivation surfaces and, like lithofacies 1, indicates a predominantly southwest paleoflow direction. Lithofacies 3, a 6-m-thick, poorly sorted, thickly bedded, boulder-rich, feldspathic, pebbly sand interbedded with well-sorted, tabular cross-stratified, white quartz medium sand and thin (<10 cm) clayey stringers, unconformably overlies lithofacies 1 and 2 (Fig. 8). The coarse gravel and boulder clasts, some exceeding 2 m in diameter, are typically subangular to subrounded Valley and Ridge-derived sandstones, siltstones, black chert, red siltstones, and Piedmont-derived schists, gabbros, quartzites, and metasediments. In contrast to lithofacies 1 and 2, cross-stratification in lithofacies 3 exhibits south and southeast paleoflow orientations (Fig. 4).

Pensauken Formation deposition indicates major shifts in the source for upper Chesapeake Bay fluvial deposits (Owens and Denny, 1979a; Owens and Minard, 1979). Lithofacies 1 and 2 correspond to the main body of the Pensauken Formation occurring in New Jersey, the lower Delaware valley, and Delmarva Peninsula and represents braided alluvial-plain deposits of the paleo-Delaware-Hudson River (Owens and Minard, 1979; Jordan, 1964). In contrast, lithofacies 3 represents a later period of coarser-grained fluvial channel deposition, which in Cecil County, Maryland, shows a Susquehanna River influence.

Many of the tabular and trough cross-stratifications exhibited by lithofacies 1 and 2 throughout Cecil County commonly exhibit reactivation surfaces indicative of a shallow tidal estuarine depositional environment (Duke and others, 1992). Although reactivation surfaces are common in any fluvial or marine environment when bi-directional flows of unequal magnitude are present (Visser, 1980), bed thickness between the reactivation surfaces or bundles in the Pensauken Formation display a systematic waxing and waning, and they form couplets with an average of 63 bundles per couplet. As the lunar month is ~29.5 days long, semidiurnal tides would be expected to be represented by 59 bundles per couplet (Visser, 1980). Thus, the Pensauken Formation bundles may have been deposited as daily records of semidiurnal tides in tidally influenced, sandy estuarine channels with a tidal range large enough to cause flow reversal. Further support for marine incursions into Pensauken Formation channels comes from the laminated silts of lithofacies 1 which may represent marine, slack-water deposits on top of a transgressive lagemplaced during a rapid eustatic rise.

Pensauken Formation age has been strongly debated, as most palynological dating efforts have been unsuccessful. Age estimates range from Pleistocene, pre-Wisconsinan glacial outwash (Salisbury and Knaap, 1917; Jordan, 1964; J. Groot, unpub. data) to early Pleistocene or late Tertiary, warm-climate deposition (Campbell and Bascom, 1933; Leverett, 1934; Berry and Hawkins, 1935; MacClintock and Richards, 1936), or late Miocene, warm-climate deposition supported by downdp petrographic correlation to the Manokin formation (Owens and Denny, 1979a; Owens and Minard, 1979) and
a recently recovered palynoflora from New Jersey (J. P. Owens, unpub. data). Relative-age data such as the presence of gibbsite, halloysite, and iron oxides, all formed in situ, support a pre-Pleistocene age for the Pennsauken Formation because deposits of known Pleistocene age on the Coastal Plain do not contain these weathering products (Owens and others, 1983; Owens and Minard, 1979; Owens and Glaser, 1989; Soller and Owens, 1991). Middle Pleistocene glacial deposits 200 km to the north, however, have been shown to contain similar weathering products (Levine and Ciolkosz, 1983), and so it remains problematic to base a pre-Pleistocene age on analysis of weathering products alone. Additionally, many labile Pennsauken Formation clasts, particularly those in lithofacies 3, display a relatively thin weathering rind (~2 cm) and still ring when struck with a hammer, characteristics not shared with the Bryn Mawr or Perryville Formations.

**Petrography.** Pennsauken Formation petrography differs significantly from the Bryn Mawr and Perryville Formations. Clast counts of the pebble-size fraction (Table 1) show that lithofacies 1 is dominated by rounded quartz and quartzite, with minor sandstone and labile constituents. In contrast, lithofacies 3 exhibits more clasts of angular to rounded and weathered to relatively fresh Valley and Ridge-derived conglomerates, sandstones, silts, and black and white chert. Smaller percentages of labile lithologies such as schist, gneiss, gabbro, granite, and partially silicified limestone clasts with clearly discernible tetra coral and brachiopod fossils also occur (Table 1).

The petrography of the sand-size light minerals is decidedly feldspathic (Fig. 7), with a total feldspar content in excess of 50% in a few samples, and a lithic content ranging from 5% to 20%. Quartz grains are predominantly of the monocrystalline variety and exhibit decidedly more strained than non-strained character (Fig. 7). The non-opaque heavy mineral suite contains relatively high amounts of metastable and labile minerals such as staurite, sillimanite, kyanite, epidote, hornblende, and garnet, and correspondingly lower amounts of zircon, tourmaline, and rutile (Fig. 7). In particular, high amounts of epidote (averaging 15%) and hornblende (averaging 25%) occur. The opaque suite is dominated by ilmenite but also shows relatively high amounts of less stable opaques, such as magnetite (Fig. 7). Lower percentages of leucoxene and ilmenite-leucoxene occur with respect to the Perryville and Bryn Mawr Formations. The clay mineralogy of the Pennsauken Formation is characterized by vermiculite, kaolinite, and small amounts of gibbsite (Owens and Minard, 1979). Halloysite occurs and gibbsite content increases in the upper meter of the deposit (Owens and others, 1983).

**Turkey Point Beds**

Pebbly, sandy, silts interbedded with finely laminated, silty clays, and buried, truncated, paleosol unconformably overlie the Perryville and Pennsauken Formations (Fig. 5) across Cecil County and possibly most of northern Delmarva (Fig. 4). These deposits are here recognized as the Turkey Point beds. Three lithofacies compose the Turkey Point beds at their type section of Turkey Point, Maryland (Figs. 4 and 8). Lithofacies 1 is a red, tan, and white, cross-stratified, fine to medium sand, exhibiting plane parallel, tabular, and shallow trough cross-stratification showing a south and southeast direction of transport (Fig. 4), interbedded with flaser and wavy-beded fine sands and laminated clayey silts. Lithofacies 1 is unconformably overlain by lithofacies 2, a wavy-beded fine sand and silt and laminated, locally burrowed, brown, and silty clay about 2 m thick. An apparent conformable, gradational contact separates lithofacies 2 from the overlying lithofacies 3, a massive, brown, fine sand and silt unit approximately 1 to 2 m thick. The soil developed in the Turkey Point beds is cumulic and polygenetic (Table 2) and reflects at least three periods of relative surficial stability and soil formation.

The presence of interbedded, cross-stratified and flaser-beded sands, silts, and clays, infrequent burrows, and numerous unconformities strongly suggests a fluvial and tidal-estuarine origin (Howard and Frey, 1975; Greer, 1975). On the northern Delmarva Peninsula, the Turkey Point beds occur as fluvial, estuarine, colluvial, and eolian deposits resulting from the reworking of the Pennsauken Formation by fluvial, mass-movement, and eolian processes (Owens and Denny, 1979a). Along the present Cecil County shoreline, the Turkey Point beds occur as paleo-Chesapeake Bay fluvial and estuarine deposits. Thus, Turkey Point bed deposition may signify the shift from large, fluvially dominated systems to more estuarine conditions.

No datable material has been extracted from the Turkey Point beds; however, relative-age data are obtained from soil stratigraphy and morphology and magnetostratigraphy. Field morphologic properties (soil color, argillic horizon development, solum thickness; Table 2) of the soil present at Turkey Point resemble the morphologic development of Coastal Plain soils (Markewitz and others, 1990) and Valley and Ridge soils (Ciolkosz and others, 1990) determined to be late Pleistocene age. A truncated and buried paleosol at the contact between lithofacies 2 and lithofacies 1 (Fig. 8) indicates a previous period of subaerial exposure and weathering of unknown duration. It is impossible to estimate the relative amount of soil development and thus the time represented by the paleosol because most of it has been removed by erosion. The sedimentologic similarities between the units directly above and below the paleosol, however, suggest that the temporal gap is not on the order of millions of years, but is rather more likely to be tens of thousands to possibly hundreds of thousands of years.

Oriented samples of the silty beds within lithofacies 1 and 2 have been collected and analyzed for their paleomagnetic signal (Fig. 8). Strong normal polarities occur both above and below the buried paleosol. If the paleosol unconformity represents less than 800 ky, then the normal polarities probably indicate the Brunhes normal epoch, given the Pleistocene age of the surficial soil. Samples collected in lithofacies 1 at least 2 m below the paleosol, however, show reversed polarities. It thus remains possible that the Brunhes-Matuyama magnetic boundary resides within lithofacies 1, asserting an early Pleistocene age for the base of the Turkey Point beds.

**Petrography.** Lithofacies 1 and 2 contain predominantly quartz with 2% microcline, 1% plagioclase, and 10% lithic fragments (Fig. 7). The quartz grains are predominantly monocrystalline and of the strained rather than unstrained variety. The non-opaque heavy mineral suite is dominated by stable components, including pink zircon, tourmaline, and rutile (Fig. 7). The Turkey Point beds differ petrographically from the underlying Pennsauken Formation on the basis of higher percentages of metastable components, predominantly staurolite, which approaches 30%, and relatively lower percentages of labile minerals in the former. The Turkey Point beds also have proportionally less epidote (<5% versus >10%) and slightly more garnet (>5% versus <5%) than the Pennsauken Formation (Fig. 7). The opaque heavy-mineral fraction shows proportionally more leucoxene + ilmenite/leucoxene and a magnetite content of 23%, which is slightly
lower than the underlying Persauken Formation (Fig. 7).

Despite the limited number of samples, the increased proportion of metastable heavy minerals, primarily staurolite, in the Turkey Point beds does not simply reflect a decreased proportion of labile heavy minerals because the stable heavy-mineral proportion also decreases. Rather, the increased heavy metastable component is hypothesized to reflect a change in sediment source from the paleo-Delaware-Hudson system to the Susquehanna River because recent sediments from the Susquehanna River are also rich in staurolite.

**DISCUSSION**

**Correlation of Upper Coastal Plain Fluviatile Deposits**

The age and origin of upper Chesapeake fluviatile deposits can be determined through petrographically and stratigraphically based lithostratigraphic correlations along the Fall Zone. This work focused on the Lower Miocene–Pliocene section along the Fall Zone and downdip into the Salisbury Embayment marine sequence (Fig. 9). The quartz-pegmatite petrography, coarse, pebbly-sand texture, stratigraphic location, deposit continuity along the Fall Zone, and to a lesser degree, provenance, suggest that the Bryn Mawr Formation is a Susquehanna River-dominated deposit lithostratigraphically correlatable to the upland gravels in New Jersey (Beacon Hill Formation), in southern Maryland (Brendwyine Formation), and Virginia (Bon Air Formation) (Fig. 9). The late Miocene formed by Pliocene age (McCurtain and others, 1990; Groot and others, 1990) of the Brandywine gravels and the fact that the upland gravels in New Jersey, southern Maryland, and Virginia locally unconformably overlie middle Miocene marine units (Owens and Minard, 1979; McCurtain and others, 1990; Howard and others, 1993; Figs. 2 and 3) strongly suggest a Miocene age for the Bryn Mawr Formation.

Bryn Mawr Formation petrography, characterized by a quartz-dominated light-mineral fraction, and zircon-tourmaline-rutil-dominated heavy-mineral fraction most closely matches the petrography of late Oligocene through middle Miocene formations of the Sussex Group and Kirkwood and Cohansay Formations of New Jersey (Figs. 7 and 9). The somewhat more mature light- and heavy-mineral suite exhibited by the Bryn Mawr Formation may reflect the effects of in situ chemical weathering and removal of labile components and/or the effects of climate and transport distance (Johnsson and others, 1991). Petrology of the Calvert Formation recently exposed in central Delaware (Cheswold Sand) has a higher proportion of quartz and stable heavy minerals than does the more down-dip Calvert Formation (Fig. 7), perhaps reflecting less mixing of multiple sources in the more proximal regions. The Cheswold Sand also lacks marine fauna, exhibiting lithofacies consistent with well-sorted shoreline deposits similar to those exhibited by member 2 of the Bryn Mawr Formation.

Similar amounts of leucocene and ilmenite-leucoxene and similar proportions of non-stained to strained and polycrystalline quartz (Fig. 7) suggest that the late Miocene Manokin and Bethany formations may also, in part, correlate to the Bryn Mawr Formation. The Bethany and Manokin formations, however, are petrographically more similar to the Bridgeton and/or Persauken Formations of New Jersey (see below) which supports the interpretations that the paleo-Delaware-Hudson River system occupied southern New Jersey and Delmarva in post-Cohansey (Choptank) time (Owens and Minard, 1979) and that the Bethany and Manokin formations are lower delta-plain to prodelta deposits of a south-flowing fluvial system (Owens and Denny, 1979a; Andres, 1986). Nevertheless, Bethany and Manokin formations petrography may reflect mixing of sediment contributed from the paleo-Delaware-Hudson and Susquehanna drainages. The composition, stratigraphic position, and to a lesser extent, elevation of the Per- ryville formation resemble (1) the Bridgeton Formation and related unnamed, fan and terraces deposits along the Fall Zone in New Jersey thought to be late Miocene–Pliocene age (Owens and Minard, 1979; Newell and others, 1989), (2) Pliocene terraces underlain by Brandywine Formation in southern Maryland (McCurtain and others, 1990), (3) the Pliocene Yorktown Formation upland gravels (McCurtain, 1989a), (4) the late Pliocene Park Hall Formation (McCurtain, 1989a), and (5) late Pliocene Bacon's Castle Formation (Ramsey, 1988). Because the Perryville for-

formation reflects local deposition by many different sources across the northern Chesapeake region, it does not correlate downdip to a specific marine unit. In the absence of dateable material, a late Tertiary, probably Pliocene age is favored for the Perryville formation, although a Miocene to early Pleistocene age is possible (Fig. 9). Pensauken Formation petrography indicates a major shift in the composition of upper Chesapeake fluvial deposits from quartz-dominated to decidedly feldspathic and/or lithic-rich (Fig. 9) beginning with Bridgeton Formation deposition in New Jersey and sweeping southwestward to the Delmarva Peninsula with the Manokin, Bethany, Pensauken, and Beaverdam Formation deposition. Petrographic similarities, used to sug-
Figure 10. Schematic cross section parallel to B–B’ of Figure 2 illustrating the relative vertical separation between proximal fluvial deposits on the upper Coastal Plain–Fall Zone region and the distal marine equivalents in the basin. Long-term basin subsidence provides accommodation space to preserve the sediments, and it results in vertical separation of deposits, whereas long-term relative base-level stability for the upper Coastal Plain and Fall Zone results in both deposition and erosion, resulting in a vertically “condensed” section. Note that if not completely removed by erosion, fluvial stratigraphy landward of the Fall Zone would also experience significant vertical separation.

gest a lithostratigraphic correlation between the Pensauken (Columbia) and Beaverdam Formations (Jordan, 1964) or the Pensauken and Bethany and Manokin formations (Owens and Denny, 1979a; Owens and Minard, 1979), in part agree with the results of this study. In this study, Bethany and Manokin Formation samples contained less feldspar and lithic fragments in the light fraction, and more metamorphic minerals (particularly staurolite and ilmenite-leucoxene) and leucoxene in the heavy-mineral fraction than did the Pensauken and Beaverdam Formations (Fig. 7). Distribution of the most common labile heavy minerals (hornblende, garnet, and epidote) demonstrates that the Bethany and Manokin formations contain significantly less epidote and hornblende and more garnet than do the Pensauken and Beaverdam Formations. In contrast, the petrography of the Bridgeton Formation very closely resembles Bethany and Manokin formation petrography for both the light- and heavy-mineral suite with virtually the same percentages of feldspar, heavy metamorphic component, epidote, and ilmenite/leucoxene and leucoxene. Thus, the Bridgeton Formation, which previously was not recognized to have a distal fluvial-deltaic or marine equivalent, represents a logical lithostratigraphic equivalent to the Bethany and Manokin formations, but the stratigraphically younger Pensauken Formation correlates to the upper Bethany through Beaverdam Formation (Hansen, 1981). These results suggest a late Miocene age for the Bridgeton Formation (Owens and Minard, 1979) and a Pliocene age for the Pensauken Formation.

**Implications for Flexural Isostatic, Tectonic, and Eustatic Processes**

Petrographically based lithostratigraphic correlations presented here have broad implications for regional flexural isostatic deformation and eustatic fluctuations on the middle Atlantic passive margin. Specifically, the clearly defined unconformities between members of the Bryn Mawr Formation could simply indicate the normal juxtaposition of depositional facies within a single broad plain. Alternatively, they may indicate major hiatuses correlative downlap to sequence boundaries in the Salisbury Embayment. Although Bryn Mawr Formation deposition has been traditionally viewed as a single aggradation event (Bascom and others, 1902; Owens, 1969; Conant, 1990), petrographic correlation to the late Oligocene–early Miocene Chesapeake Group suggests a protracted period (~10 m.y.) of Bryn Mawr Formation deposition.

Several third-order transgressions, separated by major hiatuses, occurred in the Salisbury Embayment from the late Oligocene through late Miocene (Fig. 9). Conceptually, the three members within the Bryn Mawr Formation represent at least three discrete fluvial aggradational events, most likely during major eustatic transgressions. In one possible scenario, phase 1 deposition represented by member 1 (Fig. 6) was initiated in the late Oligocene or early Miocene during Old Church or lower Calvert Formation marine deposition, respectively. Phase 2 deposition followed in the late early Miocene through middle Miocene in response to fluvial aggradation associated with the Calvert and Choptank Formations transgressions. The eustatic rise associated with Phase 2 deposition was the most extensive, resulting in the more distal braid-plain lithofacies 2, 3, and 4 of member 2 (Fig. 6). Finally, the third phase of Bryn Mawr Formation deposition, represented by the more proximal braid-plain lithofacies 5 and 6 of member 3, may have occurred from the late Miocene to earliest Pliocene in response to the less extensive St. Marys and Eastover Formations (Manokin and Bethany formations) transgressions.

Further evidence for the polygenetic nature of the Bryn Mawr Formation comes from fluvial stratigraphy of the lower Susquehanna River and regional flexural isostatic deformation of the middle Atlantic passive margin. Three distinct, texturally and compositionally mature, quartz, fluvial terraces occur along the lower Susquehanna River (Pazzaglia and Gardner, 1992 and in press [b]). The longitudinal profiles of these terraces converge downstream into the Bryn Mawr Formation on the Fall Zone and disperse through the Piedmont (Fig. 10). The marine stratigraphy of the Salisbury Embayment mirrors the terrace stratigraphy, thinning to a feather edge at the Fall Zone and thickening basinward (Fig. 10). Irrespective of the lithostratigraphic correlation between the terraces, Bryn Mawr Formation, and marine stratigraphy, geodynamic modeling, constrained by the marine stratigraphy alone, clearly shows how the upper Coastal Plain and Fall Zone behaves as a flexural hinge (Pazzaglia and Gardner, in press [a]; Fig. 10). The hinge undergoes relatively little isostatic uplift or subsidence resulting in a condensed stratigraphic section on the upper Coastal Plain and Fall Zone as it accommodates the long-term subsidence of the Salisbury Embayment and the long-term uplift of the Pied-
Figure 11. Schematic diagrams summarizing middle Atlantic Coastal Plain deposition from the late Oligocene to the middle Pleistocene. See text for detailed descriptions.
mont (Pazzaglia and Gardner, in press a). Relative isostatic stability of the upper Coastal Plain and Fall Zone leaves eustasy as the dominant component for relative base-level change. Thus, in the absence of major tectonic deformation, the elevation of fluvial deposits across the flexural hinge represents a reasonable estimation for paleo-sea level. Given the elevation of Bryn Mawr Formation member 2 on the upper Fall Zone as about 120 m amsl, and modeling results which estimate the isostatic uplift in this region between 35 and 60 m (Pazzaglia and Gardner, in press a), the middle Miocene eustatic highstand is estimated between 60 and 85 m.

Second-order tectonic processes resulting from the flexural bending of the margin and intraplate horizontal stress (Zoback and Zoback, 1989) may be reflected in the elevation of the Turkey Point beds, Pensauken Formation, and orientation of Chesapeake Bay. The Turkey Point beds at Turkey Point, Grove Point, and Beterton (Fig. 4) lie 6–8 m higher than at the mouth of the Susquehanna River. Near-vertical cliffs of the northeast shore of Chesapeake Bay, exposing the Turkey Point beds and Pensauken Formation up to 30 m amsl, contrast with the poor exposure of these deposits elsewhere, such as eastern Delmarva, where they do not exceed 18 m amsl. These elevation disparities suggest an 8 m of post-early Pleistocene offset along a northeast-southwest-trending fault beneath upper Chesapeake Bay (Fig. 4). This tectonic deformation is consistent with estimated fault orientation and offset elsewhere along the Fall Zone (Newell and Rader, 1982; Mixon and Newell, 1982; Mixon and Powars, 1984; Newell, 1985).

CONCLUSIONS

Petrography-based lithostratigraphic correlation of fluvial deposits at the head of Chesapeake Bay both along the Fall Zone and downdip into the Salisbury Embayment marine stratigraphy provides a framework through which to synthesize a regional chronostratigraphic correlation (Fig. 9), speculate on tectonic, isostatic, fluvial, and climatic processes, and reconstruct the depositional history for the Salisbury Embayment (Fig. 11). The late Oligocene middle Atlantic Coastal Plain during the Old Church transgression is envisioned to have had a broad, arcuate shape without large bays, estuaries, or salients where large quartzose braid plains prograded from the paleo-Hudson, Susquehanna, and Potomac Rivers (Fig. 11a). Relative eustatic fall in the early Miocene resulted in a shift of the shoreline basinward and subsequent abandonment of major fluvial aggradation on the upper Coastal Plain and Fall Zone. In the late early and middle Miocene, upper Coastal Plain and Fall Zone quartzose braided-plain fluvial aggradation reasumed as the proximal equivalents to the Calvert (Kirkwood) and Choptank (Cohansey) Formations (Fig. 11b). Uplift and down-to-the-southwest tilting of the entire basin (Newell and Rader, 1982), perhaps coupled with a eustatic drawdown by the late Miocene, resulted in subaerial exposure of the New Jersey Coastal Plain and the subsequent, systematic swing of river systems to the southwest (Hack, 1955; Owens and Minard, 1979). Late Miocene eustatic rise resulted in Eastover Formation marine deposition in the southwest and fluvial aggradation in the northeast as the final phase of quartzose Bryn Mawr deposition at the mouth of the Susquehanna River, quartzo-feldspathic Bridgeton Formation fluvial deposition across southern New Jersey, quartzo-feldspathic Bethany and Manokin fluvial-deltaic deposition on Delmarva, and quartzose Brandywine and Bon Air fluvial deposition across southern Maryland and Virginia (Fig. 11c).

Pliocene Chesapeake Group marine deposition (Yorktown and Chowan River Formations) followed subaerial exposure and erosion of the upper Coastal Plain during latest Miocene and early Pliocene eustatic drawdown. The eustatic rises associated with the Pliocene transgressions resulted in fluvial aggradation in the upper Coastal Plain and the deposition of the Perryville and Pensauken Formations at the head of Chesapeake Bay, unammed fan, terrace, and pediment deposits in New Jersey (Newell and others, 1989), and the Yorktown Upland Gravels, Park Hall, and Bacons Castle Formations (Ramsey, 1992) in southern Maryland and Virginia (Figs. 11d and 11e). Given the relative isostatic stability of the head of Chesapeake Bay, estimates of maximum Pliocene eustatic rise of 35 ± 18 m (Dowsett and Cronin, 1990) and 25 to 40 m amsl (Krantz, 1991) are consistent with the 30 to 45 m amsl Perryville formation elevation range. In central Delmarva, the Pliocene upper Bethany and lower Beaverdam Formations (Groot and others, 1990) were deposited as fluvial-deltaic facies between the updip, fluvial Pensauken Formation and downdip, marine Yorktown Formation (Ramsey, 1992; Fig. 11d).

Several independent lines of evidence strongly suggest periodic, but cumulative terrestrial cooling throughout the Pliocene (Webb, 1988; Groot and others, 1990; Groot, 1991), culminating with the first Northern Hemisphere glaciation in the late Pliocene at 2.4 Ma (Shackleton and Hall, 1984; Zimmerman and others, 1984; Shackleton and others, 1984; Shackleton, 1987). The Pensauken Formation, in part, represents Pliocene fluvial aggradation in response to the climatic degradation. Lithofacies 3 of the Pensauken Formation deposition may reflect large, coarse sediment contributions associated with the first Northern Hemisphere glaciation (Fig. 11e).

Late Pliocene glaciation was followed by warm terrestrial temperatures similar to those of the middle Pliocene (Webb, 1988; Travers, 1988; Groot and others, 1990) and eustatic highstands on the order of 15 to 25 m amsl (Cronin and others, 1981, 1984; Dowsett and Cronin, 1990; Krantz, 1991). Predominantly fluvial-estuarine deposition of units like the Turkey Point beds and Windsor Formation occurred at this time during the high-frequency (≈40 ky–duration), but low-amplitude, early Pleistocene eustatic fluctuations (Shackleton and Hall, 1984; Zimmerman and others, 1984; Wright, 1989; Figs. 11e and 11f). Sea-level lows during the last glacial maxima, about 120 m below present (Fairbanks, 1989) are fairly representative of middle and late Pleistocene drawdowns and are ultimately responsible for consuming the central Chesapeake Miocene-Pliocene fluvial upland (Vogt, 1991) in the creation of the modern Chesapeake and Delaware Bays (Colman and others, 1990).

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CENOZOIC ATLANTIC COASTAL PLAIN DEPOSITS

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